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Subglacial lakes and jökulhlaups in Iceland

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Abstract

Active volcanoes and hydrothermal systems underlie ice caps in Iceland. Glacier–volcano interactions produce meltwater that either drains toward the glacier margin or accumulates in subglacial lakes. Accumulated meltwater drains periodically in jökulhlaups from the subglacial lakes and occasionally during volcanic eruptions. The release of meltwater from glacial lakes can take place in two different mechanisms. Drainage can begin at pressures lower than the ice overburden in conduits that expand slowly due to melting of the ice walls by frictional and sensible heat in the water. Alternatively, the lake level rises until the ice dam is lifted and water pressure in excess of the ice overburden opens the waterways; the glacier is lifted along the flowpath to make space for the water. In this case, discharge rises faster than can be accommodated by melting of the conduits. Normally jökulhlaups do not lead to glacier surges but eruptions in ice-capped stratovolcanoes have caused rapid and extensive glacier sliding. Jökulhlaups from subglacial lakes may transport on the order of 10^7 tons of sediment per event but during violent volcanic eruptions, the sediment load has been 10^8 tons.

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1. Introduction

Volcanoes and glaciers, unlike some agents of earth-surface evolution, are spatially contained. Their interaction can be violent, leading to some of Nature's most spectacular clashes, with few parallels in magnitude or in the display of elemental power. This uneasy relationship between ice and volcanism involves extremely rapid heat transfer and explosive fragmentation of lava into tiny glass particles. Catastrophic floods burst from the glaciers, ranging from floods of water to floods of hyperconcentrated fluid–sediment mixtures and even more destructive gravity-

driven mass flow of volcanic material (viscous non-Newtonian lahars). Meanwhile within the glaciers, eruptions pile up hyaloclastite ridges and mountains. Eruptions may break through the ice cover and spread tephra over the surroundings or form lava caps on top of hyaloclastite mountains (tuyas). Injection of magma to shallow depths, however, may lead to quieter expressions of glaciovolcanism; hydrothermal systems are created that bring heat up from a magma body, continuously melting ice into water that may be stored at the glacier bed until it breaks out in jökulhlaups. Both the visible and invisible expressions of glaciovolcanic activity lead to catastrophic floods of water. While the timing of volcanic eruptions is unpredictable, jökulhlaups from subglacial lakes occur episodically as the culmination of a more subdued series of events.

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Iceland is a unique and valuable study site for glaciovolcanic interactions. At present, 10% (11,200 km²) of Iceland is ice-covered and 60% of the ice overlies the active volcanic zone (Fig. 1). During the 20th century, 15 subglacial volcanic eruptions (10 major and five minor events) took place, about one-third of all eruptions in Iceland during that century. During the Pleistocene Era when Iceland was entirely covered with ice, the Palagonite Formation, characterised by hyaloclastite ridges, mounds and tuyas, was formed across the entire country (Kjartansson, 1943; Mathews, 1947; Bemmelen and van Rutten, 1955; Jakobsson, 1979; Jóhannesson et al., 1990). Studies of the processes presently at work in glaciated volcanic areas provide information about the agents that shaped the landscape during the Pleistocene Era. This applies to the heat exchange between magma and the glacier, the dynamical response of the glacier to subglacial eruptions, the structure and growth sequence of hyaloclastite ridges and tuyas formed by subglacial eruptions, and jökulhlaups due to volcanic eruptions. Jökulhlaups, both those draining meltwater stored in subglacial lakes and meltwater produced during a volcanic eruption, have significant landscaping potential: they

erode large canyons and transport and deposit enormous quantities of sediment and icebergs over vast outwash plains and sandur deltas. Pleistocene glacial river canyons may have been formed in such catastrophic floods from subglacial lakes. Jökulhlaups have threatened human populations, farms and hydroelectric power plants on glacier-fed rivers. They have damaged cultivated and vegetation areas, disrupted roads on the outwash plains and have even generated flood waves in coastal waters. Knowledge of the sources and behaviour of jökulhlaups is essential for advanced warnings and civil defence in Iceland. Monitored seismic activity announces volcanic eruptions, and maps of bedrock and glacier surface topography can be used to delineate jökulhlaup hazard zones (Björnsson and Einarsson, 1991). Studies of glacier–volcano interactions on Earth may further understanding of the impact of volcanism on other planets such as Mars.

This paper describes glacier–volcano interactions and jökulhlaups in Iceland with an emphasis on the formation and geometry of subglacial lakes in hydrothermal areas and jökulhlaup drainage characteristics. In the present paper, we only consider the drainage of water in two ice caps, Vatnajökull and Mýrdalsjökull.

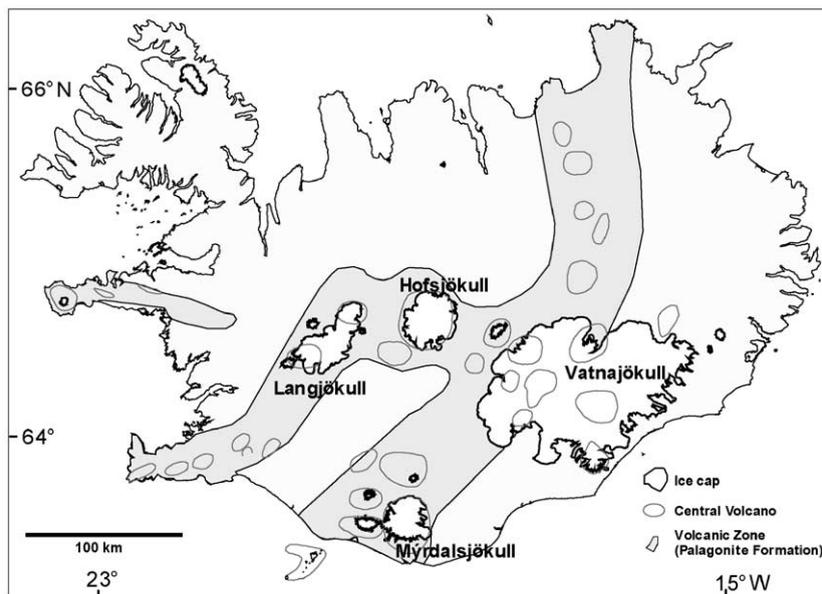


Fig. 1. Location map of Iceland showing ice caps, the volcanic zone (The Palagonite Formation) and the central volcanoes.

2. Ice caps, volcanic centres and drainage basins

Subglacial topography of all the major ice caps in Iceland has been mapped by radio echo sounding, revealing bedrock elevations from -300 to 1800 – 2000 m relative to sea level. The volcanic zone underlies the four largest ice caps and the most active volcanoes are located under the ice cap Mýrdalsjökull (600 km²) and the western part of Vatnajökull (8100 km²), which is positioned over the centre of the Icelandic mantle plume (Wolfe et al., 1997). No eruptions have taken place in Hofsjökull (925 km²) or Langjökull (925 km²) ice caps during the last millennium.

Five volcanic systems have been identified under the ice cap Vatnajökull (Fig. 1), each containing central volcanoes and fissure swarms that stretch for tens of kilometres (Björnsson, 1988; Jóhannesson et al., 1990; Björnsson and Einarsson, 1991). The most active are the Grímsvötn and Bárðarbunga volcanic systems, both of which have relief of up to 1000 m and contain calderas 600 – 700 m deep and 15 – 20 km in diameter. In Grímsvötn, about 30 eruptions have been reported over the last 400 years (Thorarinsson, 1974; Björnsson and Einarsson, 1991). More than 80 volcanic eruptions occurred during the last 800 years in Vatnajökull (Larsen et al., 1998). Since the settlement of Iceland around 870 A.D., 20 volcanic eruptions have been traced to the Katla volcanic system in Mýrdalsjökull ice cap (Larsen, 2000). This volcanic system contains a 600 – 750 m deep caldera that is 10 – 15 km in diameter (Björnsson et al., 2000).

At several of the central volcanoes under Vatnajökull and Mýrdalsjökull hydrothermal activity results from the interaction of water with magmatic intrusions at shallow depths in the crust (Thorarinsson, 1974; Björnsson, 1988; Björnsson et al., 2000, 2001a). Ice is continuously melted at the glacier bed creating permanent depressions in the glacier surface that reveal this hydrothermal activity. The meltwater, however, may be trapped in a lake at the bed due to relatively low basal fluid potential under the depression. High ice overburden pressure at the rim around the depression seals the lake. The best-known examples of such subglacial lakes are Grímsvötn in the interior of Vatnajökull and the Skaftá cauldrons (Fig. 1). Smaller lakes of this kind are located at Pálsfjall and Kverkfjöll. The current hydrothermal activity in

Mýrdalsjökull is concentrated just inside the caldera rims, where faults allow rapid vertical transport of hydrothermal fluid.

2.1. Drainage basins

On the basis of the glacier surface and bedrock maps, the ice and water drainage basins on all the major ice caps in Iceland have been delineated and the location and geometry of subglacial lakes identified (Björnsson, 1982, 1986, 1988; Björnsson et al., 2000). The glacierized portion of catchment basins for the major glacial rivers has been drawn as a continuation of the watershed outside the glacier. In principle, the watershed at the glacier base is situated along ridge crests in the fluid potential

$$\phi_b = \rho_w g z_b + p_w, \quad (1)$$

which is the sum of terms expressing the gravitational potential and the water pressure, p_w . The symbol $\rho_w = 1000$ kg m⁻³ represents the density of water, $g = 9.81$ m s⁻² is the acceleration due to gravity and z_b is the elevation of the glacier substrate relative to sea level. Water in an isotropic basal layer would flow perpendicularly to lines of equal potential. The basal water pressure is assumed to be static,

$$p_w = k p_i, \quad (2)$$

where $p_i = \rho_i g H$ is the ice overburden pressure, k is a constant, $\rho_i = 916$ kg m⁻³ is the density of ice, $H = z_s - z_b$ is the thickness of the glacier and z_s is the elevation of the ice surface relative to the sea level (Shreve, 1972). In our experience, k is close to one (Björnsson, 1982, 1988) and thus the gradient driving the water is

$$\nabla \phi_b = (\rho_w - \rho_i) g \nabla z_b + g \rho_i \nabla z_s. \quad (3)$$

This expression predicts that the glacier surface slope is about 10 times more effective than the bed slope in controlling basal water flow. This static approximation of water pressure ($p_w \approx p_i$) is plausible at low values of discharge and is therefore useful for delineating water divides.

Vatnajökull and Mýrdalsjökull have been divided into 15 and three major drainage basins, respectively (Fig. 2). Six basins in Vatnajökull, where water accu-

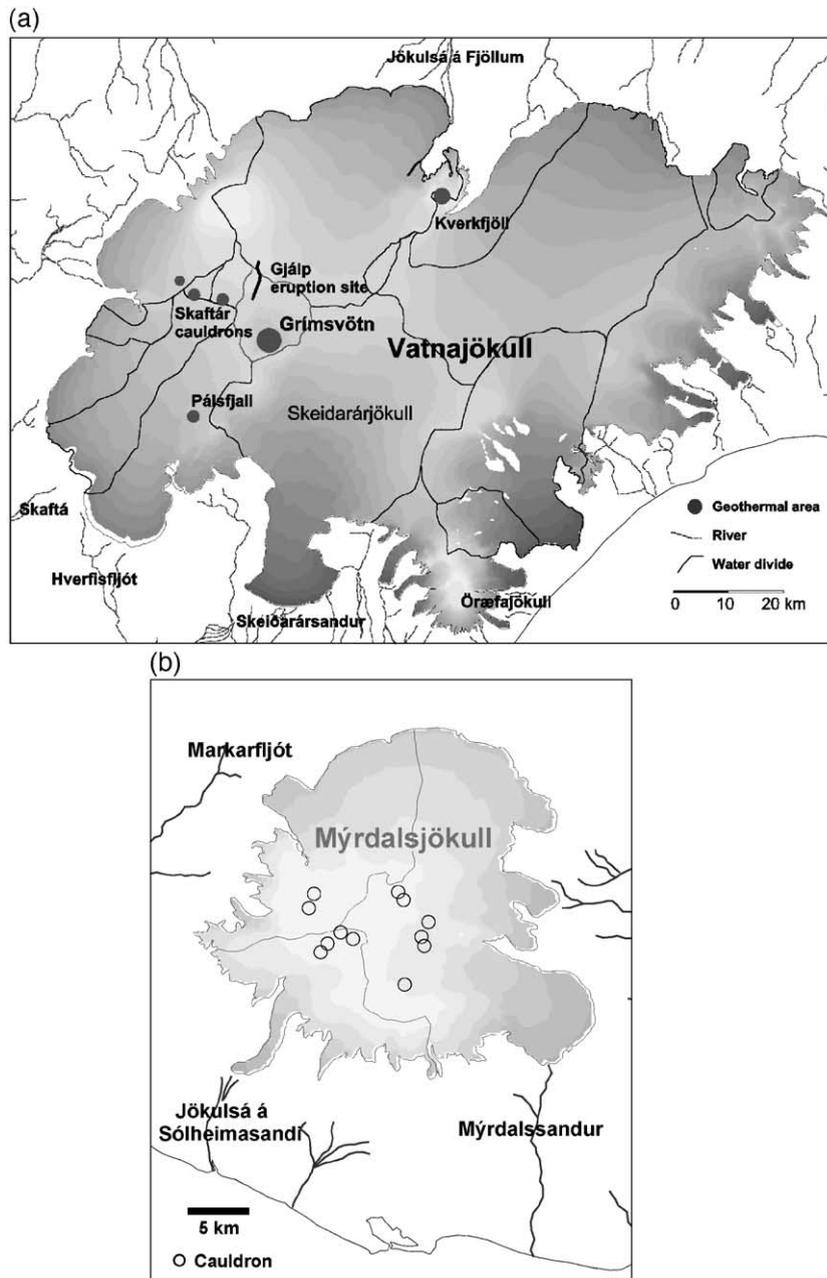


Fig. 2. Drainage basins, and location of subglacial geothermal systems. (a) Vatnajökull: The internal drainage basins of Grímsvötn and Skaftá Cauldrons drain in jökulhlaups. (b) Mýrdalsjökull: Cauldrons produced by subglacial geothermal activity.

mulates into subglacial lakes, are not connected to the glacier margin and water may accumulate under 12 small ice cauldrons in Mýrdalsjökull. During volcanic eruptions, depressions are created in the glacier surface

above the eruption site and the location of the watersheds may change. Nevertheless, if our model of a static fluid potential applies at the beginning of an eruption, we consider it likely that meltwater from the

eruption site will continue to drain through existing conduits.

2.2. Subglacial lakes

Subglacial lakes can be situated where there is no gradient in the fluid potential that drives water along the glacier bed. The condition $\nabla\phi_b = 0$ has been used to define the location and geometry of subglacial lakes. Hence

$$\nabla z_b = -(\rho_i/(\rho_w - \rho_i))\nabla z_s, \quad (4)$$

describes a relationship between the slope of the ice/water boundary of the lake and the upper glacier surface. The shape of the lake results from an equilibrium of vertical forces as the overlying glacier floats in static equilibrium. The roof of the subglacial lake slopes approximately 10 times more steeply than the glacier surface, and in the opposite direction (Fig. 3).

In Icelandic ice caps, several lakes are known to exist beneath surface depressions created above hydrothermal systems. The lakes rise as a dome above the bed, even capping mountains (Björnsson, 1975). Theory combined with topographic maps suggests that the slopes of glacier-bed depressions beneath ice caps in Iceland are not sufficient to accumulate water without accompanying depressions in the glacier surface. However, there could be lakes, not treated here, with equal in- and outflow that do not

meet the condition of zero gradient in the fluid potential.

2.3. Anatomy of a jökulhlaup

Subglacial lakes beneath ice-surface depressions drain periodically in outburst floods. A subglacial lake gradually expands as water flows toward the depression, the basal water pressure increases and the overlying glacier is lifted. Before the surface depression is completely flattened, the hydraulic seal is broken and water begins draining out of the lake at the base under the ice dam. The water escapes through narrow passages at the ice-bed interface but the pressure head maintained by a voluminous lake drops slowly. Although the pressure of the ice squeezes the tunnel draining water from the lake, water flow is primarily controlled by tunnel enlargement. In most cases, tunnel enlargement can be explained as melting of the ice walls by frictional heat generated by the flowing water and thermal energy stored in the lake (Nye, 1976; Clarke, 1982; Björnsson, 1992). The lake may become sealed again before it is empty and accumulation of water begins until a new jökulhlaup occurs.

Nye (1976) presented a general theory of flow in water-filled tunnels based on the principles of mass continuity, energy conservation, heat transfer and water discharge for a given fluid potential gradient. He derived an analytical solution predicting discharge

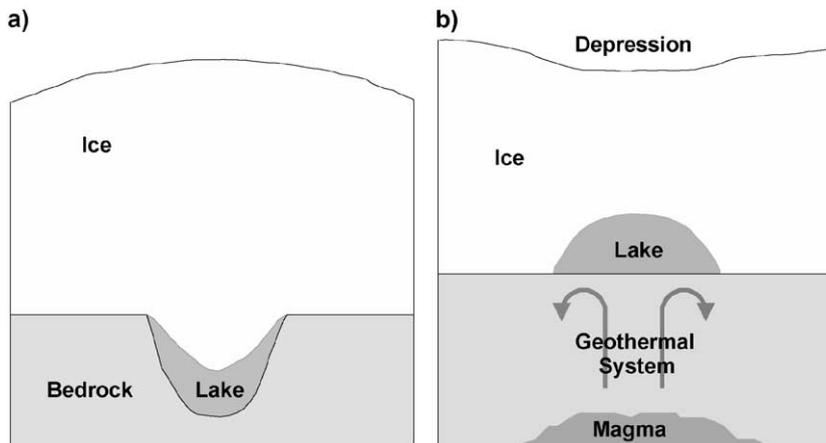


Fig. 3. Schematic drawing of two main types of subglacial lakes; (a) a stable lake, (b) an unstable lake that drains in jökulhlaups.

Q to rise asymptotically with time as $Q(t) = k(1/t)^4$ if the overburden closure is neglected and the expansion of the ice tunnel is solely due to the instantaneous transfer of frictional heat from the flowing water to the ice walls (loss of potential energy). Under the assumption of instantaneous heat transfer, water would emerge at the river outlet at the melting point, as is generally observed in jökulhlaups in Iceland, whereas the heat transfer equation would predict higher temperatures.

Occasionally, the discharge hydrograph for jökulhlaups increases faster than can be explained by the expansion of conduits by melting (Björnsson, 1997). The water pressure exceeds the ice overburden and the glacier is lifted to make space for the water (Björnsson, 1992, 1997). This event cannot be explained by the classical theory of jökulhlaups. Jóhannesson (2002) has described these floods as characterized by a propagation of a subglacial pressure wave. A preliminary model study suggests that the idea of a sheet flood preceding conduit drainage is plausible (Björnsson et al., 2001b).

3. Subglacial lakes, volcanic eruptions and jökulhlaups in Vatnajökull

Six subglacial lakes have been discovered in Vatnajökull (Fig. 2). For two of these, we will describe their geometry and growth, and the characteristics of their jökulhlaups.

3.1. Grímsvötn: geometry and drainage of ice and water

Grímsvötn, the largest subglacial lake in Iceland, is located in the western part of Vatnajökull (Fig. 4a, Table 1). The glacier covers a hydrothermal area and a 10-km wide and 300-m deep depression has been created in the ice surface (Björnsson, 1974, 1988). The extent of the subglacial lake is identified by the flat ice shelf floating on the lake and the abrupt change in surface slope at the margins. Subglacial topography is known from radio echo-soundings and seismic profiling (Björnsson, 1988; Guðmundsson, 1989, 1992). The lake is situated within the caldera floor of the Grímsvötn volcano and to the south and west the caldera walls (Mt. Grímsfjall) protrude

through the glacier surface and confine the lake, but the lake can expand to the north and northeast as the water level rises. The lowest breach of the caldera rims is 1150 m but due to the glacier surface depression the lake level at the slopes of Mt. Grímsfjall can rise more than 300 m higher. The ice-shelf thickness has been measured regularly by radio echo-soundings and computed from measured surface elevations, assuming the shelf is floating in hydrostatic equilibrium. The ice shelf gradually thickened from 150 m in the 1950s to 230–260 m in the 1980s due to reduced melting by the hydrothermal system (Björnsson and Guðmundsson, 1993). Hence, the extent and volume of the subglacial lake has gradually been reduced. The total volume of water drained out of Grímsvötn in jökulhlaups has been derived using known variations in lake level during jökulhlaups, the thickness of the floating ice cover on the lake and the bedrock topography in the Grímsvötn area (Björnsson, 1988, 1992).

In recent years the lake level has risen 10–15 m year⁻¹ and a jökulhlaup occurs when it rises 80–110 m and reaches a particular threshold (Fig. 5). The onset of lake drainage is marked by ice-quakes and subsidence of the lake level, and the arrival time of lake water to the glacier margin is identified by a sulphurous odour in the glacial river. Jökulhlaups from Grímsvötn flow beneath Skeiðarárjökull a distance of 50 km to the terminus at Skeiðarársandur. The flowpath reaches a depth of 200 m below sea level before it emerges on the outwash plain. In the most violent jökulhlaups from Grímsvötn, the entire outwash-plain, Skeiðarársandur, has been flooded. Jökulhlaups from Grímsvötn have occurred at 1– to 10-year intervals, with peak discharges of 600 to 4–5 × 10⁴ m³ s⁻¹ at the glacier margin, a duration of 2 days to 4 weeks and a total volume of 0.5–4.0 km³ (Table 2; Björnsson and Guðmundsson, 1993; Guðmundsson et al., 1995; Björnsson, 1997; Snorrason et al., 1997). In general, the frequency of jökulhlaups and the volume of water released depends upon the thickness of the ice barrier (Thorarinsson, 1974).

The typical threshold lake-level for triggering a jökulhlaup is 60–70 m lower than required for simple flotation of the ice dam (Björnsson, 1988, 1992). This suggests that some process other than lifting finally breaks the seal and permits the water to enter conduits beneath the ice dam, resulting in a jökulhlaup. Jökulh-



Fig. 4. (a) An oblique aerial photograph of Grímsvötn after the 10-km-wide and 270-m-thick ice cover had subsided 175 m during the 1996 jökulhlaup. View from the NNW toward the 400-m-high Mt. Grímsfjall (the southern caldera rim). (b) The 150-m-deep and 3-km-wide eastern Skaftá cauldron, just after a jökulhlaup in January 1982. The crevasse trending from the upper left of the cauldron suggests lifting of the ice dam. Further cauldrons are seen in the background. (c) The river Skeiðará flowing down the Skeiðarársandur outwash plain to the coast. Dykes control the river on the left hand side. The bridge across the river is 900 m long. (d) The 300-m-deep Jökulsárgljúfur canyon, north of Vatnajökull, eroded and carved during prehistoric jökulhlaups.

Table 1
Data on two subglacial lakes in Vatnajökull (Björnsson, 1988; Björnsson and Guðmundsson, 1993)

Source	Catchment basin (km ²)	Lake area (km ²)	Ice cover thickness (m)	Geothermal output (MW)	Basal melting rate (m ³ s ⁻¹)	Meltwater accumulation rate (km ⁻³ year ⁻¹)	Lake level rising rate (m year ⁻¹)	Lake level fluctuations between jökulhlaups (m)
Grímsvötn	160	5–40	150–270	2×10^3 – 4×10^3	6–12	0.2–0.5	10–15	80–175
E-Skaftá cauldron	40	<5	400	800	2.5	0.1	25	100

The basal melting rate may be underestimated as it is calculated from the drained meltwater during jökulhlaups and does not include meltwater drained into groundwater.

laups from Grímsvötn occur at any time of the year so sudden changes in subglacial drainage due to surface melting do not, in general, trigger jökulhlaups. A few jökulhlaups have occurred from Grímsvötn at lake levels far below the usual threshold. These premature jökulhlaups may have been triggered by the opening of waterways from the lake along the northeastern slopes of Grímsfjall (higher than the lowest crest at the caldera rim) facilitated by increased localized melting due to hydrothermal or volcanic activity.

Jökulhlaups from Grímsvötn terminate abruptly, often within a few hours, once the water level has dropped about 100 m. This occurs before the lake is empty, when the ice overburden pressure at the rock rim exceeds the hydrostatic water pressure by 10–15 bars. The water level in the lake does not drop to 1150 m, the level of the subglacial caldera ridge.

3.2. Characteristics of jökulhlaups from Grímsvötn

Typically jökulhlaups from Grímsvötn increases approximately exponentially to the peak, and falls

more rapidly afterward (Fig. 6). The duration of large floods tends to be shorter than that of small floods. The smaller jökulhlaups reach their peak in 2–3 weeks and terminate about 1 week later. This discharge pattern can be explained by enlargement of a single basal ice tunnel due to melting by frictional heat generated in the flowing water. The drainage affects glacier sliding only over small localized areas. All such outburst floods drain through one main ice tunnel, which feeds Skeiðará. If discharge exceeds about 3000 m³ s⁻¹, water starts to drain from other, smaller tunnels at the central part of the terminus, collecting into the river Gígja. In the most voluminous floods, 10–15 high-capacity ice tunnels develop (Björnsson, 1998). The rapid injection of water creates high water pressures, as evidenced by water forced up to the glacier surface through crevasses that formed in 300-m-thick ice.

To date, modelling based on the theory of Nye (1976) and Clarke (1982) has succeeded in simulating the increasing discharge of several jökulhlaups from Grímsvötn but fails to adequately represent the sharp transition from increasing to decreasing discharge

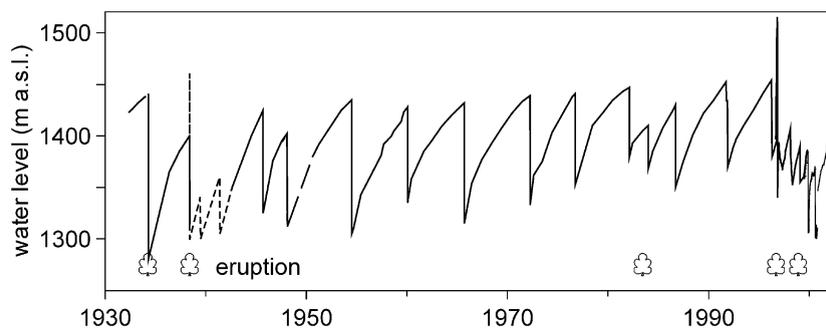


Fig. 5. The lake level of Grímsvötn 1930–2000. The lake level ascends until a jökulhlaup takes place. In 1996 the lake rose to the level required for floating of the ice dam.

Table 2

Data on typical jökulhlaups from subglacial lakes in Vatnajökull (Björnsson, 1988, 1992)

Source	Subglacial flowpath (km)	Drainage interval (year)	Peak discharge ($\text{m}^3 \text{s}^{-1}$)	Duration (days)	Volume (km^3)	Sediment load (10^6 ton)
Grímsvötn	50	1–10	$0.6–50 \times 10^3$	2–30	0.5–4	30–150
E-Skaftá cauldron	40	3	$0.2–1.5 \times 10^3$	2–15	0.1–0.3	30

(Björnsson, 1992). A cylindrical tunnel apparently cannot describe the fast closure of the conduit. Nye's (1976) prediction that the discharge Q rises asymptotically with time as $Q=(1/t)^4$ successfully described the rising limb of the jökulhlaup from Grímsvötn in 1972. Spring and Hutter (1981) applied the general equations, which include thermal energy stored in the lake to simulate the 1976 hydrograph and concluded the lake water would have been 4 °C. The lake temperature, however, was measured close to the melting

point (Björnsson, 1992). This supports the observation that the actual heat transfer by flowing water during jökulhlaups is more effective than accounted for by the empirical heat transfer equation described by Nye (1976) and Spring and Hutter (1981). Occasionally, jökulhlaup discharge has increased at a rate suggesting that thermal energy stored in the lake contributed to tunnel expansion. The rise of the jökulhlaup in 1934 can be simulated by lake water of 1 °C, and in 1938 by water of 4 °C when a peak of

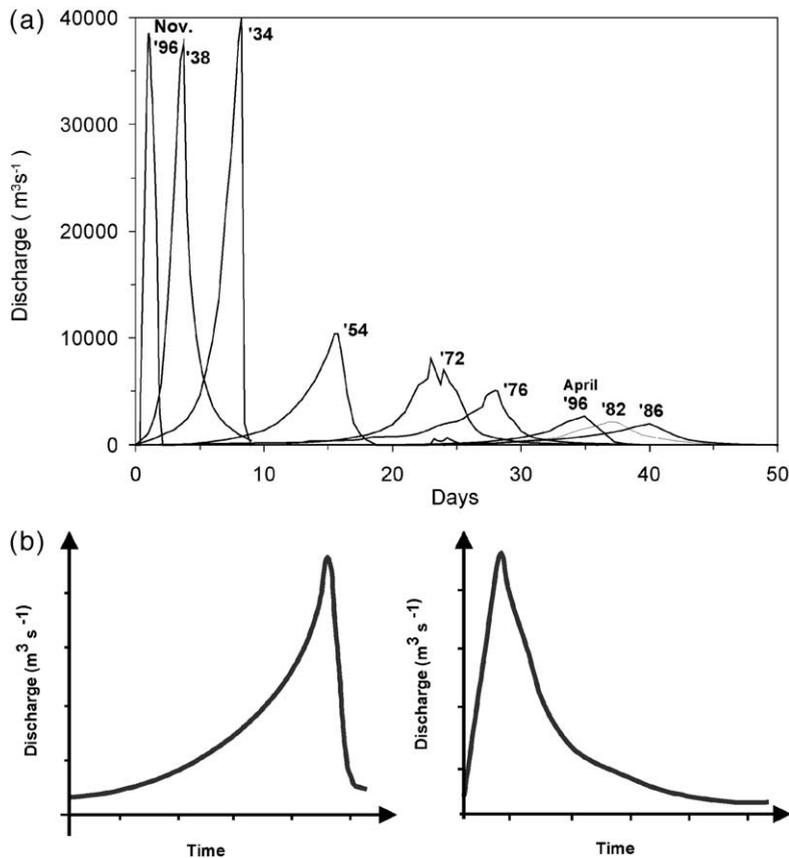


Fig. 6. (a) Typical hydrographs of jökulhlaups from Grímsvötn (1934, 1938, 1954, 1972, 1976, 1982, 1986 and 1996). (b) Left: typical shape of a hydrograph when a single basal ice tunnel enlarges due to melting. Right: Rapidly rising hydrograph that is not explained by the classical theory of jökulhlaups (typical for jökulhlaups from the Skaftá cauldrons).

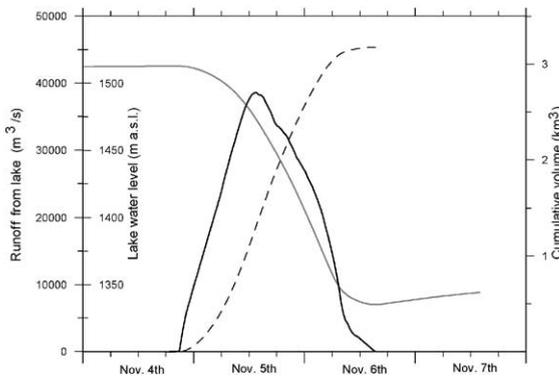


Fig. 7. Discharge hydrograph of the 1996 jökulhlaup from Grímsvötn, the cumulative volume of water drained and the subsidence of the lake level (measured by precision barometric altimetry).

$3 \times 10^5 \text{ m}^3 \text{ s}^{-1}$ was reached in 4 days (Fig. 6). Lake temperature values should not be taken seriously, however, as the heat transfer theory is questionable.

3.3. Extraordinary jökulhlaups from Grímsvötn

The jökulhlaup from Grímsvötn in November 1996 was of an extraordinary type, draining so rapidly that melting alone could not have expanded the conduits (Fig. 7). For the first time in the observational history of Grímsvötn, the ice dam was floated off the bed (Fig. 8). Downstream from the dam, water pressure exceeded the ice overburden and the glacier was lifted off the bed along the water flowpath (Björnsson,

1997; Roberts et al., 2000). Crevasses were observed along the main flowpath and icebergs were broken off the margin and spread over Skeidarársandur outwash plain. As the first jökulhlaup of this kind to take place after scientific observations began, it has provided new insight into jökulhlaup hydrology. Similar jökulhlaups may have occurred before, although descriptions do not provide unquestionable evidence for as rapid a rise in discharge (e.g. in 1861 and 1892; Thorarinsson, 1974; Björnsson, 1988).

The 1996 jökulhlaup was preceded and indirectly triggered by the Gjalp eruption, which took place inside the drainage basin of Grímsvötn (Figs. 1, 9 and 10). Meltwater was accumulated for a month until it drained in the catastrophic jökulhlaup. The eruption broke through the ice cover at one place after 30 hours of eruption, but continued subglacially for 2 weeks on a 6-km-long fissure. Melting by the eruption was measured both from the volume of the surface depressions above the eruptive fissure, and from the volume of meltwater accumulated in Grímsvötn (Guðmundsson et al., 1997; Einarsson et al., 1997). During the first 4 days meltwater was produced at a rate of $5000 \text{ m}^3 \text{ s}^{-1}$ and the heat output at the peak of the eruption was 10^{12} W (more than 100 times all the power stations producing electricity in Iceland at that time). The high rate of melting can only be due to fragmentation of the lava into glass (hyaloclastites) and rapid cooling of the fragments by quenching. The total volume of ice melted during the first 6 weeks after the beginning of the eruption was 4.0 km^3 , equivalent

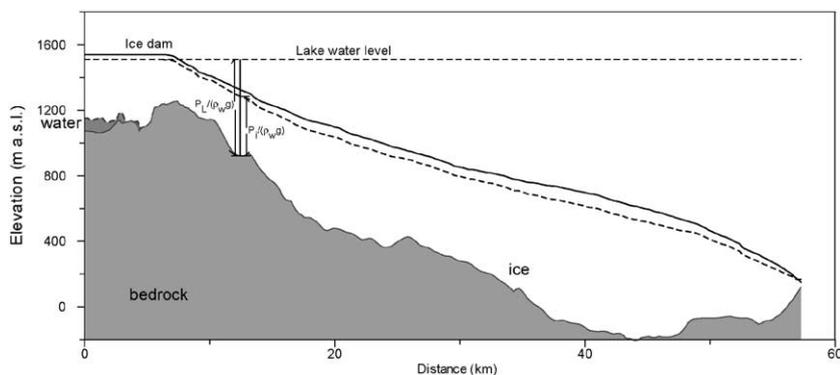


Fig. 8. A cross-section from Grímsvötn to Skeidarársandur along the flowpath of the jökulhlaup in 1996. Subglacial hydraulic conditions at the beginning of the 1996 jökulhlaup: $P_l/(\rho_w g)$ is the ice overburden pressure in metres of hydraulic head; $P_l/(\rho_w g)$ is the hydraulic head maintained by the lake. The ice dam was floated at the onset of the 1996 jökulhlaup.

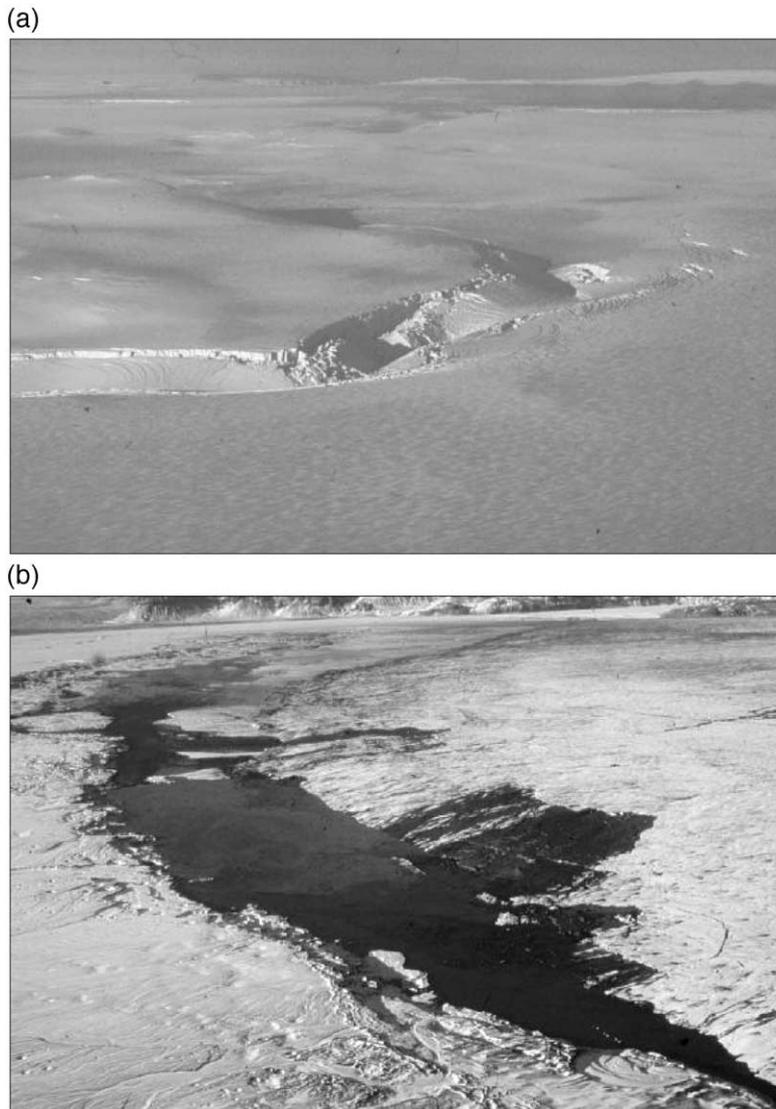


Fig. 9. (a) An oblique aerial photo of the 6-km-long, 2-km-wide and 100-m-deep depression formed above the flowpath of the 1996 jökulhlaup across the ice dam of Grímsvötn. (b) The margin of Skeiðarárjökull during the jökulhlaup of 1996. The flood water emerged from the base through crevasses 200 m higher than the margin.

to 1.1×10^{12} kg of magma cooling from 1000 to 0 °C if all the heat were used for melting. After 1 year (January 1998), the melted volume was 4.7 km³ (Fig. 10).

The meltwater accumulated in the Grímsvötn subglacial lake for a month until it drained in a catastrophic jökulhlaup on November 4–7 1996, in which 3.2 km³ of water drained from the lake within

a period of 40 hours. On November 4, the lake had risen to the level required for flotation of the ice dam, 1510 m, and ice-quakes marked the onset of lake drainage. About 10.5 hours later, water emerged from the margin of Skeiðarárjökull as a flood wave inundating Skeiðarársandur (at 100 m elevation), in the most rapid jökulhlaup ever recorded from Grímsvötn (Fig. 7). The discharge out of the lake during

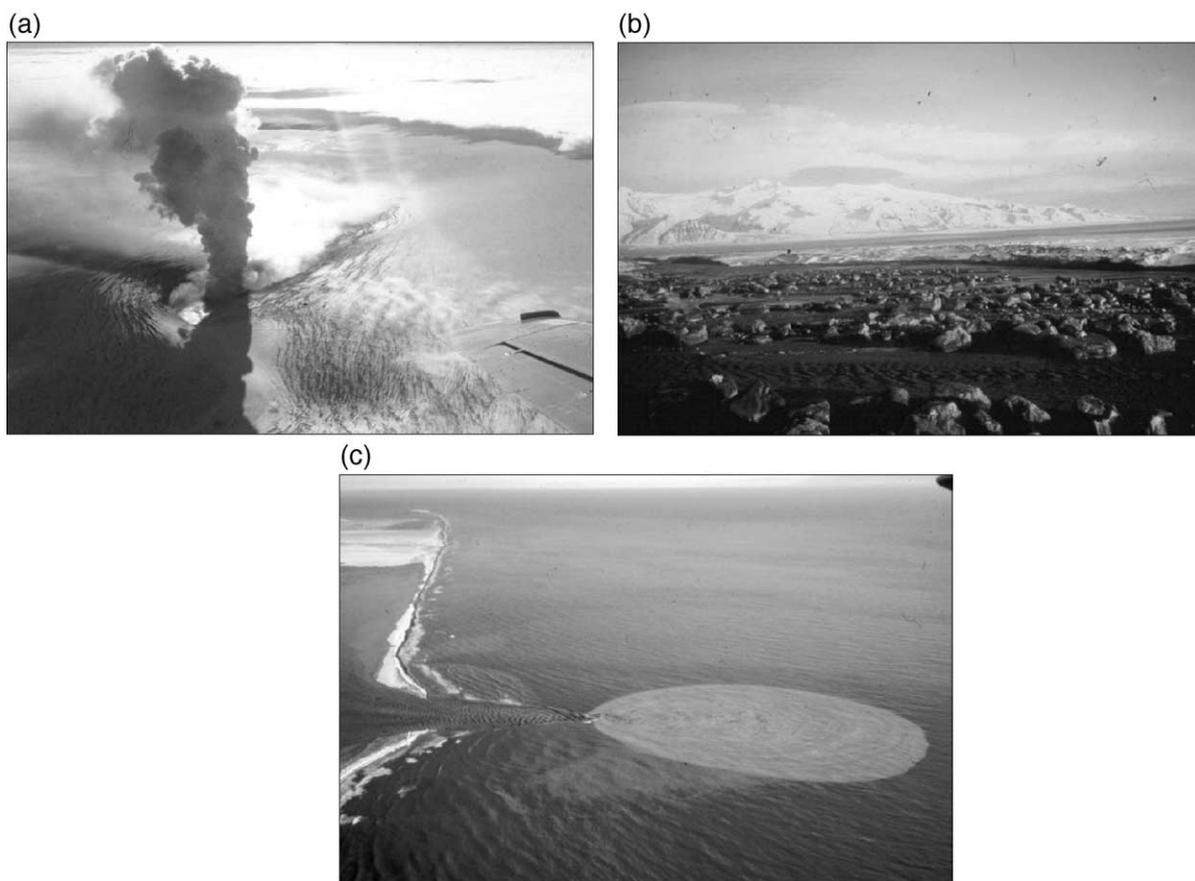


Fig. 10. Photos of the Gjalp eruption in October and the Grímsvötn jökulhlaup in November 1996. (a) A volcanic eruption through 500–700 m thick glacier. (b) Icebergs spread over the outwash plain Skeiðarársandur during the jökulhlaup. (c) The plume of suspended sediment from the Skeiðará jökulhlaup 1996 off the southern coast of Iceland. The coastal line moved seaward by about 900 m.

this jökulhlaup was derived directly from lake level observations and the known lake hypsometry. Discharge increased linearly with time and reached a peak value of $4 \times 10^4 \text{ m}^3 \text{ s}^{-1}$ in 16 hours and dropped to zero 27 hours later. The total volume of water released from the glacier was estimated at 3.2 km^3 . During the drainage the lake surface subsided by 175 m (Fig. 5), and the floating ice cover was reduced from 40 to less than 5 km^2 . While the discharge hydrograph for the lake was derived, the shape of the hydrograph for the outburst from Skeiðarárjökull is unknown and likely to be different since the flood was not drained through a single subglacial conduit. Four points on the hydrograph, two on the rising limb and two during the recession

were estimated but their accuracy is uncertain (Snorrason et al., 1997; Russel et al., 1999).

During the lake drainage, a 6-km-long, 1-km-wide and 100-m-deep depression was created by collapse of the jökulhlaup flowpath across the ice dam. The volume of the depression was 0.3 km^3 . Assuming that all of the thermal energy in the lake was used in the formation of the depression, we calculate the average temperature of the 3.2 km^3 of water released to be $8 \text{ }^\circ\text{C}$. In addition to this, we calculate that 0.1 km^3 of water was produced by frictional melting during the descent of the water from Grímsvötn to Skeiðarársandur.

When floodwater started to drain from the glacier margin a volume of 0.6 km^3 of water had accumulated under the glacier (Fig. 7). Melting enlargement

of the conduit by the frictional heat of the flowing water can only account for a portion (0.01 km^3) of the required conduit volume. Hence, lifting of the ice by water pressure in excess of the overburden took place during the beginning of the flood, prior to conduit formation. Longitudinal crevasses were observed above the entire flowpath. Approaching the glacier terminus, basal water burst out on the glacier surface through several hundred metres of ice. Icebergs were broken off the margin and spread over the outwash plain.

3.4. The Skaftá ice cauldrons and jökulhlaups

The two Skaftá cauldrons are situated over geothermal systems, 10–15 km to the northwest of Grímsvötn (Figs. 2 and 4b). Since 1955, at least 30 jökulhlaups have drained from these cauldrons to the river Skaftá. The period between these drainage events is about 2–3 years. Jökulhlaup discharge from the eastern Skaftá Cauldron typically rises rapidly and recedes slowly (Björnsson, 1977). The form of the hydrograph is a mirror image of the typical Grímsvötn hydrograph (Fig. 6). The peak discharge from the eastern cauldron is $1000\text{--}1500 \text{ m}^3 \text{ s}^{-1}$ and is reached in 1–3 days. These jökulhlaups recede slowly in 1–2 weeks and the total volume of water expelled is $200\text{--}350 \times 10^6 \text{ m}^3$ (Table 2). Theoretical simulations fail entirely to describe these hydrographs (Björnsson, 1992). The steep rise in discharge can be simulated with lake water well above the melting point ($10\text{--}20 \text{ }^\circ\text{C}$) but these values are derived from a flawed theory. Water at the melting point drains from the glacier. The rapid Skaftá jökulhlaups fall in the category that are not explained by the classical theory of jökulhlaups. Crevasses observed across the ice dam of the eastern cauldron after jökulhlaups suggest that the jökulhlaups start with flotation of the glacier (Fig. 4b). At the

time of the peak discharge, only 25% of the flood volume has drained out of the lake. The slow recession after the peak suggests that these floods do not drain through a single tunnel but spread out beneath the glacier and later slowly collect into the river outlet.

3.5. Jökulhlaups in Vatnajökull triggered directly by volcanic eruptions

Jökulhlaups triggered by volcanic eruptions can be traced to volcanic systems with the aid of maps of the subglacial topography (Björnsson and Einarsson, 1991) (Table 3). Most jökulhlaups from Vatnajökull have drained southward to Skeiðarársandur or northward to the river Jökulsá á Fjöllum (Thorarinsson, 1974; Björnsson, 1988; Björnsson and Einarsson, 1991). Major jökulhlaups occurred during two historical eruptions in the stratovolcano Öräfajökull (Fig. 2).

In general the volume of meltwater produced during subglacial eruptions is higher the longer the eruption remains entirely subglacial; hence, it tends to be higher the thicker the glacier and the longer melting of ice can be compensated by fast inflow of surrounding ice. This applies as long as the original ice thickness does not produce such high confining pressure that pillow lava is formed and the heat transfer is reduced. Two fissure eruptions under 500–700 m thick ice north of Grímsvötn, in 1938 and 1996, produced 2.7 km^3 ($0.3\text{--}0.5 \text{ km}^3$ of hylao-clastite) and 4.0 km^3 of meltwater, respectively (Björnsson, 1988; Guðmundsson and Björnsson, 1991; Guðmundsson et al., 1997). In contrast, eruptions inside lake Grímsvötn typically break through a 150–200 m thick ice shelf in 10–20 min, forming an opening 0.5 km in diameter and only melting small quantities of ice (about 0.1 km^3 , Guðmundsson and Björnsson, 1991). Eruptions within Grímsvötn do not

Table 3
Data on typical jökulhlaups triggered by volcanic eruptions (see references in Sections 3.5 and 4)

	Period (year)	Subglacial flowpath (km)	Peak discharge ($\text{m}^3 \text{ s}^{-1}$)	Duration (days)	Volume of meltwater (km^3)	Sediment load (10^6 ton)	Peak thermal output (W)	Total thermal energy (J)
Vatnajökull	5–30	50	$5 \times 10^3\text{--}1 \times 10^6$	2–30	3–5	100–300	1.7×10^{12}	10^{18}
Mýrdalsjökull	40–80	20	$1 \times 10^5\text{--}3 \times 10^5$	2–10	1–8	2000	$3 \times 10^{13}\text{--}10^{14}$	$10^{18}\text{--}10^{19}$

Thermal output of subglacial volcanic eruptions derived from production of meltwater. (Gjálp 1996, Katla 1918).

cause accumulation of water and increased pressure head, because melting of a floating ice cover does not add any mass to the system—hence they do not trigger jökulhlaups. The volume of volcanic material injected to the lake is negligible. However, jökulhlaups may trigger eruptions by the pressure release on top of the volcano subsequent to the lake level falling about 80–100 m (Thorarinsson, 1974; Guðmundsson et al., 1995). Seismic records suggest that this mechanism may have triggered several small but invisible eruptions under the Skaftá cauldrons after 1985 (Páll Einarsson, personal communication). Pressure release during a jökulhlaup may also cause explosive boiling in a subglacial hydrothermal area.

The largest and most catastrophic jökulhlaups in Iceland may be caused by eruptions in the voluminous, ice-filled calderas of Bárðarbunga and Kverkfjöll in northern Vatnajökull (see Björnsson, 1988; Björnsson and Einarsson, 1991). These calderas may be the source of prehistoric jökulhlaups in Jökulsá á Fjöllum, with estimated peak discharges of up to 4×10^5 to 1×10^6 $\text{m}^3 \text{s}^{-1}$ (Tómasson, 1973; Knudsen and Russell, 2002) that swept down Jökulsá á Fjöllum and carved a scablands and deep canyons (Jökulsárgljúfur).

Jökulhlaups during eruptions in steep ice and snow covered stratovolcanoes are swift and dangerous and may become lahars and debris-laden floods. The first contemporary description of a jökulhlaup during a volcanic eruption in Iceland dates from the 1362 eruption of the ice capped stratovolcano Örfajökull (Thorarinsson, 1958). The flood was over in less than 1 day and the peak may have been greater than 10^5 $\text{m}^3 \text{s}^{-1}$ (Thorarinsson, 1958, p. 36). The meltwater originated in the 500 m deep caldera summit area at 2000 m elevation and drained beneath the outlet glaciers. Sediment, hummocks (2–4 m high) and dead ice were spread over the lowland and into the sea, and a plain was formed where there had been thirty fathoms water (Skálholt Annal). Several farmsteads were washed away. A contemporary record described “several floods of water that gushed out, the last of which was the greatest. When these floods were over the glacier itself slid forwards over the plane ground, just like melted metal poured out of a crucible... The water now rushed down on the earth side without intermission, and destroyed what little of the pasture grounds remained... Things now assumed quite a different appearance. The glacier itself burst and many

icebergs were run down quite to the sea, but the thickest remained on the plain at a short distance from the foot of the mountain. . . we could only proceed with the utmost danger, as there was no other way except between the ice-mountain and the glacier that had slid forwards over the plain, where the water was so hot that the horses almost got unmanageable” (Report by Rev. Jón Thorláksson, cited by Henderson, 1818; Thorarinsson, 1958, p. 31). Obviously, the water conduits could not adjust to this rapid injection of meltwater, and increased water pressure reduced basal friction and facilitated sliding. A less devastating eruption took place in 1727 A.D. (Thorarinsson, 1958). The sediment and icebergs that were transported down to the lowland were later named glaciers and these place names are still used centuries after all the ice has melted away.

Similar rapid floods carrying volcanic material were produced from Eyjafjallajökull (in 1612 and 1821–1823), and during the eruption at Mt. Hekla in 1845, 1947 (Kjartansson, 1951), 1970 and 1981 in the river Rangá.

4. Subglacial lakes, volcanic eruptions and jökulhlaups in Mýrdalsjökull

Mýrdalsjökull can be divided into three major catchment basins that supply water to the outwash plains Mýrdalssandur, Sólheimasandur and Markarfljótsaurar, respectively (Fig. 2b). The glacier surface has 12 small depressions that have been created by subglacial hydrothermal activity. These ice cauldrons are typically 20–50 m deep and 500–1000 m wide. A persistent hydrogen sulphide odour from Jökulsá á Sólheimasandi indicates that meltwater is drained continuously. Meltwater accumulates beneath some of the cauldrons and frequently drains in small jökulhlaups (Rist, 1967; Björnsson et al., 2000; Russel et al., 2000).

Katla eruptions rapidly melt large volumes of ice, triggering enormous jökulhlaups and breaking off large blocks of ice from the glacier margin. Meltwater may also drain out supraglacially through crevasses and moulins. During the jökulhlaups, a mixture of water, ice, volcanic products and sediment, surges over the outwash plain. Water-transported volcanic debris has been estimated from 0.7 to 1.6 km^3 or on the order of 2×10^9 tons per event (Tómasson, 1996; Larsen,

2000). During the most violent jökulhlaups such large quantities of icebergs and sediment were clustered together that some places on the outwash plain were named for glaciers. The peak discharge of $1-3 \times 10^5 \text{ m}^3 \text{ s}^{-1}$ is reached in a few hours, and total volumes of $1-8 \text{ km}^3$ draining over 3–5 days have been suggested (Table 3; Thorarinsson, 1957, 1974; Tómasson, 1996; Larsen, 2000). The rate of increase of discharge and peak discharges are an order of magnitude higher than for any known jökulhlaup from subglacial lakes. These jökulhlaups, along with heavy fallout of tephra, make Mýrdalsjökull the most hazardous volcano in Iceland.

Fascinatingly, the Katla eruptions break through an ice cover of 400 m in 1 or 2 h. Nonetheless, the melting rate during the most violent Katla eruptions is an order of magnitude higher than during the Gjalp eruption in 1996 (Table 3, Guðmundsson et al., 1997). The heat flux of Katla may be overestimated because the peak jökulhlaup discharges are higher than the production rate of meltwater. Water may be stored at the eruption site, either at the beginning of the eruption because the conduits cannot accommodate the flow, or over a longer period due to increased heat flux from the volcano and subsequent formation of a depression in the surface. Eruptions of magma into such a subglacial water body would make possible rapid heat transfer during the eruption and ensure rapid cooling by fragmentation of the lava into glass.

Mýrdalssandur and the adjacent Sólheimasandur and Skógasandur have to a large extent been built in floods caused by volcanic eruptions of the Katla volcano. During 18 of the 20 documented eruptions the associated jökulhlaups flowed southeast down to the Mýrdalssandur outwash plain, but in two cases jökulhlaups flowed southwest to the Sólheimasandur outwash plain. During the last eruption in 1918, the jökulhlaup moved the coastline seawards by 3 km (Tómasson, 1996), and the coastline now lies 2.2–2.5 km further south than it did in 1660. Marine sediments found several hundred kilometres south of Iceland, contain debris transported from the eruption sites (Guðrún Helgadóttir and Haraldur Sigurdsson, personal communication). The most rapid jökulhlaups have produced flood waves in coastal waters. Jökulhlaups from the northern part of Mýrdalsjökull have eroded canyons in Markarfljót River (Sigurðsson, 1988). The most recent jökulhlaup taking that route was 1600 years ago (Haraldsson, 1981; Sigurðsson, 1988).

5. Conclusions

Of our main findings of the glacier volcano interactions in Iceland, we can summarize the following conclusions.

Volcanic eruptions to the glacier bed are the most violent expressions of glaciovolcanism, rapidly melting ice and producing hazardous floods outside the glacier. Subglacial hydrothermal systems are quieter expressions, generated by interaction of glacial meltwater with magma intrusions in the Earth's crust. The hydrothermal activity continuously melts ice at the glacier bed, creating a depression in the glacier surface. The meltwater drains either continuously to the glacier margin or accumulates in subglacial lakes that are situated beneath the glacier surface depressions. Some of the lakes are located in caldera depressions but they rise as a dome above the caldera rims due to the relatively low fluid pressure under the glacier depression. The release of water from the subglacial lakes can take place by two different conduit initiation mechanisms. Drainage begins either through conduits at pressures lower than the ice overburden at the ice dam (in Grímsvötn, typically 6–7 bars lower) or the lake level rises until the ice dam is floated. The subsequent drainage from the lake can take place by two different modes. The conduits may enlarge over days or weeks due to melting of the ice walls by frictional heat in the flowing water and stored lake heat. This is typical for the jökulhlaups initiated at lake levels lower than that required for flotation. These processes have been explained by classical jökulhlaup theories. Alternatively, the discharge rises faster than can be accommodated by melting of the conduits and the glacier is lifted along the flowpath to make space for the water. A preliminary modeling study suggests that the idea of a sheet flood preceding conduit drainage is plausible.

During rapidly rising jökulhlaups, water may drain through braided watercourses at high pressures. However, these passageways quickly develop into high-capacity ice tunnels, and jökulhlaups are not known to have led to surges of glaciers. Icebergs may be broken off the glacier margins and transported by water over the river courses. During eruptions in steep ice-capped stratovolcanoes (like Öræfajökull) distributed drainage of meltwater has led to sliding of the glacier down to the lowland.

Processes and products of glacier volcano interaction are displayed on a broad range of dimensions. Hydrothermal systems affect drainage areas of 5–200 km² and the area of the subglacial lakes is from 1 to 40 km² in area. The production rate of basal glacial meltwater spans over four to five orders of magnitude, from 2 to 6 m³ s⁻¹ in hydrothermal systems to 5 × 10³ to 10⁵ m³ s⁻¹ in volcanic eruptions (requiring a heat flux of 10³–10⁸ MW). The lakes vary in volume by three orders of magnitude (lakes beneath cauldrons of 2 × 10⁹ to 4 × 10¹² m³ in Grímsvötn). The lakes drain periodically with an interval of 1–10 years. The duration of the jökulhlaups may be from 2–3 days to 2–3 weeks. The peak discharge in jökulhlaups can be from 200 to 10⁶ m³ s⁻¹. Ordinary jökulhlaups from subglacial lakes may transport of the order of 10⁷ tons of sediment but during the most violent volcanic eruptions the sediment load has been 10⁸ tons.

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