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JÖKULHLAUPS IN ICELAND: SOURCES, RELEASE AND DRAINAGE

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Abstract

Jökulhlaups in Iceland may originate from marginal or subglacial sources of water melted by atmospheric processes, permanent geothermal heat or volcanic eruptions. 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. During the 20th century 15 subglacial volcanic eruptions (10 major and 5 minor events) took place, about one-third of all eruptions in Iceland during that century.

The release of meltwater from glacial lakes can take place as a result of two different conduit initiation mechanisms and the subsequent drainage from the lake occurs by two different modes. Drainage can begin at pressures lower than the ice overburden in conduits that expand slowly over days or weeks due to melting of the ice walls by frictional and sensible heat in the water. Alternatively, the lake level may rise until the glacier is lifted along the flowpath to make space for the water and water discharges rise linearly, peaking in a time interval of several hours to 1-2 days. In this case, discharge rises faster than can be accommodated by melting of the conduits. The rapidly-rising floods are often associated with large discharges and floods following rapid filling of subglacial lakes during subglacial eruptions or dumping of one marginal lake into another. Jökulhlaups during eruptions in steep ice and snow-covered stratovolcanoes are swift and dangerous and may become lahars and debris-laden floods. Normally jökulhlaups do not lead to glacier surges but eruptions in ice-capped stratovolcanoes have caused rapid and extensive glacier sliding.

Jökulhlaups 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. 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. 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.

Iceland is a unique and valuable study-site for glaciovolcanic interactions. 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. The jökulhlaups can be seen as modern analogues of past megafloods on the Earth and their exploration may improve understanding of ice-volcano processes on other planets.



CHAP.8 - MOUNTAINS GEOMORPHOLOGY p.164-184. eds. Philip Owens and Olav Slaymaker). Arnold Publishers.

GLACIAL LAKE OUTBURST FLOODS IN MOUNTAIN ENVIRONMENTS

Helgi Björnsson

1 Introduction

Floods can happen where glaciers dam lakes in mountainous areas. Occurring in both the temperate and subpolar regions of the Earth, such floods are called *débâcles* in the European Alps, *aluviones* in South America and *jökulhlaups* in Iceland. A dam of ice or sediment blocks the water to form a lake, while drainage is initiated by an opening of the hydraulic seal, which can be broken either suddenly or gradually. The impounded water is released directly into river channels, and has a typical discharge that is orders of magnitude higher than when running as a direct result of intense ablation. In regions with active, ice-covered volcanoes, meltwater is repeatedly released in floods from lakes that collect at subglacial hydrothermal areas. Occasionally, these floods take place without any significant prior storage of water, since ice melts instantaneously in volcanic eruptions. During the largest glacial floods in history, discharges reached $106 \text{ m}^3 \text{ s}^{-1}$. For comparison, the meltwater that was temporarily stored in lakes at the edges of downwasting Pleistocene ice sheets was released in bursts that were only one order of magnitude larger than this, even though the enormous volumes involved may have altered the circulation of deep water in the North Atlantic Ocean of the late Pleistocene era.

Glacial outbursts can have pronounced geomorphological impacts, since they scour river courses and inundate floodplains. Outbursts result in enormous erosion, for they carry huge loads of sediment and imprint the landscape, past and present, with deep canyons, channelled scabland, ridges standing parallel to the direction of flow, sediment deposited on outwash plains, coarse boulders strewn along riverbanks, kettleholes where massive ice blocks have become stranded and have melted, and breached terminal moraines. Some modern outbursts have produced flood waves in coastal waters (tsunamis). In the North Atlantic, outburst sediments dumped onto the continental shelf and slope have been transported great distances by turbid currents. Outburst floods wreak havoc along their paths, threatening people and livestock, destroying vegetated lowlands, devastating farms, disrupting infrastructure such as roads, bridges and power lines, and threatening hydroelectric plants on glacially fed rivers.

Knowledge of jökulhlaup behaviour is essential for recognizing potential or imminent hazards, predicting and warning of occurrences, enacting preventive measures, assessing consequences and responding for the purpose of civil defence. The goal of this chapter is to outline and describe the following:

- (1) the location and geometry of glacial flood sources, including the properties of dams that impound the water;
- (2) the accumulation of water leading to outbursts and the conditions in which they begin;
- (3) the mechanisms and discharge characteristics of outbursts; and
- (4) case histories of floods, illustrating potential hazards.

2 Anatomy of glacial lakes and their outburst floods

2.1 Sources of glacial outbursts

Glacial lakes are found in various topographic settings. Meltwater may be impounded within the glacier (englacially and subglacially), on the surface of the glacier (supraglacially) or in lakes formed by dams at the glacier's margin (proglacially). The general characteristics, geometry and filling of glacial lakes can be described by the basic physics of hydrology. The movement of a thin film of water along the glacier base follows the gradient 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 , assuming the effects of kinetic energy to be negligible. The symbol w represents the density of water and equals 1000 kg m^{-3} , whereas g equals 9.81 ms^{-2} and represents the acceleration due to gravity, while z_b is the elevation of the glacier substrate above sea level (Figure 8.1). When describing the regional flow of basal water, it has proved useful to present a static approximation of water pressure, equal to the overburden pressure from the ice, where i equals 916 kg m^{-3} and represents the density of ice, H equals $z_s - z_b$ and is the thickness of the glacier and z_s is the elevation of the ice surface in relation to sea level (Shreve, 1972). Thus the gradient driving the water is

$$\nabla \phi_b = (\rho_w - \rho_i) g \nabla z_b + g \rho_i \nabla z_s \quad (2)$$

This formula predicts that the surface slope of the glacier is some ten times more influential than the slope of its bed in determining the flow of basal water. Glacial lakes may be situated at points where water flows towards a minimum in the total fluid potential. Trapped in such a lake, water is surrounded by a hydraulic seal. As the overlying glacier floats in a state of static equilibrium, and the vertical force balance is reflected by the shape of the lake. The roof of a subglacial lake slopes approximately ten times more steeply than the surface of the ice, and in the opposite direction (Figure 8.2). While resembling a grounded ice shelf, the ice on the perimeter of the lake essentially comprises an ice dam. The state in which there is no gradient driving water inside the lake,

$\nabla \phi_b = 0$, can be used to define the location and geometry of a subglacial reservoir. Hence describes the



relationship of the slope in the ice/water outline of the lake to the slope at the upper surface of the glacier (note that there might also be lakes with equal inflow and outflow which would not meet the condition of a zero gradient in relation to fluid potential). Given the glacier's surface and the bedrock geometry, it is possible to assess whether a water reservoir is likely to exist on the glacier bed.

$$\nabla z_b = -[\rho_i/(\rho_w - \rho_i)] \nabla z_s \quad (3)$$

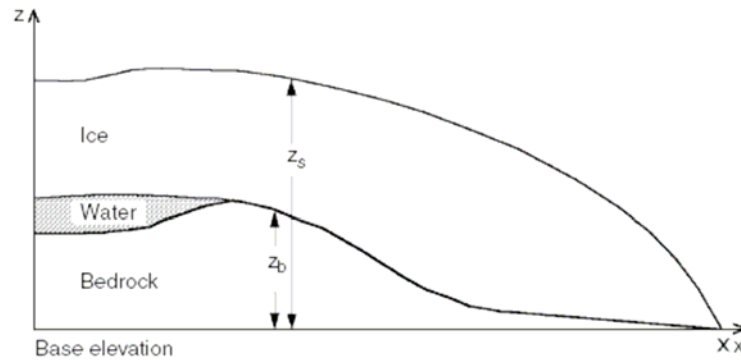


Figure 8.1 Parameters for describing the principles of water flow in glaciers

The general relationship in eq. (3) applies to all types of glacial lakes (Figure 8.2, Table 8.1). Water may gather in a bedrock hollow beneath a relatively flat or dome-shaped glacial surface, (Figure 8.2(a)) – in a type of reservoir which has been delineated beneath today's Antarctic Ice Sheet (Oswald and Robin, 1973; Ridley *et al.*, 1993; Siegert *et al.*, 2001) and was presumably widespread during the Quaternary glaciations. Another type of lake can form a cupola above the glacier bed, accompanied by a depression in the glacier's surface (Figure 8.2(b),(c)). Subglacial water reservoirs of this sort are common in Iceland, owing to volcanic and geothermal activity which causes the ice to melt and thereby creates depressions in the glacier surface (Björnsson, 1974, 1988). Meltwater flows along the bed and accumulates beneath the depression.

Subaerial glacial lakes (marginal or proglacial lakes) are confined by ice on one side and by topographic barriers on the other, for instance by the edges of ravines, riverbeds or main valleys (Figure 8.2(d)). The glacier's surface slopes toward the lake, which stores meltwater from the glacier along with runoff from the surrounding terrain. The lakes formed during the summer at the margins of subpolar glaciers may be sealed by an ice dam frozen to the bed (e.g., Maag, 1969).

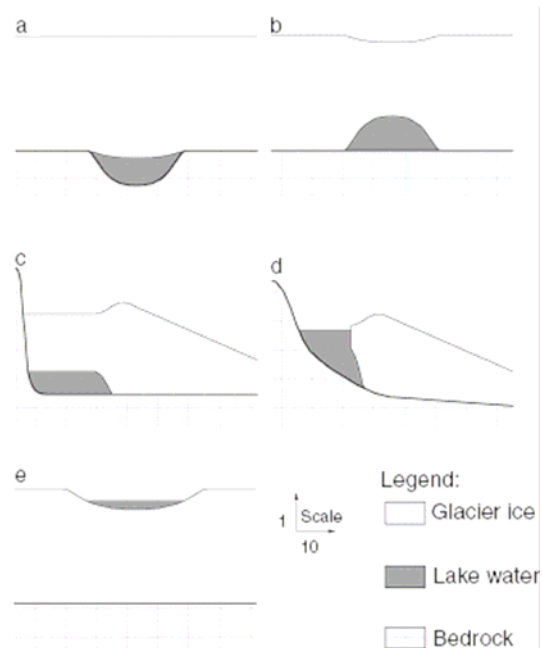


Figure 8.2 Schematic diagram of the main types of glacial lake

In other instances, a proglacial lake may lie in a depression behind moraine at the front of the glacier, which is typically retreating rapidly, while sediment is deposited too slowly to fill up the over-deepened bed. This happened in the case of the receding late Pleistocene ice sheets and applies to many shrinking twentieth century glaciers. Supraglacial lakes are isolated in depressions on the glacier surface (Figure 8.2(e)), hydrologically separated from the basal drainage system. A final addition to our list might be water stored in cavities located here and there inside the glacier (Kamb *et al.*, 1985; Walder, 1986; Kamb, 1987).



2.2 Drainage of glacial lakes

In general, glacial lakes of every type can drain either continuously or in episodic bursts. Marginal and proglacial lakes sometimes drain constantly via subaerial spillways over the rock or sediment barrier that contains them. Such barriers might consist of moraine, volcanic material or landslides. Nevertheless, lakes held back by these materials may break forth abruptly if the dams fail. Floods released suddenly when sediment or ice impoundments are breached rise almost linearly over a period of minutes or hours. They reach a high, sharp discharge peak and then fall along a steep recession limb (Figure 8.3(b)). Sediment dams can fail because of overtopping or accelerated fluvial erosion, which results in retrogressive incision enlarging the point of outflow. Moreover, a break in sediment dams can be caused by seepage and piping (progressive groundwater movement) within the sediment barrier, resulting in liquefied flows and embankment slips that weaken or disintegrate the dam. Tectonic activity and landslides are further threats to sediment dams, while breaching may also be triggered by flood water entering the lake or by heavy rain leading to a rapid rise in the lake level.

Finally, dam failure can be generated by waves from landslides, rock falls, ice avalanches, calving icebergs or glacier surges into the lake. Providing the topographic barriers hold, on the other hand, proglacial lakes serve to dampen jökulhlaups from other areas. Any outburst that does occur may be a one-time event if it ruins the dam. Where the barrier is a wall of ice, it may break down mechanically, similarly to a sediment dam. Marginal lakes at subpolar glaciers, where the ice barricade is frozen to the bed, are typically breached as lake water spills over the top of the dam into a supraglacial channel that melts into a bigger breach – commonly at the juxtaposition of the glacier and a rock wall (e.g., Schytt, 1956; Maag, 1969). The rate of ice melt, the level of the lake in relation to the outlet and the hypsometry of the water reservoir dictate the progress of lake drainage. As the lake surface falls, the breach may broaden and deepen, even undercutting the ice wall so as to cause calving. When subglacial lakes are situated in bedrock hollows under flat or dome-shaped glacial surfaces, they are not expected to expand and drain in sudden outbursts, except perhaps during rapid deglaciation which drastically alters glacier geometry and drainage (Goodwin, 1988). In contrast, when icedammed lakes, regardless of whether they are positioned subglacially or marginally, receive water inflow and are gradually made to expand, basal water pressure will increase and the overlying ice will be raised. Eventually, the hydraulic seal of the ice dam will be ruptured, so that water will begin to drain from the lake via a basal pathway. Seepage beginning beneath the ice blockage causes enlargement of the drainage system, initiating a flood under the surrounding ice. In a typical case, though for reasons not yet fully understood, leakage starts before the lake has reached the level which would cause flotation of the ice dam (Björnsson, 1975, 1988). After discharge has begun, pressure from the ice constricts the passageway, and water flow at an early stage in the jökulhlaup correlates primarily with enlargement of the ice tunnel owing to heat from friction against the flowing water and to thermal energy stored in the lake (Nye, 1976; Spring and Hutter, 1981, 1982; Clarke, 1982; Björnsson, 1992). Increasing as an approximate exponent of time over a matter of hours or days, the discharge falls quickly after peaking (Figure 8.3(a)). The recession stage of the hydrograph sets in when tunnel deformation begins to exceed enlargement by melting. The overlying ice may collapse abruptly into the tunnel and seal the lake again, sometimes before it is empty, with new water beginning to accumulate until another jökulhlaup occurs. The timing of bursts depends on what lake level will provide the subglacial water pressure necessary for breaking the hydraulic seal affected by the overburden pressure at the ice dam; therefore, if long-term records of lake levels are available, the elevation at which a jökulhlaup will begin can be predicted with some precision. Since the frequency of outbursts depends on the rate at which a lake is filled, marginal lakes often experience them at the end of the ablation season, when meltwater storage climaxes. Fluctuations in the thickness of the blocking ice, resulting from climatic variations or surges, may modify the outburst cycle or even stop bursts completely. Occasionally, jökulhlaups are triggered by flotation of the ice dam. Rather than initial drainage from the lake being localized in one narrow conduit, the water is suddenly released as a sheet flow, surging downhill and propagating a subglacial pressure wave, which exceeds the ice overburden and lifts the glacier in order to create space for the water. In this instance, discharge increases faster than can be explained by conduits expanding through melting (Björnsson, 1992, 1997, 2002; Björnsson *et al.*, 2001; Jóhannesson, 2002). The resulting hydrograph (Figure 8.3(b)) presents a rapid rise to its peak, succeeded by a more gentle waning stage. While sheet flows may be followed by drainage through high-capacity conduits, swift floods may instead manage to cause hydrofracturing and force their way englacially from the base of the glacier to its surface (Liestøl, 1977; Roberts *et al.*, 2000; Russell *et al.*, 2000; Roberts, 2002).

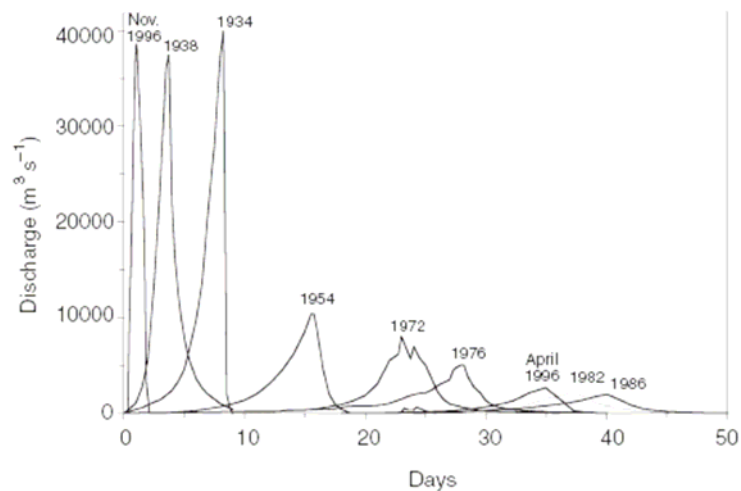


Figure 8.4 Discharge hydrograph of various jökulhlaups from Grímsvötn, Iceland

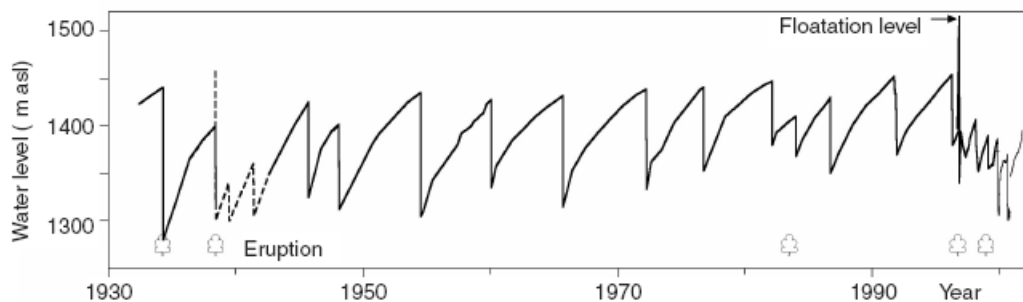


Figure 8.5 The water level of lake Grímsvötn, 1930–2000. The crater level rises until a jökulhlaup takes place. In 1996 the lake rose to the level required for floatation of the ice dam

Supraglacial lakes, on the other hand, are usually transient features (e.g., Björnsson, 1976), because they form englacial channels and drain down to the subglacial water system. Water that collects in lakes on polar ice masses frequently freezes and unites with the ice during the winter. Where glacial surges occur, floods may come about when a surge terminates and drainage of the water from subglacial cavities switches from a distributed system – conducting water slowly downglacier – to a quickly effective network of tunnels (Kamb *et al.*, 1985; Kamb, 1987). While the outlets for these bodies of water are not fixed, they may be expected to empty into rivers running from the glacier before the surge. As a final case, there have been reports of frequent abrupt releases of water, lasting from minutes to some hours, from diminutive amounts of subglacial or englacial storage, but the nature of these releases is poorly described (Haeberli, 1983; Drieger and Fountain, 1989; Walder and Diederger, 1995).

2.3 Impacts of jökulhlaups

The impact of outbursts corresponds to their discharge rate, the peak and nature of the flow, sediment load, routeway topography and frequency. Depending on the concentration of sediment, jökulhlaups range across a continuum from water floods to hyperconcentrated flows and debris flows. Most jökulhlaups from lakes dammed by ice contain relatively small concentrations of sediment (<40% by weight and 20% by volume) and can be described adequately in terms of standard hydraulics in turbulent water flow (calculating flow by Manning's equation and sediment transport by Einstein's equation). In the most catastrophic floods, intense erosion may occur through cavitation; the enormous velocity of flow reduces absolute pressure, causing air bubbles to form, which collapse and generate shock waves. Conditions for this are favourable in constricted channels at low water depths, producing channelled scabland features, whereas more uniform flow produces a topography of subdued erosion.

Jökulhlaups triggered by the failure of a sediment dam or by volcanic eruptions may temporarily contain high concentrations of sediment. Hyperconcentrated jökulhlaups represent moderately turbulent to laminar flow of non-Newtonian fluids, with sediment concentrations varying between 40% and 70% by weight (20% and 47% by volume). Debris flows are laminar, containing a uniform sediment concentration ranging from 70% to 90% by weight and from 47% to 77% by volume (Costa, 1988). Hyperconcentrated flows produce crudely stratified, less finely sorted deposits than water floods, and the deposits from debris flows are composed of poorly sorted clasts. River courses and landforms on outwash plains have adjusted to the magnitude of frequent or cyclic jökulhlaups, i.e., those occurring with annual regularity or at intervals of several years. During rare extensive floods, however,



erosion, transportation and downstream deposition modify river courses extensively. During the aftermath, flood channels remain destabilized, which may lead to slumping and debris flows.

3 Estimating the discharge of jökulhlaups

3.1 Measured and calculated discharge

Direct measurements of water current velocity and discharge are usually precluded in major jökulhlaups. Conventional stream-gauging stations are rarely set up for recording extreme floods, and even if the streams have been gauged no developed curves are available for comparing stage and streamflow. Large-scale floods overflow the normal riverbanks, distort river courses and damage or destroy instruments. Although a number of jökulhlaups have been gauged some distance downstream from glacier margins, the flood wave may have been significantly attenuated in the intervening interval by passing through such storage bodies as braided channels or lakes.

Indirect methods can be applied to estimate average stream velocity, using hydraulic calculations along with field evidence of channel geometry (including cross-sectional area, heights and dates of flood levels, and channel roughness). This approach employs Manning's equation, which was developed empirically for open, canalized waterways and which relates energy dissipation to the roughness of flood paths. In such an approximation, the surface slope at high water becomes the basis for determining the energy slope. Because of nonuniform flow, computations must take into account the conservation of mass and energy as the flood moves forward (Webb and Jarrett, 2002). The maximum flood stages are reconstructed by inspecting eroded channel margins, finding the highest flotation elevation of the largest boulders that were encased by ice and drifted downstream (ice-rafted erratics), and locating water divides where floodwater spilled over cols. The reconstruction of flood discharge obviously may be complicated when sedimentation at the end of the flood changes the area of the channel cross-section as well as the roughness of the flood course.

3.2 Prediction of discharge by theoretical and empirical methods

For the purpose of risk assessment, both empirical and theoretical methods can be relevant to predicting the probable rate of change of discharge, peak discharge and duration of jökulhlaups.

3.2.1 Subglacial drainage

Theoretical models of jökulhlaups discharging through water tunnels in the glacier base have been derived by Nye (1976), Spring and Hutter (1981) and Clarke (1982). These models are based on the physics of mass continuity, momentum, energy conservation and heat transfer, and they describe how water is driven by a fluid potential gradient through a tunnel of given roughness. From his general model, Nye (1976) derived an analytical solution, which predicted discharge, Q , to rise asymptotically with time as $Q \propto t^{-1/4}$, if confinement by the overburden was neglected and expansion of the ice tunnel was solely attributed to the instantaneous transfer of frictional heat (i.e., the dissipation of potential energy) from the flowing water to the enclosing ice. Numerical models for simulating the entire hydrograph were derived by Spring and Hutter (1981), who included lake temperature, and Clarke (1982), who added lake geometry. Clarke (1982) successfully simulated jökulhlaups from Lake Hazard (the Yukon, Canada) and Summit Lake (British Columbia, Canada), which are dammed by marginal ice. Björnsson (1992) tested Clarke's (1982) modification of Nye's (1976) general model, applying it to Icelandic jökulhlaups and concluding that the amended model in some respects successfully simulated jökulhlaups from the subglacial lake Grímsvötn in Vatnajökull. In general, the ascending slopes simulated in the graph corresponded to the measured ones. The peak in the computed hydrographs, however, was not as sharp as the actual climaxes, rendering the simulation of the descending limbs unsatisfactory. This presumably happens because, in the model, the tunnel is assumed to be cylindrical. It was only possible to simulate the rapid rise of some outbursts in Iceland (from subglacial as well as marginal lakes) by assuming a lake temperature several degrees above the melting point, which may suggest that thermal energy stored in these bodies of water contributes to tunnel expansion and thereby affects the discharge rate. The values computed for lake temperatures, however, should not be taken seriously, as the theory of heat transfer is questionable. The simulation of such swift jökulhlaups completely failed to illustrate the closure of tunnels and the recession of discharge.

An empirically based regression relation between the peak discharge Q_{max} (in $m^3 s^{-1}$) and the total volume V_t (in $10^6 m^3$) of water passing through subglacial tunnels in jökulhlaups was formulated by Clague and Mathews (1973):

$$Q_{max} = K V_t^b \quad (4)$$

For jökulhlaups emerging from ten marginal ice-dammed lakes, which ranged in volume from 10^6 to $10^9 m^3$, Clague and Mathews obtained $K=75/s m^{-0.99}$ and the power coefficient $b=0.67$. A recent update for 26 marginal ice-dammed lakes throughout the world yielded $K=46 s^{-1} m^{-1.02}$ and $b=0.66$ (Walder and Costa, 1996). Referring to eleven jökulhlaups from the Icelandic lake Grímsvötn, Björnsson (1992) found $K=4.15 \cdot 10^3 /s m^{-2.52}$ and the power coefficient $b=1.84$. Clarke (1982) derived theoretical relationships for predicting peak discharge, suggesting that b depended on the lake's contribution of stored thermal energy in proportion to the frictional energy dissipated in the jökulhlaup. The outcome of his calculations was that $b=0.8$ when the thermal energy originating in the lake was dominant in melting the tunnel and $b=1.33$ when its share was negligible. Ng and Björnsson (2003) point out that



incomplete lake draining, a phenomenon contrasting with common behaviour in marginal lakes dammed by ice, may explain the deviation of Grímsvötn from statistically derived value.

Table 8.1 – Sources of glacial floods and characteristics of drainage^a. p_w stands for water pressure and p_i for ice overburden pressure

Source/type of lake	Flood initiation and drainage routes				
	Subglacial drainage	Englacial drainage	Subaerial drainage (overtopping, downcutting, dam failure)		
Marginal lake ¹	Through ice tunnels at	Upwards from the base, creating outlets by hydrofracturing the ice and retrofeeding moulins at	Drainage over a subaerial breach	Fluvial erosion of sediment dams; piping	Mechanical failure caused by tectonic activity; rockfalls; landslides
Subglacial eruption, often in the absence of significant storage ²	$p_w < p_i$ or in a sheet flow at				
Subglacial lakes in hydrothermal areas ³	$p_w = p_i$, propagating a flood wave at	$p_w > p_i$			
Supraglacial lakes in cauldrons and sink holes ⁴		Downward to the base through englacial tunnels			
Englacial storage ⁵					
Linked subglacial cavities ⁶	Distributed drainage system switches to a tunnel system				

^a Superscript numbers refer to publications as follows:

- ¹ Bretz (1925, 1969); Thorarinsson (1939); Liestøl (1956); Stone (1963); Maag (1969); Post and Mayo (1971); Clague and Mathews (1973); Björnsson (1976); Lliboutry *et al.* (1977); Haeberli (1983); Sturm and Benson (1985); Yamada (1998).
- ² Thorarinsson (1957, 1958); Sturm *et al.* (1986); Björnsson (1988); Thouret (1990); Pierson *et al.* (1990); Trabant and Meyer (1992); Guðmundsson *et al.* (1997).
- ³ Björnsson (1974, 1988, 2002; Guðmundsson *et al.* (1995).
- ⁴ Björnsson (1976); Russell (1993).
- ⁵ Haeberli (1983); Driedger and Fountain (1989); Walder and Driedger (1995).
- ⁶ Kamb *et al.* (1985); Kamb (1987); Björnsson (1998).

There is neither a theory nor sufficient empirical data yet available for describing how jökulhlaups drain subglacially at a faster rate than can be explained by the expansion of conduits through melting. This quicker drainage involves the ice overburden being exceeded by water pressure, which lifts the glacier and allows space for the water. Initially propagated as a pressure wave, the water then moves forward in a sheet flow, prior to draining through conduits (Björnsson, 1997, 2002; Björnsson *et al.*, 2001; Jóhannesson, 2002).

3.2.2 Subaerial drainage

A subaerial burst following the sudden breaching of an ice dam is typically over sooner than a flood exiting through a basal ice tunnel. Discharge in the former outburst may increase linearly with time and, for a given lake volume, peak discharges are significantly higher than for drainage through subglacial tunnels. The size of the outlet and the quantity of the impounded water body determine the peak discharge through the breach.



Physical models of drainage over a subaerial breach have been drawn up along similar lines to those for drainage through subglacial tunnels. Walder and Costa (1996) successfully simulated hydrographs for Lake George in Alaska. The widening of the opening is assumed to be controlled by melting, while the lake discharge spilling over the breach is dictated by the hydraulic conditions of critical flow through an open channel.

Furthermore, an empirical power-law was derived in the form of eq. (4), with $K=1100/s\ m^{1.69}$ and $b=0.44$, i.e., similar figures to those for regression during known outbursts that breached man-made earthen dams, in which instances $K=1200/s\ m^{-1.56}$ and $b=0.48$ (Costa, 1988). Raymond and Nolan (2000) made the first attempt to identify and describe by a physical model the processes controlling unstable drainage over a spillway with an ice floor; in other words, they formulated criteria for when discharge water would melt the spillway path down faster than the lake level dropped. Their work successfully explained observations of supraglacial drainage from Black Rapids Glacier, Alaska.

4 Case studies

4.1 Floods from sediment-dammed lakes

Floods in which lakes breach sediment dams carry large amounts of debris. Generally unpredictable, they are regarded as the most dangerous kind of outbursts, but have been successfully prevented at many hazardous locations by artificial drains through the sediment dams, which restrict lake depths. Advances and subsequent retreats of glaciers in the Little Ice Age created many unstable moraine dams, with examples reported from the European Alps, the Himalaya and Peru. An Austrian valley was flooded in 1874 when the terminal moraine of the Madatschferner glacier gave way and the lake escaped from in front of the retreating ice. The French Tête Rousse débâcle of 1892, released $2 \cdot 10^5\ m^3$ of water and $8 \cdot 10^5\ m^3$ of sediment, killed 175 people (Liboutry, 1971). In 1926 a flood released from the Himalayan Shyok glacier devastated a village and cultivated land 400 km from the source (Mason, 1929), while the Peruvian Jancarurish outburst of 1950 discharged $2 \cdot 10^6\ m^3$ of water and $3 \cdot 10^6\ m^3$ of sediment (Liboutry *et al.*, 1977). In British Columbia, debris flows up to 20 m thick have travelled as far as 20 km downvalley after breaching terminal moraines (Evans and Clague, 1994). As still another example, an ice avalanche from the Langmoche glacier in Nepal broke a moraine dam in 1985, resulting in a flood that destroyed bridges, houses and a hydroelectric plant (Vuichard and Zimmermann, 1986, 1987)

4.2 Floods from marginal ice-dammed lakes

Jökulhlaups from ice-dammed lakes at glacial edges have posed a serious threat to roads, power lines and the human population in many countries of Europe, North and South America, Asia and New Zealand. In most countries, however, thinning of the ice dams by the warm climate of the twentieth century has led to smaller and more frequent outbursts from marginal lakes dammed by ice. Walder and Costa (1996) recently compiled a comprehensive review of outburst floods from glacially dammed lakes.

Some of the best-documented modern lakes blocked by ice are found in North America. Maag (1969) identified 125 lakes with ice dams on Axel Heiberg Island, Canada, and studies of several icedammed lakes in that country have been of major importance for the scientific understanding of jökulhlaups, for example, investigations of Tulsequah Lake, British Columbia (Marcus, 1960); Summit Lake, British Columbia (Mathews, 1965; Clarke, 1982; Mathews and Clague, 1993); Hazard Lake, Yukon Territory (Clarke, 1982); and Flood Lake, British Columbia (Clarke and Waldron, 1984).

Post and Mayo (1971) identified 750 lakes behind marginal ice dams in Alaska; of these, many have been thoroughly researched, such as Lake George (Stone, 1963; Hulsing, 1981; Lipscomb, 1989), Snow River (Chapman, 1981) and Strandline Lake (Sturm and Benson, 1985). The largest observed terrestrial flood from a breached impoundment took place on 8 October 1986, below the ice-dammed Lake Russell, Alaska. A surge had caused Hubbard Glacier, North America's largest tidewater glacier, to advance across the entrance of Russell Fjord, turning it into Lake Russell. A moraine was pushed up in front of the advancing glacier, raising inland water 25.5 m asl before the lake overflowed and cut the dam at the junction of the ice and the valley wall. The entire ice dam was removed within a few hours, with released water amounting to $5.4 \cdot 10^9\ m^3$ and the peak discharge reaching $1.5 \cdot 10^6\ m^3\ s^{-1}$ (Mayo, 1988, 1989). A similar event, though somewhat smaller, occurred in the summer of 2002.

In the Swiss Alps, damage from glacial floods occurs, on average, biennially; 60–70% of the outbursts originate in lakes at glacial margins and 30–40% in pockets of water within glaciers (Haeberli, 1983). Nowadays, about 15 ice-marginal lakes in Iceland drain by means of jökulhlaups (Thorarinsson, 1939; Björnsson, 1997) and frequently cause problems for bridges and other road structures. A pioneering thesis was written by Liestøl (1956) on glacially dammed lakes in Norway. Outburst floods have been scientifically described in Greenland (Schytt, 1956; Lister and Willie, 1957; Dawson, 1983; Russell, 1989), Argentina (Heinsheimer, 1954, 1958; Fernández *et al.*, 1991), Pakistan (Gunn *et al.*, 1930; Hewitt, 1982), and the former USSR (Glazyrin and Sokolov, 1976; Krenke and Kotlyakov, 1985; Konovalov, 1991).

The largest outbursts from ice-marginal lakes known through geological records occurred at the end of the last glaciation (Bretz, 1969; Baker, 1973, 1983; Teller and Clayton, 1983; Waitt, 1984; Clarke *et al.*, 1984; Bond *et al.*, 1992; Shaw, 1996, 2002). Jökulhlaups drained lakes that were dammed by margins of the Laurentide ice sheet in North America (Lake Agassiz and others) and by the Fennoscandinavian ice sheet (situated over the Baltic Sea). The Eurasian ice sheet had blocked off lakes, the outbursts of which emptied into northward-draining rivers in Siberia. Pulses of meltwater amassed in major proglacial lakes were released into the Mississippi and St



Lawrence Rivers and the Arctic and Hudson Straits, eventually to enter the North Atlantic Ocean. These floods emitted enormous amounts of freshwater, which may have affected ocean circulation and deep-water production in the North Atlantic (Teller, 1990). Giant jökulhlaups from the Laurentide ice sheet during the last glacial cycle may have played a part in depositing the Heinrich layers in the North Atlantic (Heinrich, 1988; Alley and MacAyeal, 1994; Colman, 2002). Floods from glacial Lake Missoula at the border of the Laurentide ice sheet repeatedly transported around 2×10^{12} m³ (2000 km³) of water in less than four days, with flow velocities of up to 30 m s⁻¹ and estimated peak discharges of 2×10^7 m³ s⁻¹ (Baker, 1973; Baker and Bunker, 1985; O'Connor and Baker, 1992), so that they comprise the greatest discharges of freshwater known in geological history.

The Lake Missoula floods lasted approximately one week, and the magnitude of their peak discharge indicates either supraglacial release or suddenly failing barriers, because such discharges cannot be explained by the gradual enlargement of basal ice tunnels. The floods from Lake Missoula produced the huge channelled scablands on the Columbia Plateau, Washington State.

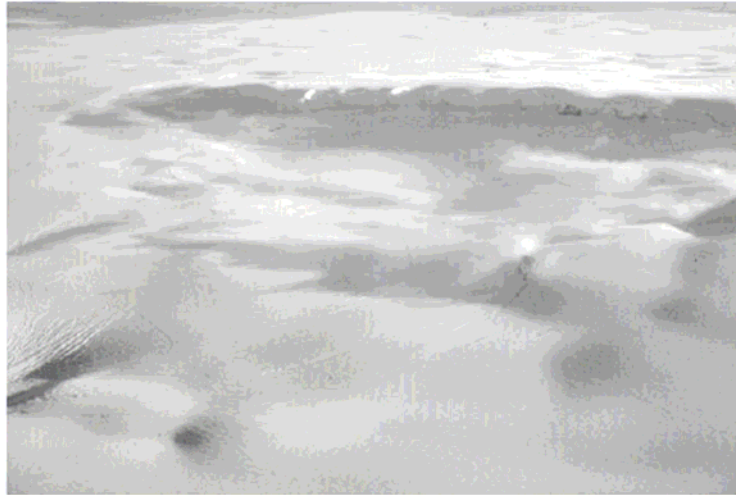


Figure 8.6 A 10-km-wide depression formed by continuous melting at the subglacial hydrothermal area of Grímsvötn, Vatnajökull, Iceland. View towards the 400-m-high caldera rim, Mt. Grímsfall. Photo: H. Björnsson

4.3 Floods from subglacial lakes

The best-documented floods from subglacial lakes occur in Iceland, where jökulhlaups regularly drain six lakes of this type. They are located in hydrothermal areas, where geothermal activity continuously melts ice on the glacier bed, creating a depression in the glacier surface, under which water accumulates until released in floods. The best-known jökulhlaups drain the lake Grímsvötn in Vatnajökull glacier (Björnsson, 1974, 1988; Thorarinsson, 1974) and occur at 1- to 10-year intervals. Their peak discharge ranges from 600 to 4×10^5 m³ s⁻¹ at the outwash plain, their duration is two days to four weeks and their total volume at each event is 0.5×10^9 m³. The greatest floods have peaked in less than one week and subsided in two days, whereas the smaller ones peak in two to three weeks, after which they usually terminate in approximately one week (Figures 8.4, 8.5 and 8.6). Jökulhlaups from Grímsvötn flow a distance of some 50 km beneath the ice to the glacier terminus by the Skeiðarársandur outwash plain (Figure 8.7). The most violent Grímsvötn jökulhlaups have flooded this entire plain, measuring 1000 km² (e.g., Björnsson, 1988). Typically, lake drainage begins at water pressures 6–7 bar lower than those exerted by the overburden at the ice dam. Conduits are enlarged in the course of several days or up to three weeks, so that the floods develop from small initial discharges which have been explained by classic jökulhlaup theories. Occasionally, nevertheless, the lake level rises until the ice dam floats; in this case, discharge increases faster than can be accommodated by the melting of conduits, and the glacier is lifted along the flow path as the water forces open space for itself. An outburst of this type took place in November 1996, when meltwater from the Gjalp eruption collected in Grímsvötn before flowing out in a disastrous flood: this is discussed below in connection with volcanism.

Other well-known jökulhlaups from subglacial Icelandic lakes originate at the Skaftá cauldrons (Figure 8.8), 10–15 km northwest of Grímsvötn, and result in floods of $50 \text{--} 350 \times 10^6$ m³. They rise over one to three days to a peak discharge of $200 \text{--} 1500$ m³ s⁻¹, before receding slowly for one to two weeks. So far, the hydrographs of these outbursts have not been simulated theoretically. Although their speedy rise suggests a water temperature at the reservoir well above the melting point (10–20°C), this must not be considered definite, because the theory of heat transfer is debatable.

4.4 Floods related to subglacial volcanic eruptions

In active volcanic areas of Iceland and of North and South America, eruptions frequently trigger catastrophic glacial floods (Thorarinsson, 1957, 1958; Sturm *et al.*, 1986; Björnsson, 1988; Trabant and Eyer, 1992; Pierson, 1995; Guðmundsson *et al.*, 1997; Thouret, this volume). Lava erupts directly into water stored under the ice and the boiling water transfers heat readily. Meltwater created by subglacial eruptions may discharge instantly toward



the glacier margins, depending on the rate of volcanic melting. The bursts may either travel subglacially in tunnels or spread out in a broad sheet, heaving up the ice. Moreover, the water may overtop the glacier, melt channels on its surface, and crack off giant blocks of ice. Alternatively, the meltwater forming beneath substantial ice masses may accumulate in subglacial lakes, eventually to exit from them in jökulhlaups. A recent example of such a flood occurred when meltwater from the 1996 Gjalp eruption gathered in the subglacial lake Grímsvötn, entering at a rate of up to $5 \cdot 10^3 \text{ m}^3 \text{ s}^{-1}$, before draining from the lake in a catastrophic flood, with discharge increasing almost linearly to a peak of $4 \cdot 10^4 \text{ m}^3 \text{ s}^{-1}$ in 16 hours. Within a period of 40 hours, $3.2 \cdot 10^9 \text{ m}^3$ of water had left the lake. Even though the outburst had begun as a flood wave, so that the initial discharge was of the order of $5 \cdot 10^3$ to $5 \cdot 10^4 \text{ m}^3 \text{ s}^{-1}$, an exponential increase would have taken around two days to reach the estimated peak.



Figure 8.8 A 3-km-wide and 150-m-deep cauldron after a jökulhlaup in Vatnajökull, Iceland.
Photo: H. Björnsson

Eruptions of snow-clad volcanoes may generate huge floods when pyroclastic, turbulent fluid melts the icy cover, producing slushy, debris-rich flows which speed down valley slopes. Thouret (this volume) describes how a relatively small eruption from the snow-capped volcano Nevado del Ruiz (5400 m high) in Colombia in 1985 caused disastrous lahars that killed more than 23 000 inhabitants of the town Armero, 72 km away. In 1966–68 an eruption at Mt Redoubt, Alaska, melted or washed away $6 \cdot 10^7 \text{ m}^3$ of glacial ice (Sturm *et al.*, 1986) and in 1989–90 another eruption achieved double that amount (Trabant and Meyer, 1992).

A contemporary description of the outburst during the 1362 Öræfajökull eruption in Iceland provides an example of abrupt, destructive jökulhlaups from an ice-capped stratovolcano (Thorarinsson 1957, 1958). In less than one day, the flood may have reached its peak of at least $10^5 \text{ m}^3/\text{s}$ (Thorarinsson, 1958). Water originating in the 2000 m high vicinity of the summit streamed subglacially down the slopes before bursting out to wash away several farmsteads and to leave dead ice, sediment and hummocks covering the lowlands at a depth of 2–4 m. Outwash plains appeared where sea depths had previously measured 50 m. A contemporary record (report by Rev. Jón Thorláksson, cited by Henderson (1818); Thorarinsson, 1958: p. 31) describes how “several floods of water 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 [sic] 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”.

A less devastating eruption took place in AD 1727 (Thorarinsson, 1958). The swiftest, greatest outbursts of glacial meltwater noted in world history accompany volcanic eruptions in the Katla caldera, which rests under the Icelandic ice cap Myrdalsjökull. Occurring on average twice per century, the outbursts have durations of 3–5 days, peak discharges estimated at 10^5 – $10^6 \text{ m}^3/\text{s}$, and total volumes of 1–8 $\cdot 10^9 \text{ m}^3$ (Thorarinsson, 1957, 1975; Maizels, 1989, 1995; Tómasson, 1996; Larsen, 2000). Immense blocks of ice break off the glacial margins, and a mixture of water, ice, volcanic emissions and sediment surges over the outwash plain at velocities of 5–15 m/s. During a segment of the flow event, the water may consist of a hyperconcentrated fluid–sediment mixture. The amount of volcanic debris produced per event and carried away by the water has been estimated to range from 0.7 to 1.6 $\cdot 10^9 \text{ m}^3$, or be of the order of 2 $\cdot 10^9 \text{ t}$ (Tómasson, 1996; Larsen, 2000). During the last eruption in 1918, the neighbouring coastline advanced 3 km into the sea. Although many loose deposits have been washed away in succeeding years, the affected seashore remains 2.2–2.5 km further south than its position in 1660.



Jökulhlaups have played a dominant role in building up the outwash plain, which like other such plains in Iceland is called a *sandur* – a term that has acquired international use. Marine sediments containing debris transported from the eruption site are found in the ocean several hundred kilometres to the south. Jökulhlaups from Katla have threatened local communities, damaged vegetation, ruptured roads on the alluvial flats surrounding the ice cap and even generated flood waves in coastal waters. On the northern side of Myrdalsjökull, the outbursts have eroded deep canyons in the bedrock.

The largest, most catastrophic jökulhlaups in Iceland, with peak discharges of the order of $10^6 \text{ m}^3/\text{s}$ were caused by prehistoric eruptions in the voluminous, ice-filled calderas of Bárðarbunga and Kverfjöll in the northern reaches of Vatnajökull (see Tómasson, 1973; Björnsson, 1988; Björnsson and Einarsson, 1991; Knudsen and Russell, 2002; Waitt, 2002). Floods sweeping down Jökulsá á Fjöllum carved a conspicuous scabland and a deep canyon Jökulsárgljúfur; Figure 8.9); erosion by cavitation is considered to have been a significant factor in their creation.



Figure 8.9 A canyon eroded and carved by jökulhlaups (the 300-m-deep Jökulsárgljúfur, north of Vatnajökull, Iceland)
Photo: Björn Rúriksson, reproduced with permission.

5 Concluding remarks and future prospects

This chapter has reviewed the current knowledge of glacial floods and their sources in impounded glacial meltwater, and on the dams holding back the water and dam failure. Also, I have described the characteristics of flood discharges, their impact in the proglacial zone, and their significance for society. Despite some outburst sources lying hidden, field inspection or remote sensing can detect most of them. Potential jökulhlaup hazards can be recognized by examining the properties of reservoir dams and evaluating their stability. While the onset of outbursts cannot be timed exactly, those stemming from ice-dammed lakes may be predicted empirically – though no more precisely than to the nearest month, based on records of lake levels at the beginning of previous outbursts. The discharge rates of change and impacts of jökulhlaups depend on their release mechanisms, i.e., whether a sediment dam suddenly fails, an ice dam is abruptly lifted or overtopped, or the dam seal begins to leak through a subglacial tunnel at gradually increasing rates. The nature of the flood hazard depends on reservoir volume and the temperature of the fluid, along with the character and content of the sediment that is entrained in the flow. Empirical and theoretical relations help to forecast the possible magnitude of jökulhlaups. Theoretical models of flood discharge may help to predict the slopes of the ascending limbs of the hydrograph if the drainage occurs through tunnels under an ice dam. However, formulating such models requires information on water temperature and bottom roughness, and these data are rarely available. At present, theoretical models are unreliable for predicting peak discharge, although empirical relationships may provide some estimate, given the total volume of the reservoir at the beginning of the outburst. Once the flood has started, the subsequent discharge rate may be predicted on the grounds of reservoir volume and of flow measured during the initial phase of the flood. No theory is yet available for subglacially draining jökulhlaups, which increase faster than can be explained by conduits widening through melting. Glacier lakes are ephemeral features, dependent on variable conditions such as the advance or recession of glacial termini. Continuous monitoring of impounded water and inflow rates is needed for assessing risks and delineating danger zones. Outbursts from lakes behind sediment dams at glacial margins can be prevented by installing drainage pipes to control water levels. Some englacial sources of floods may be difficult to detect, and subglacial volcanic eruptions are hard to predict, but maps of bedrock topography and the glacier surface can be used to identify potential hazard zones with regard to jökulhlaups. Field studies of landforms, sediments and other evidence of varying flow processes would improve



understanding of the outburst mechanism. At present, both empirical and theoretical calculations suffer from inaccurate estimates of discharges and particularly of peak flows. Further studies are required on the physics of how jökulhlaups begin to enable more precise timing of their onset. Additionally, studies of supraglacial outbursts that melt gorges through ice dams and of subglacial bursts that lift the glacier to propagate a flood wave are still at an early stage.

Studies of the physics of modern glacial floods of every magnitude may fundamentally improve our understanding of palaeofloods in the late Pleistocene and early Holocene. They released enormous amounts of freshwater into the ocean, perhaps altering deep-water circulation and the global climate. In this respect, the glacial floods of today may become a key to the past.

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