JÖKULHLAUPS: A REASSESSMENT OF FLOODWATER FLOW THROUGH GLACIERS

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[1] In glaciated catchments, glacier-generated floods (jökulhlaups) put human activity at risk with large, sporadic jökulhlaups accounting for most flood-related fatalities and damage to infrastructure. In studies of jökulhlaup hydrodynamics the view predominates that floodwater travels within a distinct conduit eroded into the underside of a glacier. However, some jökulhlaups produce subglacial responses wholly inconsistent with the conventional theory of drainage. By focusing on Icelandic jökulhlaups this article reassesses how floodwater flows through glaciers. It is argued that two physically separable classes of jökulhlaup exist and that not all jökulhlaups are an upward extrapolation of processes inherent in events of lesser magnitude and smaller scale. The hydraulic coupling of multiple, nonlinear components to the flood circuit of a glacier can induce extreme responses, including pressure impulses in subglacial drainage. Representing such complexity in mathematical form should be the basis for upcoming research, as future modeling results may help to determine the glaciological processes behind Heinrich events. Moreover, such an approach would lead to more accurate, predictive models of jökulhlaup timing and intensity.

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1. INTRODUCTION

1.1. Background and Research Perspective

[2] Jökulhlaup is an Icelandic term derived from the words jökull (glacier) and hlaup (literally meaning sprint or burst). In glaciological parlance a jökulhlaup is a sudden release of meltwater from a glacier or a moraine-dammed lake, culminating in a significant increase in meltwater discharge over a period of minutes to several weeks [e.g., *Sturm and Benson*, 1985; *Paterson*, 1994] (Table 1). Contemporary understanding of jökulhlaup physics has evolved from inspiring field-based studies in Iceland, Norway, and Canada [*Thórarinsson*, 1939; *Liestøl*, 1956; *Mathews*, 1973]. Estimates of maximum discharge range from 10^0–30×10^4 m^3 s^{-1} for historical jökulhlaups [*Mayo*, 1989; *Tómasson*, 1996; *Klingbjer*, 2004] to $\sim 10^5$–20×10^6 m^3 s^{-1} for late Pleistocene events [*Baker*, 2002]. The irregular timing and magnitude of some jökulhlaups occasionally causes fatalities and widespread damage to infrastructure within the flood tract [*Richardson*, 1968; *Young*, 1980; *Rist*, 1983; *Thouret*, 1990; *Clague and Evans*, 2000; *Haeberli et al.*, 2001; *Björnsson*, 2002, 2004].

[3] Jökulhlaups can originate from meltwater stored in ice-marginal, *subglacial*, *englacial*, and *supraglacial* locations (Figure 1 and Table 1), although volcanogenic and rainfall-induced floods can occur without significant water storage. (Italicized terms are defined in the glossary, after the main text.) Meltwater reservoirs form by a combination of favorable hydraulic pressure gradients, local topography and, sometimes, geothermal or hydrothermal heat [*Björnsson*, 1974; *Nye*, 1976]. Some reservoirs are thought to have persisted for millennia [*Dowdeswell and Siegert*, 1999, 2002], others are thought to have persisted for several hundred years [*Teller et al.*, 2002], and some are thought to have existed for just hours, reflecting transient hydrologic conditions [*Roberts et al.*, 2003].

[4] Physical understanding of jökulhlaup dynamics is based on empirical data from a comparatively small population of well-studied events, where it is assumed that a single, straight subglacial conduit links a pressure-coupled reservoir of meltwater directly to the *glacier terminus* [*Glazyrin and Sokolov*, 1976; *Nye*, 1976; *Spring and Hutter*, 1981; *Clarke*, 1982; *Fowler*, 1999]. Jökulhlaups exhibiting an exponential rise to maximum discharge are a product of a runaway increase in the flow capacity of a water-filled, subglacial conduit. Such a response is perpetuated by frictional melting of the conduit walls due to heat convected by the latent heat of fusion [*Nye*, 1976] (Figure 2). The amount of heat produced is proportional to the water flux. If meltwater temperature is above the *pressure-determined melting point*, then the melt-widening effects of heat conduction prevail (Figure 2), forcing discharge to increase at a greater-than-exponential rate [*Spring and Hutter*, 1981].

[5] However, empirical and theoretical evidence demonstrates that some jökulhlaups peak too suddenly to be explained by invoking conventional views of conduit melt widening along the entire flood path [*Thórarinsson*, 1957; *Haeberli*, 1983; *Walder and Driedger*, 1995; *Björnsson*,

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TABLE 1. Recognized Types of Jökulhlaup and Candidate Trigger Mechanisms[a]

Recognized Types of Jökulhlaup	Postulated Trigger Mechanisms
Type 1, drainage of an ice-marginal, ice-dammed lake	ice dam flotation due to hydrostatic stress [*Thórarinsson*, 1939] Glen mechanism: viscoplastic deformation of the ice dam when hydrostatic stress exceeds glaciostatic stress and the confining strength of ice [*Glen*, 1954][b] supraglacial overspill (common at cold-based glaciers), resulting in thermodynamic erosion of an ice-lined spillway hydraulic tapping of a meltwater reservoir by intraglacial drainage [*Fisher*, 1973; *Anderson et al.*, 2003] Darcian flow at the base of the ice dam [*Fowler and Ng*, 1996; *Fowler*, 1999] enhanced glacier sliding close to the ice dam, leading to mechanical rupture of the dam base [*Liestøl*, 1956] hydrodynamic enlargement of a breach formed between substrate and glacier ice [*Walder and Costa*, 1996]
Type 2, drainage of a supraglacial lake	hydraulic tapping of a meltwater reservoir by intraglacial drainage (see type 1 for references) nonlinear rate of spillway lowering due to viscous dissipation and conduction of heat from flowing meltwater [*Raymond and Nolan*, 2000] release of meltwater into intraglacial drainage due to descent of a hydrofracture from the base of a supraglacial lake [*Björnsson*, 1976; *Boon and Sharp*, 2003]
Type 3, volcanically induced jökulhlaup	profuse ice melt due to subglacial volcanism, causing either subglacial ponding (see type 4 for trigger mechanisms) or immediate drainage of floodwater with no significant storage [*Björnsson*, 1988; *Pierson*, 1989]
Type 4, drainage of a subglacial lake	ice dam flotation due to hydrostatic stress (see type 1 for reference) Glen mechanism (see type 1 for explanation and reference) Darcian flow at the base of the ice dam (see type 1 for references)
Type 5, drainage of an intraglacial cavity	englacial rupture of a water-filled vault [*Haeberli*, 1983] hydraulic tapping of a meltwater reservoir by intraglacial drainage (see type 1 for references) enhanced glacier sliding close to the ice dam, leading to mechanical rupture of the dam base (see type 1 for reference) sudden input of meltwater to the glacier bed via crevasses and moulins [*Walder and Driedger*, 1995]
Type 6, drainage of a moraine-dammed lake, including those dammed by ice cored moraines	sudden and significant increase in lake level or a rapid displacement of lake mass, leading to heightened rates of spillway incision (Rapid increases in lake height are attributable to high influxes of meltwater to the lake, caused by rainfall in the ablation zone, pronounced ice surface melting, or the subglacial drainage of an ice-dammed lake within the same water catchment [*Haeberli*, 1983; *Clague and Evans*, 2000].) trigger mechanisms for lake displacement including direct impacts by snow and ice avalanches, landslides, or rockfalls (Other trigger mechanisms include unstable Darcian flow through porous sediments and lake seepage through fractured ice cored moraines [*Haeberli et al.*, 2001].)
Type 7, meltwater release during surge termination	disruption of distributed subglacial drainage by the reestablishment of canalized subglacial drainage, resulting in the sudden evacuation of stored meltwater [*Kamb et al.*, 1985; *Björnsson*, 1988]

[a]Note that the numbered type entries correspond to the reservoir sites and meltwater sources depicted in Figure 1.
[b]In the Glen mechanism, putative viscous pushing of glacier ice is unrealistic; if glaciostatic flotation occurs, a hydraulic breach will likely develop by lateral hydrofracturing at the glacier base [see *Fowler*, 1999].

1992, 2002; *Jóhannesson*, 2002; *Flowers et al.*, 2004]. Insight gleaned primarily from (1) lake drawdown and flood transit times [*Björnsson*, 1998]; (2) comparative measurements of meltwater temperature at the flood source and glacier terminus [*Rist*, 1955; *Björnsson*, 1992]; (3) borehole pressure transducers [*Anderson et al.*, 2003]; (4) seismic surveys [*Nolan and Echelmeyer*, 1999]; (5) ice-dam kinematics [*Walder et al.*, 2005]; (6) *proglacial* hydrograph reconstructions [*Björnsson*, 1977]; (7) volumetric estimates of subglacial sediment removal [*Fowler and Ng*, 1996]; (8) hydrochemical observations [*Anderson et al.*, 2003]; and (9) sudden changes in *outlet* location [*Roberts et al.*, 2000], suggests that jökulhlaups permeate large zones of the *glacier bed* before reaching the glacier terminus.

1.2. Purpose and Scope of This Review

[6] The aims of this review are (1) to reassess how floodwater flows through glaciers; (2) to develop a qualitative view of subglacial processes during jökulhlaups; and (3) to suggest research priorities for future jökulhlaup studies. I use mostly examples from Iceland to illustrate geophysical processes, and I extend the scope of previous jökulhlaup reviews [*Björnsson*, 1988, 1992; *Maizels and Russell*, 1992; *Walder and Fountain*, 1997; *Tweed and Russell*, 1999; *Snorrason et al.*, 2000; *Björnsson*, 2002, 2004] by focusing on interactions between various components of a glacier and its drainage system under extreme hydraulic transience. Because this article emphasizes subglacial processes, it makes only incidental reference to important topics such as drainage of moraine-dammed lakes, jökulhlaup hazard assessment, engineering mitigation strategies, magnitude-frequency regimes, and geomorphic impact. These topics are considered elsewhere more thoroughly [see *Rist*, 1983; *Major and Newhall*, 1989; *Walder and Costa*, 1996; *Clague and Evans*, 2000; *Huggel et al.*, 2002; *Kattelmann*, 2003].

2. THEORETICAL MODELS, ASSUMPTIONS, AND THE EMERGENCE OF EXTRAORDINARY JÖKULHLAUPS

[7] The work of *Nye* [1976] represents one of the most pioneering, complete, and heuristic studies of jökulhlaup physics. Nye's model of jökulhlaup hydrodynamics built on earlier empirical studies by *Liestøl* [1956], *Mathews* [1973], and *Björnsson* [1974, 1975] and theoretical treatments of steady state water flow by *Röthlisberger* [1972], *Shreve* [1972], and *Weertman* [1972]. Through analysis of differential equations for nonsteady water flow at the start of the subglacial flood path, Nye demonstrated that an exponential rate of discharge increase could be explained by thermal melting of a water-filled, ice-encased conduit due to the frictional effect of turbulent water flow. In resolving differ-

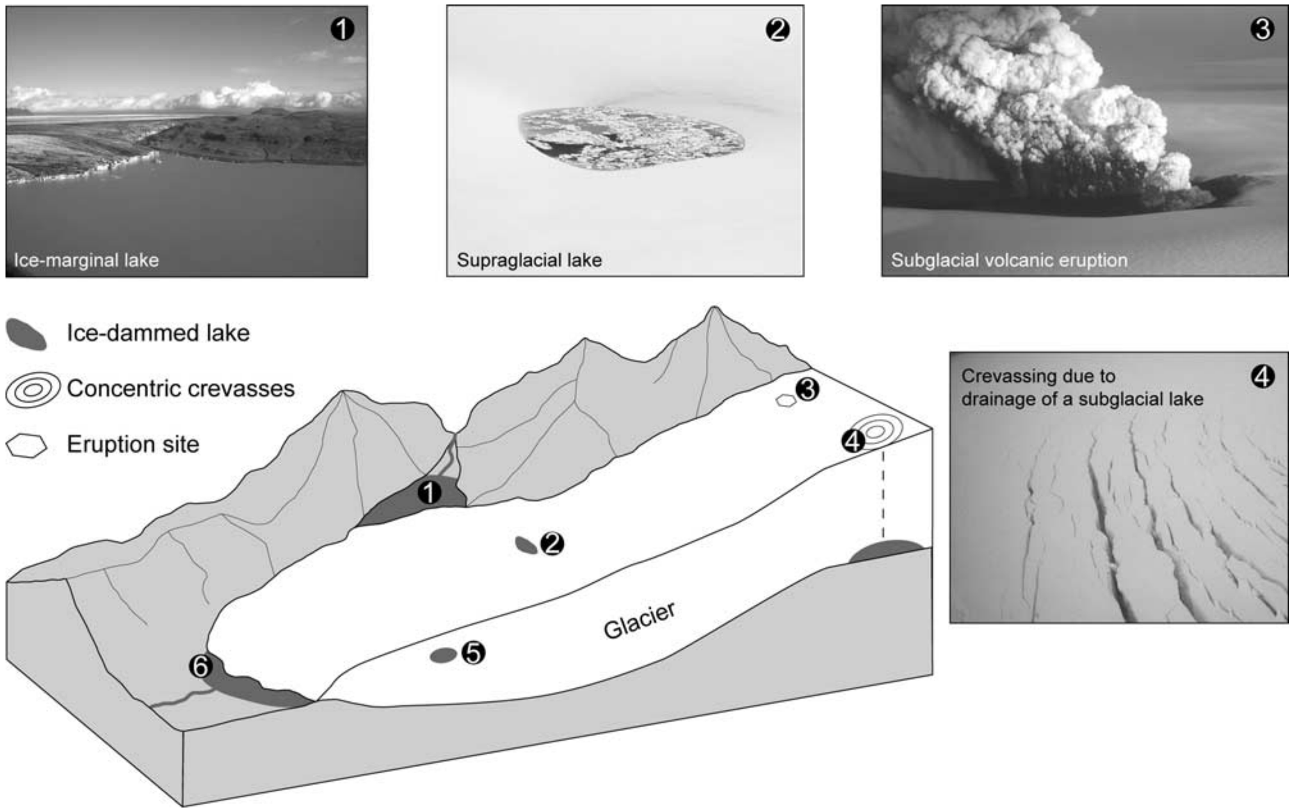

Figure 1. Reservoir sites and meltwater sources for jökulhlaups. See Table 1 for a summary about each numbered location. Photographer M. J. Roberts.

ential equations for jökulhlaup dynamics at Skeiðarárjökull, Iceland, *Nye* [1976, pp. 187–190] assumed a cylindrical, preexisting subglacial conduit of constant cross-sectional diameter and hydraulic gradient as the means for conveying floodwater directly to the glacier margin. Further, albeit necessary, simplifications included a reservoir isotherm of 0°C and a spatially uniform, antecedent base flow down the flood conduit. Although later studies have advanced Nye's original work substantially by improving some of the assumptions regarding reservoir temperature and hypsometry [*Spring and Hutter*, 1981; *Clarke*, 1982; *Björnsson*, 1988, 1992; *Elvehøy et al.*, 2002], heat transfer and conduit melt rate [*Spring and Hutter*, 1982; *Jóhannesson*, 2002; *Clarke*, 2003], and dynamic conduit geometry and motion [*Spring and Hutter*, 1982; *Clarke et al.*, 1984; *Fowler and Ng*, 1996], the main principles of his theory endure.

2.1. Anatomy of an Exponentially Rising Jökulhlaup

[8] On the basis of *Nye*'s [1976] model, its later corroboration [*Spring and Hutter*, 1981, 1982], and subsequent critical testing and revision [*Clarke*, 1982] I describe the rudimentary physics of a conventional jökulhlaup. In its simplest form the revised model explains exponentially rising discharge from a subglacial ice-dammed lake by melt widening of a single, preexisting conduit, with necessary heat being provided by dissipation of potential energy [*Liestøl*, 1956] and advection of thermal energy from the meltwater reservoir [*Clarke*, 1982]. Visualize a meltwater lake, encased by a finite lining of temperate glacier ice, located beneath a depression in a glacier surface (Figures 1 and 3); viewed in profile, the morphology of the lake resembles an inverted bowl (cupola). A bedrock-ice inter-

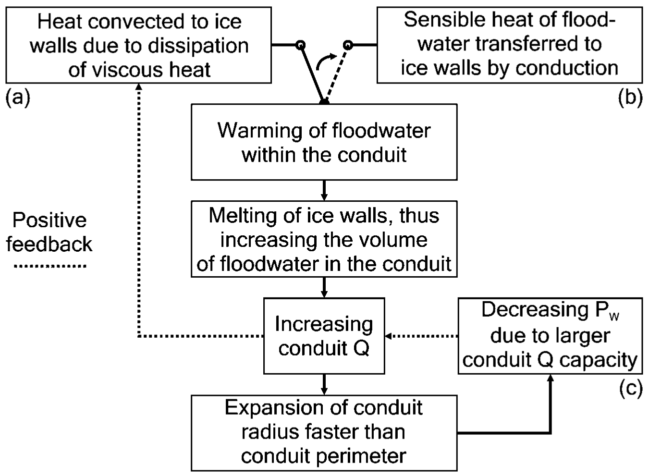

Figure 2. Subglacial conceptualization of jökulhlaup thermodynamics, following the theoretical tenet of *Spring and Hutter* [1981]. (Note that Q and P_w signify discharge and hydrostatic pressure, respectively.) When the effect of process a prevails, Q rises exponentially because of positive, thermodynamic feedback. Conversely, when process b dominates, Q rises more swiftly than would be possible under the forcing of process a, and the effect of consequence c diminishes temporarily.

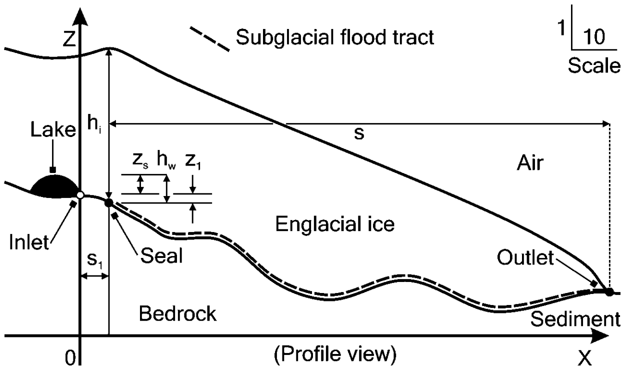

Figure 3. Schematic view of an idealized subglacial lake and contiguous flood tract. Note the location of the conduit seal at the point of maximum ice dam thickness. See section 2.1. for an explanation of the notations used.

face exists at the base of the lake, and meltwater and ice are driven toward the lake center [*Björnsson*, 1975, 1988]; hence for the most part a local hydraulic gradient exists between glacier and lake. Increasing water flux, coupled with fluid-ice transfer of sensible heat (enthalpy decrease), forces lake volume to increase. Under idealized conditions, where (1) the effective ice dam is glaciostatic and structurally flawless, (2) a potential gradient exists between the lake and proximal subglacial drainage over a comparatively narrow zone [*Björnsson*, 1988], and (3) a tight seal is maintained, lake drainage is practicable when

$$P_w = \rho_w g h_w \propto P_i = \rho_i g h_i, \qquad (1)$$

where P_w and P_i are hydrostatic and glaciostatic pressure, respectively; ρ_w and ρ_i are respective freshwater and pure ice densities (998 kg m^{-3} (at 4°C) and $\sim$900 kg m^{-3}); g is gravitational acceleration ($\sim$9.81 m s^{-2}); h_w is lake surface height above the seal (m); and h_i is ice thickness above the location of the seal (m) (Figure 3). Therefore water pressure at the seal is approximated by the hydrostatic pressure of the lake [*Clarke*, 1982]

$$P_w(s_1, t) = \rho_w g h_w(t), \qquad (2)$$

where hydrostatic pressure is a function of time (t) with respect to downstream distance from the conduit inlet to the lake seal (s_1), and $h_w(t) = z_s(t) - z_1$, where z_s is the lake surface height above the conduit inlet (not to be confused with the seal) and z_1 is the height of the conduit inlet above the seal (Figure 3).

[9] Drainage is initiated when hydrostatic pressure at the seal is equal to minimum hydraulic potential in the region damming the lake [e.g., *Sturm and Benson*, 1985]. However, it is stressed that many subglacial and ice-marginal lakes drain long before conditions for ice dam flotation are met [*Mathews*, 1973; *Björnsson*, 1974, 1976, 1988, 1992; *Clarke*, 1982; *Fowler*, 1999; *Anderson et al.*, 2003; *Roberts et al.*, 2005], thus suggesting the feasibility of processes such as (1) Darcian flow beneath an ice dam [*Gilbert*, 1971; *Fisher*, 1973; *Fowler and Ng*, 1996]; (2) mechanical breaching of the effective dam area to cause hydraulic short circuiting [*Nye*, 1976; *Fowler*, 1999]; (3) *Nye*'s [1976] notion of a buoyant, inverted cantilever acting on the lake seal because of a temporary lack of glaciostatic balance during lake drawdown; and (4) strong feedbacks between low-pressure subglacial drainage and a meltwater reservoir [*Kessler and Anderson*, 2004]. As meltwater flows under the buoyed region of the ice dam, thermal and frictional energy dissipation cause flow to localize into a discrete channel, often manifest at the glacier surface by an elongated collapse zone delineating the start of the conduit inlet [e.g., *Sturm and Benson*, 1985; *Björnsson et al.*, 2001]. Because of lake surface lowering, hydraulic head at the conduit inlet decreases as drainage progresses, thus inhibiting sustained sheet flow ($P_w \propto P_i$) down glacier from the seal [*Shoemaker*, 1992a, 2002]. Therefore the hydraulic potential

$$\phi = p_w g z + p_i \qquad (3)$$

driving meltwater down the conduit decreases with time, where z is the elevation above some datum; similarly, the hydraulic pressure opposing plastic creep or fracture closure of the conduit ($\partial S/\partial t$, where S is conduit cross-sectional area (m^2)) decreases proportionately. Jökulhlaup discharge (Q) at the seal of the subglacial lake can be derived from the water balance equation

$$\frac{\partial V}{\partial t} = Q_{\mathrm{IN}} - Q_{\mathrm{OUT}} - Q, \qquad (4)$$

where V is lake volume, Q_{IN} is the antecedent rate of lake inflow, and Q_{OUT} is the corresponding rate of nonflood water escape, either because of lake seepage, evaporation, or a spillway controlling lake surface elevation [*Clarke*, 1982]. Assuming the start of a jökulhlaup coincides with a full reservoir, the difference between Q_{IN} and Q_{OUT} can usually be neglected.

[10] *Mathews* [1973], *Nye* [1976], and *Clarke* [1982] utilized the Gauckler-Manning Q equation, derived originally for engineering purposes [*Chow*, 1959], to approximate instantaneous flow rate in a circular *intraglacial* conduit of uniform down glacier cross section:

$$Q = \frac{S^{4/3}\langle -\partial\phi/\partial s\rangle^{1/2}}{(4\pi)^{1/3}(\rho_w g)^{1/2} n}. \qquad (5)$$

In equation (5), n is Manning's roughness coefficient [*Chow*, 1959] for the conduit walls, and $-\partial\phi/\partial s$ expresses the hydraulic gradient (when $-\partial\phi/\partial s < 0$), where s denotes the inferred conduit length from inlet to outlet (Figure 3). The utility of the perimeter-averaged Gauckler-Manning equation for plausible estimations on subglacial Q is debatable [*Clarke*, 2003]. The equation was conceived for a steady state efflux at atmospheric pressure within a geometrically regular channel [*Chow*, 1959]; but the subglacial reality is pressurized, turbulent flow along multiple, irregularly shaped routeways formed primarily from the solid phase of the fluid flowing through it.

[11] For glaciers overlying coarse, unconsolidated sediments, initial flow resistance along the flood path would be comparable to shallow flow through a wide, boulder-strewn channel ($n \sim 0.1$ m$^{-1/3}$ s). Empirically simulated hydrographs require an n value between 0.08 and 0.12 m$^{-1/3}$ s (compare value of 0.03 m$^{-1/3}$ s derived by *Fowler and Ng* [1996]) to obtain a satisfactory fit with measured hydrographs [*Nye*, 1976; *Clarke*, 1986; *Björnsson*, 1992]. Assuming conduit shape is preserved with increasing Q [*Nye*, 1976], then n decreases with increasing S at the rate $S^{1/12}$ [*Spring and Hutter*, 1982]. It is known, however, that subglacial flood paths through unconsolidated sediments widen laterally to engender broad, shallow flows (i.e., a small Q/S ratio) [e.g., *Shreve*, 1985; *Fowler and Ng*, 1996]. Under these circumstances the effects of relative conduit roughness (n') dominate, and n, which applies strictly to conduits of geometrically similar cross sections (equation (5)), fails to apply; consequently, n' changes with size while n remains constant [*Clarke*, 2003]. Little is known about spatial and temporal variations in flood path roughness beneath and within a glacier, and given the need to parameterize mean conduit roughness in Q formulae, further evaluation of the applicability of experimentally determined frictional laws and the potential for spatially inhomogeneous values of n' is required. Moreover, the significance of roughness imparted dynamically by sediment-laden floodwater has received virtually no assessment in glaciological literature (see sections 3.3 and 4.5).

[12] At the onset of a jökulhlaup the immediate existence of an incipient conduit providing direct hydraulic coupling from the conduit inlet to the outlet is doubtful for most situations [*Shoemaker*, 2002; *Anderson et al.*, 2003]. Nonetheless, semipermanent conduits are known to exist for some meltwater reservoirs overlying geothermal fields [e.g., *Björnsson*, 1988]. For the majority of reservoirs a more realistic situation is efflux from the conduit inlet dissipating into preexisting intraglacial drainage at a finite distance down glacier from the reservoir seal. Jökulhlaups are effective at purging meltwater from hydraulically isolated zones of subglacial drainage; consequently, antecedent *glaciohydraulic* conditions can modulate the intraglacial transit time and routing of minor jökulhlaups [e.g., *Merrand and Hallet*, 1996; *Anderson et al.*, 2003].

[13] Although I have considered physical processes during drainage of a subglacial lake, meltwater reservoirs can also form (1) from multiple, water-filled subglacial cavities [*Kamb et al.*, 1985; *Warburton and Fenn*, 1994; *Walder and Driedger*, 1995]; (2) where glacier ice overlies a geothermal field or hydrothermal system [*Björnsson*, 1988; *Pierson*, 1989]; (3) on the glacier surface, when meltwater accumulation in depressions exceeds ice permeability [*Björnsson*, 1976; *Boon and Sharp*, 2003]; and (4) in ice-marginal locations where suitable glaciohydraulic and topographic conditions allow meltwater to accumulate against the flank of a glacier [*Björnsson*, 1976; *Walder and Costa*, 1996]. Large ice-marginal lakes such as Grænalón ($\sim 5 \times 10^8$ m^3, Skeiðarárjökull) and Hidden Creek Lake ($\sim 3 \times 10^7$ m^3, Kennicott Glacier, Alaska, United States) exist because P_i at the base of the ice dam is high enough to force some intraglacial meltwater to flow toward the lake. Only meltwater reservoirs created in subglacial or ice-marginal locations have the ability to initiate a jökulhlaup by localized glacier flotation [*Nye*, 1976]. Supraglacial lakes drain in response to hydromechanical opening of drainage pathways [*Boon and Sharp*, 2003], changing subglacial water pressure [*Sturm and Benson*, 1985], or thermal erosion of ice spillways [*Raymond and Nolan*, 2000]. Once subglacial flow channelization occurs, regardless of whether floodwater enters a glacier from its base or surface, the same glaciohydraulic models described earlier are tenable.

2.2. Volume of Floodwater Drained

[14] Ignoring base flow, total flood volume is a product of initial reservoir volume (or hydrothermal flux) and the addition of water derived from the thermal enlargement of conduit walls, assuming isothermal ice at the pressure-determined melting point. Variations in floodwater temperature, flood duration, hydraulic potential, and the length of conduit drainage consequently affect the volumetric contribution of frictional ice melt [see *Liestøl*, 1956]. Additionally, release of antecedent meltwater from intraglacial storage, caused by the subglacial passage of floodwater, can augment flood volume [*Anderson et al.*, 2003].

2.3. Conduit Growth and Perpetuation of an Exponential Rise in Discharge

[15] By comparing a prescribed rate of conduit melt widening against the effect of conduit closure due to gravitational ice creep, *Nye* [1976] elucidated the physics of conduit growth using the relation

$$\frac{\partial S}{\partial t} = \frac{M}{\rho_i} - K_0 S N^{n^*}. \tag{6}$$

In equation (6), M expresses volumetric ice melt per unit distance per unit time (m^3 m^{-1} s^{-1}, see equation (18)), K_0 is a constant derived from the flow law of ice [*Paterson*, 1994], N signifies *effective pressure*, and n^* is a material exponent (≈ 3) from Glen's flow law [*Paterson*, 1994]. *Nye* [1976] resolved the rise of the 1972 jökulhlaup from Skeiðarárjökull to a simple differential equation

$$\frac{\partial Q}{\partial t} = K Q^{5/4}, \tag{7}$$

which he evaluated against a proglacial rating curve for the same flood by *Rist* [1973]; Nye's model mimicked Rist's field data perfectly. Later physical analyses of the 1972 jökulhlaup achieved $Q(t)$ and Q/S approximations similar to that of Nye's [see *Spring and Hutter*, 1981, 1982; *Björnsson*, 1992; *Fowler and Ng*, 1996]. Equation (7) has the solution

$$Q = \left(-\frac{4}{Kt}\right)^4, \tag{8}$$

where K ($\mathrm{m}^{-3/4}\ \mathrm{s}^{-3/4}$) is a coefficient expressed as

$$K = \frac{(4/3)(\rho_w/\rho_i)g}{(S/\pi R^2)^{1/4} L n^{3/4}} (\Delta s/\Delta z_w)^{11/8}. \quad (9)$$

In equation (9), $S/\pi R^2$ represents the hydraulic radius of a circular, water-filled conduit (where R denotes conduit radius), L is the latent heat of fusion for glacier ice ($\sim$333.5 kJ kg^{-1}), and z_w is the vertical elevation difference between the lake surface and the primary outlet. Assuming a finite time remaining (τ) before rising stage Q becomes infinitely high, it follows from equation (8) that $Q \propto \tau^{-4}$ [*Nye*, 1976].

[16] Using field data from the 1978 jökulhlaup from Hazard Lake, Canada, *Clarke* [1982] simplified *Nye*'s [1976] jökulhlaup model to a concise mathematical description. This dimensionless model characterized reservoir hypsometry and the relative magnitudes of conduit enlargement by transfer of frictional and sensible heat and conduit closure by ice creep motion due to gravity. Three dimensionless parameters were invoked to define the relative influence of reservoir hypsometry (*M*), lake temperature (β), and conduit closure (α) on jökulhlaup dynamics. *Clarke*'s [1982] model has the greatest utilitarian value because it can be applied to a variety of jökulhlaups [e.g., *Clarke et al.*, 1984; *Clarke*, 1986; *Walder and Costa*, 1996; *Raymond and Nolan*, 2000; *Elvehøy et al.*, 2002], although the dimensionless lake parameter (*M*) is applicable only to subaerial reservoirs and not subglacial cupolas, thus precluding direct application of Clarke's entire model to ice cap and ice sheet settings. Notwithstanding this, *Björnsson* [1992] exploited elements of Clarke's model, using known hypsometric data [*Björnsson*, 1988], to analytically simulate jökulhlaups from subglacial lake Grímsvötn, Iceland.

2.4. Factors Governing the Duration of Waning Stage Discharge

[17] Dynamics responsible for jökulhlaup cessation are generally enigmatic, and most computations fail to mimic river stage recession from Q_{MAX} [*Björnsson*, 1992]. Conduit inlet Q_{MAX} represents maximum localized hydrodynamic efficiency and minimum closure by plastic deformation of ice. Immediately after the incidence of Q_{MAX}, glaciostatic and glaciohydraulic processes determine the period of waning stage Q. For jökulhlaups emerging from thin Alpine glaciers, say, <80 m ice thickness, glaciostatic pressure is below the threshold for instantaneous α activation [*Paterson*, 1994]; therefore conduit closure by ice creep has little direct effect on waning stage hydrodynamics [*Clarke*, 1982]. Instead, mechanical blockage of subglacial conduits, particularly at the conduit inlet [*Mathews*, 1973; *Sturm and Benson*, 1985] and outlet [*Paige*, 1955], can suppress and occasionally impede the release of waning stage efflux (see section 4.6). Other factors known to impede subglacial passage of floodwater include the (1) down-glacier presence of thermal boundary in the form of ice below the pressure-determined melting point [*Liestøl*, 1977; *Skidmore and Sharp*, 1999]; (2) interaction of a jökulhlaup with the distributed drainage network of an actively surging glacier [*Björnsson*, 1998]; and (3) intraglacial formation of supercooled floodwater by *hydraulic supercooling*, which acts to roughen and throttle flood conduits because of growth and ice wall adhesion of sediment-bearing frazil [*Shreve*, 1985; *Roberts et al.*, 2002].

[18] *Björnsson* [1992] acknowledged that the assumption of a cylindrical conduit inlet may be erroneous for several jökulhlaup systems, concluding that settling of a flat-based ice dam onto smoothed bedrock may expedite termination of discharge at the seal. Where the flank of a glacier flows partly into a tributary valley, subglacial conduits conveying water from the valley are subject to lateral compression with respect to their longitudinal axis [*Spring and Hutter*, 1982]. Previous glaciological studies have given scant regard to the effects of lateral deformation on subglacial hydrodynamics, and as *Tweed and Russell* [1999] noted, lateral deformation is probably an important and underestimated processes of jökulhlaup cessation.

2.5. Empiricisms for Estimating Jökulhlaup Magnitude

[19] Remarkably simple, independently corroborated empiricisms exist for estimating Q_{MAX} at a glacier margin. Using field data from 10 ice-marginal lakes (including prehistoric Lake Missoula), *Clague and Mathews* [1973] were the first to identify a power law relation between Q_{MAX} and *V*, revealing that *Q* is not a direct function of time since jökulhlaup onset but some power of the cumulative volume of water released from storage. The Clague-Mathews relation

$$Q_{\mathrm{MAX}} = bV^a \quad (10)$$

is valid for most tunneled jökulhlaups. In equation (10), Q_{MAX} and V are measured in $\mathrm{m}^3\ \mathrm{s}^{-1}$ and $10^6\ \mathrm{m}^3$, respectively; and b and a are regression coefficients derived from field data, varying significantly between flood systems [*Walder and Costa*, 1996]. Subsequent revisions of the Clague-Mathews relation have refined the parameters b and a to fit (1) modified and extended data sets [*Walder and Costa*, 1996; *Ng and Björnsson*, 2003]; (2) specific field sites [*Clarke and Mathews*, 1981; *Desloges et al.*, 1989; *Björnsson*, 1992]; (3) jökulhlaups generated from pervasive collapse of subglacial water-filled cavities [*Haeberli*, 1983]; (4) subaerial failure of glacier-proximal lakes [*Evans*, 1986]; and (5) drainage of ice-marginal lakes by breach widening of an ice dam [*Walder and Costa*, 1996].

[20] The physical underpinning of the Clague-Mathews relation for tunneled jökulhlaups has been demystified by *Ng and Björnsson* [2003], who explained how Q_{MAX} in equation (10) arises from *Nye*'s [1976] model of time-dependent water flow in a subglacial conduit coupled to an emptying lake; additionally, they argued that the power law relation in equation (10) is an artifact of the flood trigger mechanism and attendant hydrodynamics. Therefore site-specific factors such as conduit geometry and hydraulic potential are responsible for the order-of-magnitude variations of b and a reported in glaciological literature. Although the Clague-Mathews relation is a simple first-order

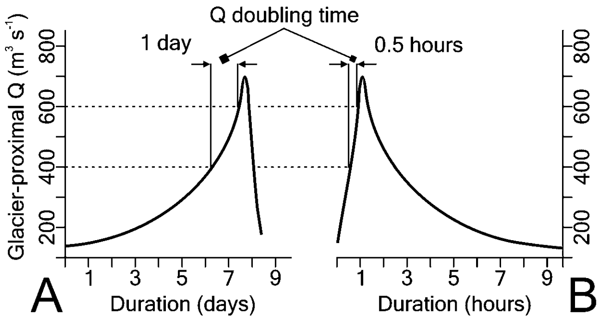

Figure 4. Idealized (a) exponentially and (b) linearly rising hydrographs constructed from glacier-proximal Q derivations. Note the contrasting timescales represented in the abscissas of Figures 4a and 4b.

approximation of jökulhlaup magnitude, it is stressed that potentially unknown factors such as the thermal potential of the reservoir can augment Q_{MAX} drastically.

[21] Jökulhlaups ensuing from sudden release of water from subglacial cavities [*Walder and Driedger*, 1995], occurring typically over a period of ~15–60 min [*Haeberli*, 1983], can achieve a glacier-proximal $Q_{\mathrm{MAX}} \sim 1/1000$ of total flood volume, resulting in downstream propagation of a flood wave. Such flood magnitudes correspond to a water equivalent volume ~10–100 mm thick over the entire ice catchment [e.g., *Merrand and Hallet*, 1996]. Expressing floodwater release as a product of affected subglacial area (A) and water layer depth (d), and assuming a glacier-proximal hydrograph with a symmetrically shaped rise and fall, *Walder and Fountain* [1997] proposed

$$Q_{\mathrm{MAX}} \approx \frac{2Ad}{t}. \tag{11}$$

Despite the inherent difficulties of estimating plausible bounds on A and d such basic equations are crucial for constraining likely flood magnitudes and consequential hazards.

2.6. Inferences Derived From Jökulhlaup Hydrographs

[22] To help elucidate jökulhlaup dynamics, it is useful to consider the relation $Q = f(t)$. To illustrate this, Figure 4 depicts two idealized classes of glacier-proximal hydrograph. Discharge in Figure 4a rises exponentially (i.e., increasing by a constant percentage of net Q per unit time), whereas Q in Figure 4b increases linearly at a rate significantly greater than exponential (i.e., by a constant amount per unit time, suggesting equality between Qt); following Q_{MAX}, both hydrographs display an intermediate relative minimum rate of decrease. For Q derivations from glacier-proximal river stage data a sudden, linear rise to Q_{MAX} produces a distinctive hydrograph shape, recognizable by the skewed time series asymmetry of the rise to Q_{MAX} and subsequent recession to base flow. When the time series ratio of rising to waning stage efflux is $\ll 1$, hydrograph shape is characteristic of a linearly rising jökulhlaup (Figure 4b); conversely, a ratio ≥ 1 suggests exponentially rising Q (Figure 4a). However, it is emphasized that the true form of glacier-proximal hydrographs are masked frequently by (1) hydraulic dampening effects of glacier contact lakes [*Merrand and Hallet*, 1996], (2) downstream flood wave retardation and attenuation [*Desloges et al.*, 1989], (3) inputs from rainstorms and groundwater [*Björnsson*, 1977], and (4) processes identified in section 2.2.

[23] From equation (8) it follows that ΔQ in the hydrograph in Figure 4a increases by a factor of ~2.7 over progressively shorter intervals approximated by

$$\Delta Q = \frac{1}{KQ^{1/4}}. \tag{12}$$

During the initial rising stage of the jökulhlaup portrayed in Figure 4a, the Q doubling time is relatively long; although as Q increases, the doubling time shortens significantly. In contrast, the doubling time of the hydrograph in Figure 4b remains consistently short during the rising stage. Incorporating factors such as thermal potential and length of the subglacial flood path, jökulhlaups conforming to Figure 4a attain Q_{MAX} commonly 1–21 days after the onset of flooding [e.g., *Björnsson*, 1992], whereas events conforming to Figure 4b reach Q_{MAX} within hours to a day [e.g., *Flowers et al.*, 2004]. Generally, the duration of large jökulhlaups tends to be shorter than that of smaller ones [*Björnsson*, 1992, 2002].

[24] Hydrographs in Figures 4a and 4b reflect fundamentally different subglacial processes, as a linear relationship between Q and t cannot be explained solely by positive feedback between water flow and conduit enlargement [*Björnsson*, 1992, 2002, 2004; *Flowers et al.*, 2004]. Instead, linearly increasing Q suggests a short-term inefficient heat transfer between floodwater and glacier ice, as first advocated by *Björnsson* [1992]. However, field measurements of floodwater temperature at glacier termini confirm that during all types of jökulhlaups, floodwater temperature remains at (or slightly below) the pressure-determined melting point (≈0°C) of artesian discharge [*Rist*, 1955; *Björnsson*, 1988; *Snorrason et al.*, 2002]. Therefore sheet-like flow across large portions of the glacier bed is required to explain linearly rising jökulhlaups [*Björnsson*, 1992; *Jóhannesson*, 2002; *Flowers et al.*, 2004].

2.7. Precedents for a Revised Understanding of Jökulhlaup Dynamics

[25] Despite the utilitarian value of *Nye*'s [1976] model and its subsequent revisions some jökulhlaups reveal physical processes that confound conventional understanding [*Björnsson*, 1977, 1988, 1992, 2002; *Roberts et al.*, 2000, 2003; *Jóhannesson*, 2002; *Flowers et al.*, 2004] (Figure 5). These jökulhlaups are characterized by a steep, linear rise to Q_{MAX} within a period of minutes to a day [e.g., *Thórarinsson*, 1957, 1958; *Haeberli*, 1983; *Sigurðsson et al.*, 1992; *Walder and Costa*, 1996; *Walder and Fountain*, 1997] and maximum discharges several orders of magnitude greater than preflood base flow [e.g., *Tómasson*, 1996]. The

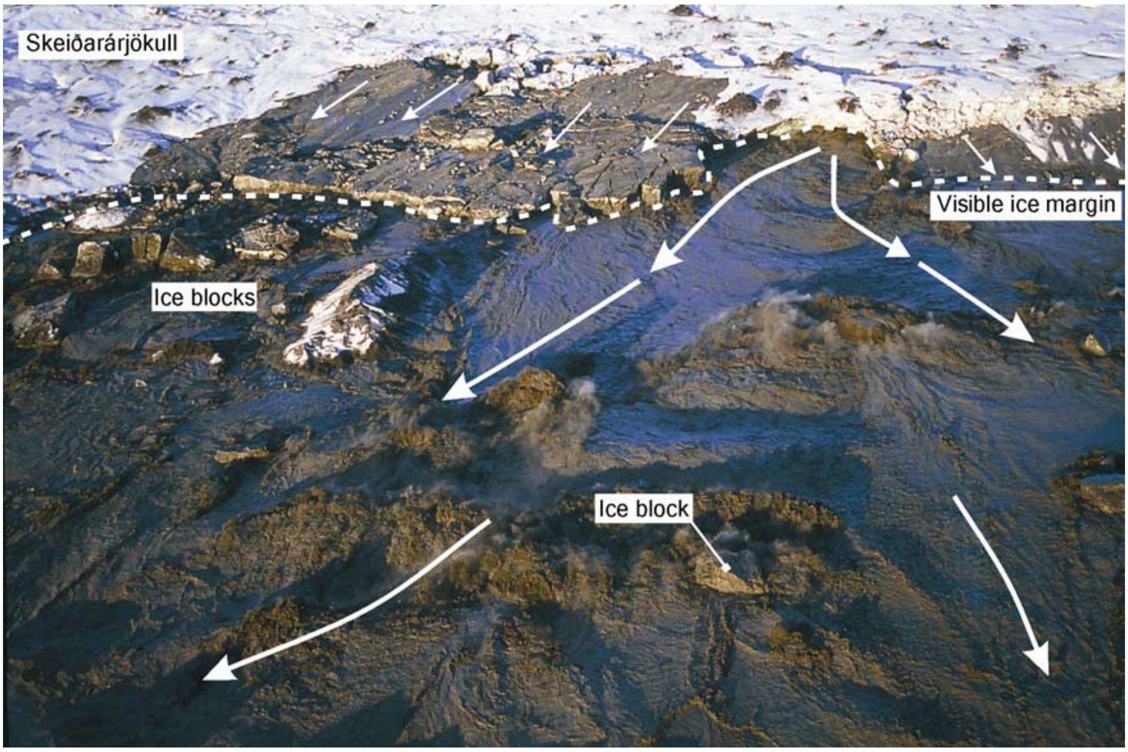

Figure 5. Oblique aerial photograph of a 400-m-wide portion of the terminus of Skeiðarárjökull, Iceland. The photograph was taken midway through the 18-hour rise of the November 1996 jökulhlaup (Q_{MAX} 4.5–5.3 × 10^4 m^3 s^{-1}). Prodigious quantities of floodwater are bursting simultaneously from beneath and above the ice margin. Ice blocks >20 m in diameter are in transit within hyperconcentrated floodwater. Photographer R. T. Sigurðsson.

mean acceleration of rising stage discharge ($\Delta Q/\Delta t$, m^3 s^{-2}) provides a simple means of differentiating linearly rising jökulhlaups from their exponential counterparts (Figure 6). For 35 gauged Icelandic jökulhlaups, $\Delta Q/\Delta t$ spanned 10^{-3}–10^1 m^3 s^{-2}, and data clustered in the regions $\leq$0.01 and $\gg$0.1 m^3 s^{-2} (Figure 6). From the previous data set I contend that a mean rate of increase of $\geq$0.1 m^3 s^{-2} is suggestive of a linearly rising jökulhlaup.

[26] Some linearly rising jökulhlaups can be explained by envisaging instantaneous water release from a subglacial cavity network [*Haeberli*, 1983]. In this context, flooding is triggered by rapid and pronounced water input to the glacier bed, usually from intense rainfall in the ablation zone [*Walder and Driedger*, 1995]. Alternatively, some linearly rising jökulhlaups originate from discrete reservoirs in either ice-marginal or subglacial locations [*Björnsson*, 2002, 2004]. Arguably, because of their infrequency, there has never been a precedent for a physical understanding of linearly rising jökulhlaups [*Björnsson*, 1992]. This explains why accepted physical models that aim to facilitate generic understanding [e.g., *Nye*, 1976] are incompatible with extraordinary jökulhlaups. Moreover, the seminal works of *Nye* [1976], *Spring and Hutter* [1981, 1982], and *Clarke* [1982] were based on prevailing knowledge of steady state subglacial drainage and attendant hydrodynamics [i.e., *Röthlisberger*, 1972; *Shreve*, 1972; *Weertman*, 1972].

[27] Since the early 1970s, glaciologists have revised their view of subglacial hydrodynamics substantially [*Fountain and Walder*, 1998], especially through progressive realization that subglacial drainage is a dynamic, feedback-driven system capable of attaining multiple metastable equilibria in time and space [e.g., *Clarke*, 1996]. Despite high-magnitude jökulhlaups being the ultimate manifestation of in-phase variations in P_w and Q, glaciologists have yet to undertake quantitative jökulhlaup analysis based on the assertion that unstable, nonlinear changes in flow geometry and drainage configuration are possible on a timescale of minutes to several hours.

[28] Disparity between theory and observation reflects judgments about the relative importance, applicability, and feasibility of processes that can be modeled. For instance, some field data suggest pervasive leakage of floodwater from subglacial conduits [e.g., *Nolan and Echelmeyer*, 1999] (see section 3.3), but theoretical modelers neglect this process because they infer that the conduit leakage will have no significant effect on the overall hydrodynamics of the jökulhlaup. Clearly, this belief is not applicable to linearly rising jökulhlaups, as significant conduit losses are a prerequisite to a linear rate of $Q(t)$.

3. TRIGGERING OF LINEARLY RISING JÖKULHLAUPS

[29] Thermodynamic and hydromechanical processes exclusive to the conduit inlet and seal area of a meltwater reservoir must combine to impart a sudden and sustained subglacial water flux capable of forcing $Q(t)$ into equity. In

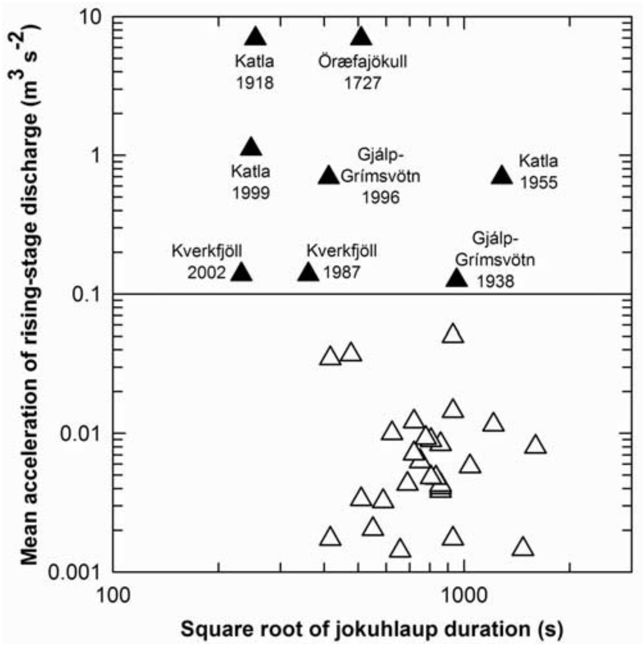

Figure 6. Log-log plot of rising stage $\Delta Q/\Delta t$ and flood duration for 35 Icelandic jökulhlaups. Data sources for solid triangles are Katla [*Thórarinsson*, 1957; *Tómasson*, 1996]; Kverkfjöll [*Sigurðsson et al.*, 1992; *Sigurðsson et al.*, 2000]; Gjálp-Grímsvötn [*Björnsson*, 2002]; and Öræfajökull [*Thórarinsson*, 1958]. Data sources for open triangles are Brúarjökull, 1986 [*Sigurðsson et al.*, 1992]; Gjánúpvatn, 1951 [*Thórarinsson*, 1974]; Grímsvötn [*Thórarinsson*, 1974; *Björnsson*, 1988; S. Ó. Elefsen, unpublished data, 2004]; Grænalón, 1939 and 1973 [*Thórarinsson*, 1974]; eastern Skaftá cauldron, 1984, 1986, 1989, 1991, 1995, 1997, 2000, and 2002 [*Zóphóníasson*, 2002]; western Skaftá cauldron, 1988, 1994, 1996, 2000, and 2002 [*Zóphóníasson*, 2002]; and Vatnsdalón, 1898 and 1974 [*Thórarinsson*, 1974]. Errors in Q_{MAX} accuracy are considerable because of the highly variable quality of hydrometric data.

sections 3.1–3.3 I assess likely processes responsible for the initiation of linearly rising jökulhlaups.

3.1. Sudden Release or Generation of Meltwater

[30] Hydromechanical rupturing of an ice dam could facilitate rapid, unsteady water release. Processes responsible for dam rupturing include (1) calving of a large, deep-seated section of an ice dam [*Walder et al.*, 2005], caused either by a sudden rise in lake level or by the presence of water-filled crevasses in the dam area [*Sturm and Benson*, 1985]; (2) intrusive hydrofracturing at the base of an ice dam, when hydrostatic pressure exceeds ice strength [*Glen*, 1954; *Fowler*, 1999]; and (3) for meltwater reservoirs in zones of intense seismicity, earthquake-induced faulting of an ice dam. Despite the theoretical tenability of dam weakening by strong seismic radiation, there are no published accounts of such a process connected to jökulhlaup initiation [*Tweed and Russell*, 1999]. Nonetheless, large-scale ice fracturing can occur when an earthquake's epicenter is located beneath a glacier [e.g., *Thouret*, 1990].

[31] Triggering of the November 1996 jökulhlaup from Skeiðarárjökull deserves particular consideration. Between late September and mid October 1996 a subglacial volcanic eruption occurred at Vatnajökull [*Gudmundsson et al.*, 1997, 2004]. Over 35 days, ~2.7 km^3 of eruption-induced melt entered Grímsvötn, adding to ~0.5 km^3 of meltwater already in the lake [*Gudmundsson et al.*, 1997]. A jökulhlaup was expected when lake height corresponded to 1450 m above sea level (asl); nevertheless, meltwater continued to collect until lake height reached 1510 m asl. At this height, Grímsvötn exerted hydrostatic pressure equivalent to the minimum hydraulic potential in the region damming the lake [*Björnsson*, 2002]. Consequently, at 2300 UT on 4 November a large zone of the Grímsvötn ice dam floated, allowing subglacial escape of floodwater into Skeiðarárjökull. This type of trigger mechanism had not operated at Grímsvötn for $\geq$60 years [*Björnsson*, 2002]. I regard the combination of warm floodwater (section 3.2) and excess hydrostatic stress at the conduit seal as fundamental to the extraordinarily swift emptying of Grímsvötn in November 1996 [see also *Fowler*, 1999]; similar processes are thought to have operated during volcanically related jökulhlaups from Skeiðarárjökull in 1861, 1892, 1934, and 1938 [*Björnsson*, 1988].

[32] Some linearly rising jökulhlaups are an immediate corollary of subglacial volcanism [*Thórarinsson*, 1957, 1958; *Baker et al.*, 1969; *Major and Newhall*, 1989; *Pierson*, 1989; *Best*, 1992; *Thouret et al.*, 1995; *Tómasson*, 1996]. Tremendous melt results from phreatomagmatic eruptions at the base of an ice mass, with rapid enthalpy release occurring via lava quenching and attendant vigorous convection of boiling water and steam [*Höskuldsson and Sparks*, 1997; *Smellie*, 2002; *Wilson and Head*, 2002]. Melt rate at an ice interface by convection of a hot fluid can be estimated by

$$M_R = \frac{H_{\mathrm{FLUX}}}{\rho_i[C_i(T_{\mathrm{MELT}} - T_{\mathrm{SOLID}}) + L]}, \tag{13}$$

where M_R expresses melt rate (m s^{-1}), H_{FLUX} is the heat flux of the solid being melted (W m^{-2}), C_i is the heat capacity of ice (2.01 $\times$ 10^3 J kg^{-1} °C^{-1}), T_{MELT} is the melting point of ice (~0°C), and T_{SOLID} is antecedent ice temperature (~0°C). Assuming instantaneous and complete enthalpy release, one unit of lava melts about 14 times the same volume of ice [*Wilson and Head*, 2002]. Meltwater within a growing eruption cavity can reach temperatures nearing 20°C [*Gudmundsson et al.*, 2004], causing transient changes in cavity shape and volume. Besides melting through subaqueous lava-ice interaction, gaseous juvenile water can be released in large quantities from the eruption edifice [*Mastin*, 1995]. While an eruption remains subglacially confined, condensed juvenile water can add to the volume of ice melted (V_{ICE}), assisting the growth of a large and inherently unstable meltwater reservoir. *Smellie* [2002] proposed that dynamic advection of gaseous volatiles and steam against an ice face should promote melt rates approximately twice greater than would be possible across a meltwater interface; however, the thermodynamic significance of gas-driven ice melting is currently unknown. Although cavity volumetric changes induced by lava extrusion (V_{LAVA}) and associated ice melt (V_{MELT}) are roughly equal, slight imbalances affect cavity P_w significantly [*Björnsson*, 1988; *Gudmundsson et al.*, 2004]. Cavity P_w (C_{Pw}) is resolved by

$$\frac{\Delta V}{V_{\mathrm{MELT}}} = \frac{(C_{PW} - P_i)}{B}. \tag{14}$$

In equation (14), ΔV = (V_{MELT} + V_{LAVA} − V_{ICE}) and B represents the bulk modulus (volumetric strain); for water, B is 2.05 GPa [*Blake*, 1981]. Because ΔV is typically small, massive increases in C_{PW} are probable, sometimes exceeding ambient glaciostatic stress. Consequently, for an eruption under hundreds of meters of ice, lava intrusion can generate lithostatic forces much greater than glacier yield strength, thus facilitating ductile deformation of the cavity walls and roof. However, because of the rapidity of melting, ice deformation cannot equilibrate C_{PW} and P_i; hence meltwater can escape by localized glacier flotation [*Gudmundsson et al.*, 2004]. Imposition of negative effective pressure therefore inhibits the formation of a large water-filled cavity; instead, it facilitates the generation of a nontunneled subglacial flood wave (see section 3.3). In summary, subglacial volcanism results in direct meltwater dispersal and little in situ storage [*Gudmundsson et al.*, 2004], with the rate of magma extrusion and enthalpy extraction determining the amount of melt draining down glacier [*Wilson and Head*, 2002].

3.2. Rapid Heat Exchange Between Water and Ice

[33] *Gilbert* [1971] and *Mathews* [1973] simulated dynamic fluid enthalpy release in a circular conduit of a uniform cross section using the classic, dimensionless power law extracted from *McAdam* [1951], where

$$Nu = 0.023 Re^{4/5} Pr^{2/5}. \tag{15}$$

In equation (15), Nu is the Nusselt number, defined by the ratio of actual heat transfer to purely conductive heat transfer, Re is the Reynolds number, expressing the degree of flow turbulence, and Pr is the Prandtl number, characterizing the thermal and mechanical properties of the fluid. Dimensionless Nu, Re, and Pr numbers are derived as

$$Nu = hD/k_w, \quad Re = \rho_w vD/\eta, \quad Pr = C_w\eta/k_w, \tag{16}$$

where h is an empirical heat transfer coefficient, valid only for prescribed Q conditions and conduit geometries [*Spring and Hutter*, 1982; *Clarke*, 2003]; D is conduit diameter (m); k_w is the thermal conductivity of water (~0.56 W m^{-1} °C^{-1}); v is mean flow velocity (m s^{-1}); η is dynamic water viscosity (~1.79 × 10^{-3} kg m^{-1} s^{-1}); and C_w is the specific heat capacity of water (~4196 J kg^{-1} °C^{-1}). For enthalpy release to conduit walls, floodwater temperature (θ_w) must exceed ice temperature (θ_i); however, it is likely that $\theta_w - \theta_i$ varies significantly with down-glacier distance from the jökulhlaup source. It is assumed widely that heat generated by potential energy reduction is transferred to conduit walls at the same rate it is released [e.g., *Paterson*, 1994]. Hence floodwater temperature decreases with increasing distance down glacier to some value (θ_0). From *Clarke and Mathews* [1981] the volume of ice melted (M) per unit distance per unit time is given by

$$M = \frac{Q\left(-\dfrac{\partial\phi}{\partial s}\right)}{L'} + \frac{0.205}{L'}\left(\frac{2Q\rho_w}{\pi^{1/2}S^{1/2}\eta}\right)^{4/5} k_w(\theta_o - \theta_i). \tag{17}$$

In equation (17), L' expresses the effective heat of melting, as defined by $L + C_w(\theta_w - \theta_i)$; the term second from right is an expression for enthalpy release from a turbulent fluid flowing in a straight, cylindrical pipe [see *McAdam*, 1951]. This empiricism is rooted in all physical treatments of jökulhlaup hydrodynamics [e.g., *Mathews*, 1973; *Nye*, 1976; *Spring and Hutter*, 1981, 1982; *Clarke*, 1986]. *Jóhannesson* [2002] demonstrated that a simple modification to equation (17) allows rectangular channel cross sections to be represented. Despite the physical inaptness of an ice-lined cylindrical or semicircular conduit as the ideal for jökulhlaup drainage, glaciologists have yet to see quantitative thermodynamic analysis based on more glaciologically realistic flow configurations and experimentally determined relationships between Nu, Re, and Pr numbers [*Clarke*, 2003].

[34] On the basis of simulations of $\Delta Q/\Delta t$, *Björnsson* [1992] contended that actual enthalpy release from turbulent floodwater is markedly more effective than results of equations (15) and (17) suggest. *Jóhannesson* [2002] tested equation (17) in relation to observations of a supraglacial canyon, representing ~0.3 km^3 ice loss [*Björnsson*, 2002], which outlined 6 km of the 50 km subglacial flood path from Grímsvötn during the November 1996 jökulhlaup [*Björnsson et al.*, 2001]. Assuming that the canyon was a surface expression of subglacial ice melt and that down glacier from the canyon the front of the jökulhlaup was at the ambient ice temperature (−0.19°C), mean lake temperature was ~8°C [*Björnsson*, 2002]. On the basis of this insight, Jóhannesson demonstrated that maximum enthalpy extraction had occurred over a distance at least an order of magnitude less than that predicted by the model of *Nye* [1976]. The spatial and temporal distribution of thermal energy along a subglacial flood path is therefore imbalanced, exhibiting rapid exponential decay.

3.3. Propagation of a Subglacial Flood Wave

[35] A subglacial distension wave is a body of pressurized fluid ($P_w \geq 1\ P_i$) capable of decoupling a glacier locally from its bed to produce ice surface uplift by hydraulic jacking [*Iken et al.*, 1983; *Jóhannesson*, 2002; *Clarke*, 2003; *Flowers et al.*, 2004] (Figure 7). Kinematic waves of subglacial meltwater are often observed indirectly during intense rainfall episodes [*Raymond et al.*, 1995],

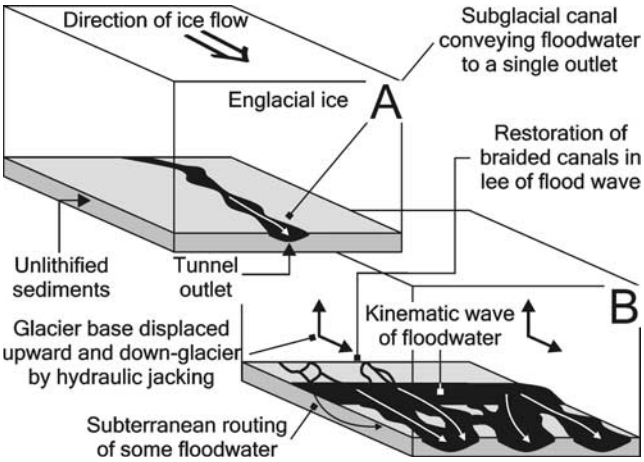

Figure 7. Cartoons of subglacial drainage configuration during an (a) exponentially rising and (b) linearly rising jökulhlaup.

spring events [*Iken et al.*, 1983; *Röthlisberger and Lang*, 1987], surges [*Kamb et al.*, 1985; *Björnsson*, 1998], and jökulhlaups [*Nolan and Echelmeyer*, 1999; *Flowers and Clarke*, 2000]. The magnitude of negative effective pressure at the leading edge of the flood wave is determined dynamically by the enormity of short-period hydraulic impulses (i.e., $Q/S \geq 1$) [*St. Lawrence and Qamar*, 1979; *Spring and Hutter*, 1982; *Kavanaugh and Clarke*, 2000]; local hydraulic gradients and drainage connectivity [*Flowers and Clarke*, 2000]; the confining properties of glacier ice [*van der Veen*, 1998]; and the hydraulic properties of macroporous subglacial sediments [*Nolan and Echelmeyer*, 1999]. Pronounced glacier uplift and enhanced basal sliding are often accompanied by ice-sourced seismic tremors [*St. Lawrence and Qamar*, 1979], icequake swarms [*Wolf and Davies*, 1986], ice surface rupturing (see section 4.1), and, in extreme situations, supraglacial outbursts of subglacially derived meltwater (see sections 4.1. and 4.2.).

[36] For the rise of the 1996 jökulhlaup from Skeiðarárjökull ($\Delta Q/\Delta t \sim 0.62$ m^3 s^{-2}), *Björnsson* [1998] reasoned that $\sim 10^8$ m^3 of floodwater was in subglacial transit 6 hours before Q_{MAX} ($4.5-5.3 \times 10^4$ m^3 s^{-1}) was reached. Assuming a triangular-shaped area of floodwater inundation beneath Skeiðarárjökull, with a source point at Grímsvötn and corner points at the distal edges of Skeiðarárjökull, gives a mean subglacial water depth of $\sim$1.7 m over $\sim$575 km^2 at the time of maximum floodwater storage. Subglacial effluxes proportional to a $\geq$1 m deep channel covering a full glacier cross section, together with observations of prominent ice surface upheaval, imply that localized tracts of the glacier bed are subject to fleeting sheet flow conditions during cataclysmic jökulhlaups [*Shoemaker*, 1992a]. It is apparent that linearly rising jökulhlaups translate work done by moving subglacial floodwater mainly to hydrostatic pressure rather than to additional ice melt by viscous dissipation of heat. This notion explains why linearly rising hydrographs have reversed symmetry when compared to their exponential counterparts. For volcanogenic jökulhlaups the relative density and dynamic viscosity of the subglacial flood wave determines the magnitude of vertical ice surface displacement. Phreatic activity can force the subglacial propagation of debris flows composed of juvenile eruptives [*Major and Newhall*, 1989]. Flows of hot rocks and gas melt ice progressively to produce slurries with increasingly lower kinematic viscosity; therefore it is expected that slurry dilution and associated fluid-density reduction occur dynamically during subglacial transit [see *Iverson*, 1997].

[37] Valley glaciers subjected to volcanogenic jökulhlaups are liable to incur more hydromechanical disruption than piedmont glaciers such as Skeiðarárjökull [e.g., *Branney and Gilbert*, 1995; *Thouret et al.*, 1995]. This is because the $Q(t)$-dependent switch from predominantly channelized subglacial drainage to distributed drainage is likely to occur simultaneously throughout a comparatively large area of the submerged glacier bed [e.g., *Weertman and Birchfield*, 1983], especially if the density of the jökulhlaup reduces effective pressure significantly. Consequently, negative effective pressure is generated over large tracts, and insidious, irreversible modifications in glacier stress ensue, leading to abnormally high basal shear stresses, resultant tensional ice fracturing [*Shoemaker*, 1992a], and ultimately runaway glacier advance [*Weertman*, 1962]. Those skeptical of high-magnitude jökulhlaups triggering calamitous glacier upheaval should review descriptions by *Thórarinsson* [1958, pp. 29–33]. Neither subglacial distension waves nor Bingham flow characteristics can be reconciled to current hydrodynamic theories of jökulhlaup propagation [e.g., *Shoemaker*, 1992a; *Walder*, 1994]. Similarly, spatially inhomogeneous deficits in effective pressure, particularly at the leading edge of a flood wave, have no parallel in prevailing jökulhlaup knowledge [*Jóhannesson*, 2002; *Clarke*, 2003; *Flowers et al.*, 2004].

[38] Further quantitative thought must be given to the hydraulic properties of floodwater in subglacial transit [*Spring and Hutter*, 1981; *Björnsson*, 1992; *Clarke*, 2003]. For instance, profuse intraglacial growth of frazil ice can result in the glacier-proximal formation of a fluid-ice mix with Bingham flow properties; but the hydraulic significance and erosive effects of a turbulent, two-phase medium within intraglacial drainage remains to be determined [*Clarke*, 2003]. It is manifestly clear that freshwater density assumptions are unrepresentative of rising stage efflux; moreover, it is self-evident that hydrodynamically significant fluid-density augmentation occurs during the initial stages of jökulhlaup propagation. My Iceland-centric jökulhlaup examples suggest that floodwater be treated as a dilatant, compressible, non-Newtonian fluid. On this premise, Q along the subglacial flood path should be considered anisotropic as a non-Newtonian flood wave would promote dynamic water storage, hence spatial variations in instantaneous Q.

[39] Eradicating the assumption that $\nabla\phi$ and S are constant for the subglacial flood path [*Walder*, 1994; *Clarke*, 2003] will allow more accurate determination of the initial route and probable breadth of preferential drain-

age axes. Such an approach requires data input from laborious surveys of three-dimensional glacier geometry [e.g., *Elvehøy et al.*, 2002]. Modification of the existing Gauckler-Manning formula (equation (5)) facilitates Q approximations for broad, rectangular cross sections. For a subglacial flood wave whose horizontal dimensions greatly exceed water depth, the Q flux (m^3 m^{-1} s^{-1}) is

$$Q_{\mathrm{FLUX}} = \frac{d^{5/3}}{2^{2/3}n}\sqrt{(-\partial\phi/\partial s)/(\rho_w g)}. \tag{18}$$

However, like equation (5), the Achilles' heel of equation (18) is the inescapable parameterization of a dubious value of n. However, *Walder and Fowler* [1994] and *Fowler and Ng* [1996] made a laudable attempt to resolve effective Manning's roughness by considering the effects of drag imparted by the unconsolidated floor and ice-covered roof of subglacial canals [see also *Clarke*, 2003].

[40] Contemporary observations of linearly rising jökulhlaups [*Björnsson*, 2002] and theoretical treatments of sheet flows [*Weertman and Birchfield*, 1983] debouching from piedmont ice lobes confirm that channelized subglacial drainage prevails eventually (Figure 8). For example, when the 1996 jökulhlaup neared Q_{MAX}, concomitant changes in outlet location and geometry signified the pervasive disablement of a distributed drainage system and the restoration of high-capacity channelized drainage [*Flowers et al.*, 2004] (Figure 8). It can be concluded that channelized subglacial flow is the most stable means of transmitting large volumes of meltwater [cf. *Shoemaker*, 1992a, 2002], especially if flood channels are embedded in unconsolidated sediments [*Walder*, 1994; *Walder and Fowler*, 1994; *Catania and Paola*, 2001].

4. GLACIER RESPONSES TO LINEARLY RISING JÖKULHLAUPS

4.1. Outbursts of Floodwater Onto the Glacier Surface

[41] Supraglacial outpourings of subglacially derived floodwater are a corollary of subglacial hydraulic transience [*Mathews*, 1973; *Roberts et al.*, 2000; *Clarke*, 2003], generated either by a temporary mechanical constriction [e.g., *Reid and Clayton*, 1963], a thermal boundary [*Skidmore and Sharp*, 1999], or by a propagating wave of floodwater (section 3.3 and references therein). Supraglacial outbursts created during jökulhlaups emanate from a variety of outlet morphologies. Documented outlet types include (1) moulins [*Warburton and Fenn*, 1994]; (2) surface crevasses [*Goodsell et al.*, 2003], ice folia, tension veins, and shear planes [*Goodwin*, 1988]; (3) exposed englacial conduits [*Thouret*, 1990]; and (4) discrete, flood-induced fractures [*Liestøl*, 1977]. Artesian outpourings 1–2 m higher than the ice surface (i.e., $P_w \approx 1.1\ P_i$) are common for supraglacial outlets that evacuate subglacial floodwater [e.g., *Goodwin*, 1988]. However, significantly larger hydraulic imbalances are implied from historical reports. For example, observers of the 1823 Katla eruption, which occurred beneath the Mýrdalsjökull ice cap, Iceland, witnessed the supraglacial breakout and up-glacier movement of giant, curtain-like fountains of turbid floodwater that extended tens of meters above the ice surface (i.e., $P_w > 1.2\ P_i$) [*Austmann*, 1907].

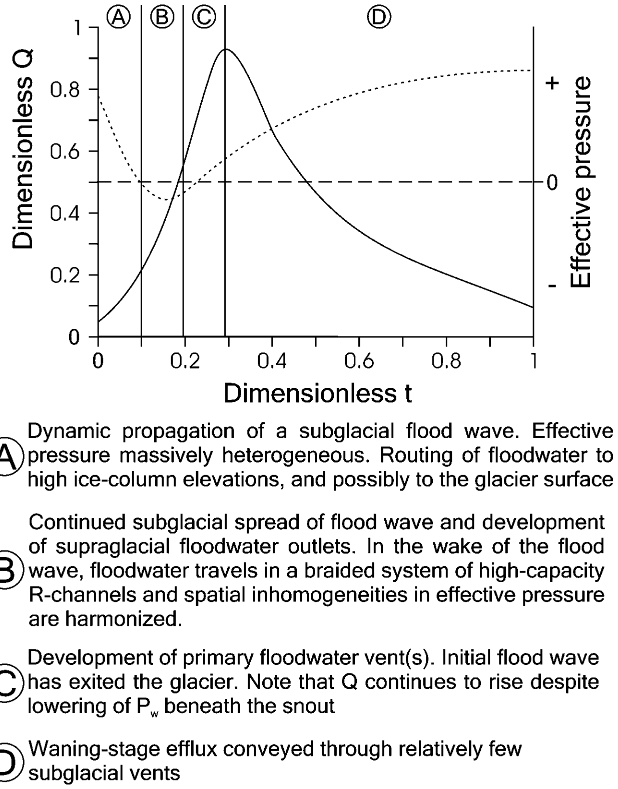

Figure 8. Conceptualized hydrograph (solid line) showing changes in effective pressure (dotted line) and hydraulic conditions during the subglacial propagation of a flood wave.

[42] Subglacial flood paths composed of geometrically diverse orifices engender hydraulic impulses because of spatially nonuniform changes in water velocity [*Clarke et al.*, 2004]. Consequent wave propagation times range from slow ($t = \gg 2L/C$) to instantaneous ($t = 0$), where L denotes the length of the affected reach of the drainage system and C defines characteristic wave celerity (m s^{-1}) [*Walski et al.*, 2003], expressed as

$$\sqrt{\frac{B}{\rho_w}} = \sqrt{\frac{\Delta P_w}{\Delta\rho_w}}. \tag{19}$$

For a freshwater density of 998 kg m^{-3} and a corresponding elasticity assumption of 2.05 GPa, C is 2 orders of magnitude greater than g; in contrast, C is at least an order of magnitude less than g for a notional fluid density and modulus of elasticity of 1300 kg m^{-3} and 0.001 GPa, respectively. Thus the compressibility of floodwater determines the propensity for hydraulic transience within subglacial drainage [see also *Clarke*, 2003]. Because subglacially propagating slurries cause effective pressure to lower (section 3.3), any ensuing supraglacial outburst is

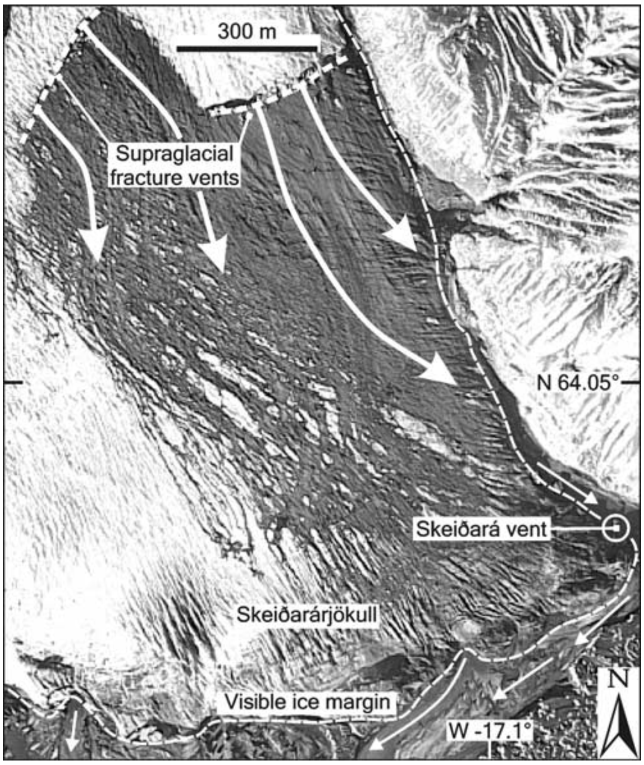

Figure 9. Near-vertical aerial photograph of a supraglacial outburst of floodwater on the eastern flank of Skeiðarárjökull, Iceland, during the November 1996 jökulhlaup. Ice thickness in the region of the fracture outlets is ~300 m. Photograph courtesy of Landmælingar Íslands, 1997.

due to buoyancy-driven extrusion of Bingham floodwater [*Jóhannesson*, 2002]. In contrast, supraglacial outbursts of Newtonian floodwater rely on subglacial hydraulic transience to achieve overpressurization.

[43] *Roberts et al.* [2000] related the temporal occurrence of supraglacial outbursts to the magnitude of Q during the 1996 jökulhlaup (section 3.3), concluding that the gradient of near-ice-surface P_w in excess of hydrostatic P_w determines the type of intraglacial drainage topology that forms. A near-surface P_w gradient equivalent to the static weight of freshwater (i.e., ~9.8 kPa m^{-1}) allows moulins, open crevasses, and exposed englacial conduits to decant floodwater across the glacier surface. Conversely, a near-surface P_w gradient $\geq$2 kPa m^{-1} in excess of hydrostatic conditions facilitates water pressure-induced fracturing at the glacier surface (Figure 9).

4.2. Water Pressure-Induced Ice Fracture

[44] Hydraulic waves propagating rapidly (i.e., t = $<2L/C$) within subglacial flood circuits can induce water hammer events because of high pulses in P_w [e.g., *Kavanaugh and Clarke*, 2000]. Hydraulic pressure in excess of P_i and a critical value of the stress intensity factor of glacier ice (~100 kPa m$^{1/2}$) will result in the near-instantaneous growth of a vertical, fluid-driven crack in a tensile stress field toward the glacier surface [*van der Veen*, 1998]. Brittle features at the glacier base provide an englacial opening that can be hydraulically fractured apart, thus allowing floodwater to race toward the fracture tip with each successive split. Sediments preserved in relict hydrofractures record instantaneous fracture widths of 0.001–2 m [*Ensminger et al.*, 2001; *Roberts et al.*, 2001].

4.3. Removal of Ice Fragments

[45] Jökulhlaups produce an abundance of ice fragments, ranging from millimeter-sized shards to decimeter-sized blocks (Figure 10). Ice fragmentation is not confined solely to the flanks of subglacial outlets; it can occur simultaneously at the jökulhlaup source [*Sturm and Benson*, 1985], on the glacier surface [*Major and Newhall*, 1989], within intraglacial locations [*Warburton and Fenn*, 1994], and at the ice margin [*Russell and Knudsen*, 1999]. Voluminous (e.g., $\geq 10^8$ m^3) ice fragmentation can exceed the volume of flood-induced melt from subglacial conduits, making ice release a dominant mass transfer process during some jökulhlaups.

[46] The smallest ice fragments ($\leq$0.1 m in diameter) released from outlets come from the ice interface of the intraglacial flood path [*Mathews*, 1973; *Björnsson*, 1992], and candidate erosion processes include (1) hydrofracturing at the crystal boundary scale [*Iverson*, 1991], (2) water vapor implosions (cavitation) [*Barnes*, 1956], and (3) ice comminution due to percussive striking of bed load and suspended load against conduit walls [*Spring and Hutter*, 1981]. Observers of high-pressure artesian outlets (fountain velocities $\geq$10 m s^{-1}) often note effervescent floodwater [e.g., *Lawson*, 1993], diagnostic of water at the critical pressure for cavitation [*Barnes*, 1956]. Because of order-of-magnitude variations in $-\nabla\phi$ and high Q, subglacial floodwater coursing through geometrically heterogeneous channels is liable to repeatedly satisfy vapor pressure conditions for bubble formation [*Mathews*, 1973], and channel roughness imparted by frazil ice is likely to accentuate cavitation rates. Cavitation is a rarely cited process in jökulhlaup literature, yet it probably erodes ice more effectively and pervasively than intraglacial hydrofracturing or in situ ice comminution. The erosive significance and hydraulic and thermodynamic consequences of intraglacial cavitation deserves further evaluation by jökulhlaup scientists.

[47] Ice fragments $\geq$1 m in diameter originate typically from the flanks of floodwater outlets. Few ice blocks are released from artesian outlets that remain stationary; however, hydraulic gradients imposed by linearly rising jökulhlaups can force outlets to migrate. Under such conditions a drifting artesian outlet can hydraulically fragment large areas of ice [*Roberts et al.*, 2000]. Because of structural instability, ice-roofed outlets are capable of massive ice release. Rapid fluvial undercutting of ice walls results in mechanical splaying and ensuing ice collapse, which weakens the ice roof further. Giant, cube-shaped ice blocks are produced in deepwater zones by rapid flexure of ice in response to torque imposed by glacier buoyancy [*Sturm and Benson*, 1985; *van der Veen*,

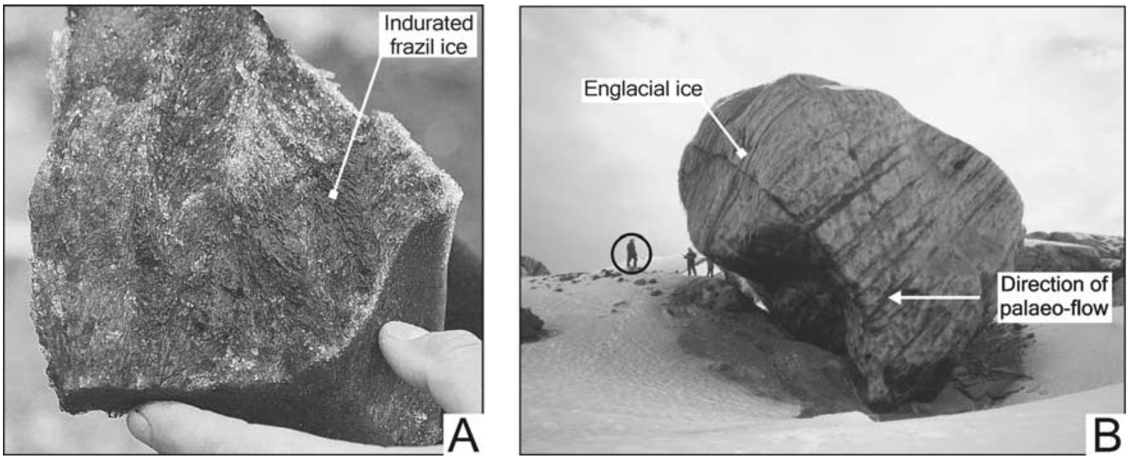

Figure 10. Contrasting sizes of ice fragments released from Skeiðarárjökull, Iceland, during recent jökulhlaups. (a) Angular fragment of frazil ice. (b) Subangular block of englacial ice fractured from Skeiðarárjökull during the 1996 jökulhlaup (note person (circled) for relative scale). Photographers M. J. Roberts (Figure 10a) and R. T. Sigurðsson (Figure 10b).

1998]. The most massive ice dislocations occur when a subglacial flood wave speeds toward a comparatively thin ice margin, thereby reducing normal stresses and amplifying shear strain; this causes shear faulting perpendicular to principal ice stresses and the creation of immense, slab-shaped glacier fragments.

[48] Circumstantially, it is clear that as $\Delta Q/\Delta t$ increases, so too does ice block size [e.g., *Russell and Knudsen*, 1999]. However, if scientists are to contribute effectively to hazard mitigation strategies for jökulhlaups, then statistical analyses of ice block geometry, frequency, and spatial distribution in relation to specific jökulhlaup types and outlet configurations is required. Relevant quantitative data and improved hydromechanical understanding of ice fracture processes will allow tighter constraint on the probable ice-marginal locations of intensive ice erosion during jökulhlaups.

4.4. Development of Giant Floodwater Outlets

[49] Sustained rising stage efflux from outlets in zones where flood-imposed $-\phi$ is asymptotic will force intense mechanical breakup of an ice margin, and, if ice fragments are removed hydraulically, a sizable ice-walled canyon will form [*Richardson*, 1968; *Trabant et al.*, 1994; *Fleisher et al.*, 1998; *Russell et al.*, 2001]. The 1918 Katla eruption [*Tómasson*, 1996] generated a spontaneous jökulhlaup ($\Delta Q/\Delta t \sim 14$ m^3 s^{-2}) that cut a 2-km-long, 0.5-km-wide ice canyon ($\approx 1.5 \times 10^8$ m^3 ice depletion) [*Tómasson*, 1996] (Figure 11a). *Richardson* [1968] remarked that a 90-m-deep, 270-m-wide canyon extended into the terminus of Kautz Glacier (Washington State) following a jökulhlaup in 1947. *Fleisher et al.* [1998] noted the development of a flood-induced ice canyon during surges of Bering Glacier, Alaska, in 1966–1967 and 1993–1994 (Figure 11b). *Russell et al.* [2001] described a twin-chambered ice canyon, corresponding to $\sim 5.7 \times 10^6$ m^3 of ice loss, which formed during the late rising stage of the 1996 jökulhlaup (Figure 11c). The four previous examples share commonality because a pressurized, supraglacial release of floodwater occurred at each site before canyon formation [*Tómasson*, 1996; *Richardson*, 1968; *Roberts et al.*, 2000; P. J. Fleisher, personal communication, 2000]. Furthermore, for observed jökulhlaups of equivalent or higher magnitude the same type of canyon has formed at the same location [e.g., *Fleisher et al.*, 1998], implying repeated imposition of axes of preferential drainage.

4.5. Fluvial Erosion of Subglacial Sediments

[50] Scouring of unconsolidated sediment from beneath Skeiðarárjökull during the 1996 jökulhlaup is an outstanding example of the sediment scavenging potential of subglacial floodwater. Because of the residence time of volcanogenic floodwater in Grímsvötn (section 3.1), no pyroclastic fragments from the 1996 eruption were detected in floodwaters from Skeiðarárjökull [*Maria et al.*, 2000]. Hence bed load and suspended sediment load came from fluvial erosion of older volcaniclastic material and overridden glaciofluvial sediments in storage along the subglacial flood route [*Stefánsdóttir et al.*, 1999]. *Snorrason et al.* [2002] inferred that 1.8×10^{11} kg of suspended sediment load was carried in rivers draining floodwater from Skeiðarárjökull. From satellite radar interferograms, *Smith et al.* [2000] estimated 7.3×10^7 m^3 of glacier-proximal bed load deposition ($\equiv 1.3 \times 10^{11}$ kg, assuming dry, unconsolidated sand with a density of 1750 kg m^{-3}) during the 1996 jökulhlaup. Compositely, these values give a total sediment yield of $\geq 3.1 \times 10^{11}$ kg ($\approx 1.8 \times 10^8$ m^3) and a mean flux of 1.8×10^6 kg s^{-1} over the 47-hour duration of the jökulhlaup. Averaging 1.8×10^8 m^3 of sediment over the putative area of glacier bed impacted by floodwater ($\sim 5.8 \times 10^8$ m^2, section 3.3), implies 0.3 m (1.8×10^{-3} mm s^{-1}) of subglacial erosion. Remarkably, the 1996 jökulhlaup imposed an effective erosion rate comparable to the minimum annual denudation rate (0.3 m yr^{-1}) reported for the heavily glaciated region of southeast Alaska [*Hunter et al.*, 1996]. Clearly, intensive subglacial erosion occurs when turbulent bursts of floodwater scour the glacier bed.

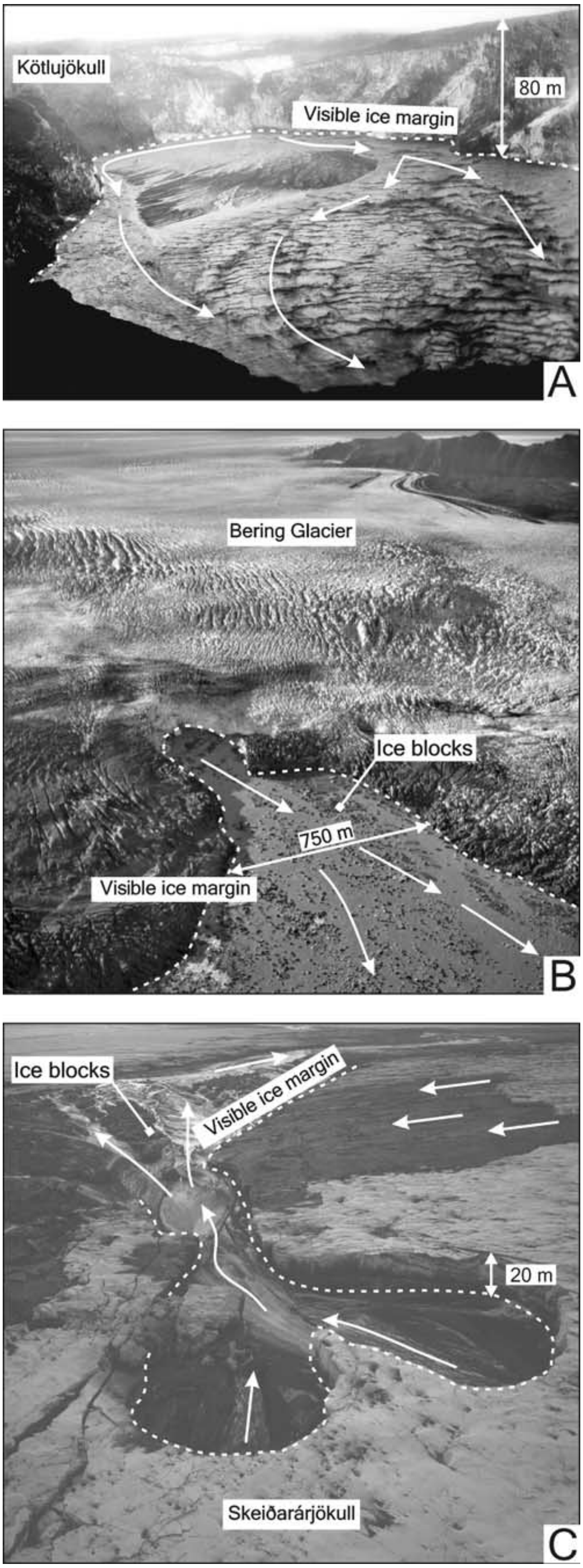

Figure 11. Giant ice-walled canyons at (a) Kötlujökull, Iceland; (b) Bering Glacier, Alaska; and (c) Skeiðarárjökull, Iceland. Single arrows illustrate primary directions of floodwater flow. Photographers: K. Guðmundsson (Figure 11a), November 1918 (restored photograph courtesy of R. T. Sigurðsson); A. Post (Figure 11b), U.S. Geological Survey, aerial photograph 663-28, September 1966 (photograph courtesy of P. J. Fleisher); and R. T. Sigurðsson (Figure 11c), November 1996.

4.6. Postflood Response of Subglacial Drainage System

[51] Major subglacial disturbances result from the passage of a high-magnitude jökulhlaup. Visible aftereffects include (1) development of elongated, trough-like sags in the ice surface [*Sturm and Benson*, 1985]; (2) postflood abandonment of preflood outlets [*Larson*, 2000]; (3) occupation of former floodwater outlets by melt-dominated efflux [*Thórarinsson*, 1974; *Röthlisberger and Lang*, 1987]; and (4) periods of low, nonvarying Q [*Anderson et al.*, 2003]. Intraglacial and ice-marginal ice rubble can create hydraulic constrictions capable of briefly impounding waning stage Q [*Paige*, 1955]. Temporary floodwater storage during subglacial volcanism can instigate repeated, progressively smaller, jökulhlaups [*Thouret et al.*, 1995; *Tómasson*, 1996]. Linearly rising jökulhlaups can exert lasting mechanical damage at the flood source. For example, the Grímsvötn ice dam (section 3.1) was damaged severely during the 1996 jökulhlaup [*Björnsson et al.*, 2001], resulting in nearly continual leakage of Grímsvötn meltwater until August 2000. (F. Pálsson, personal communication, 2003)

[52] After the 1996 jökulhlaup a slightly sinuous, 0.5-km-wide trench extended 10 km up glacier obliquely from the head of the ice-walled canyon (Figure 11c); floodwater thermodynamics (section 3.2) and sediment evacuation rates (section 4.5) during the 1996 jökulhlaup imply that this trench delineates part of a *tunnel valley* [cf. *Shaw*, 1996; *Piotrowski*, 1997]. Late Pleistocene tunnel valley genesis is a contentious topic [see *Ó Cofaigh*, 1996], and a jökulhlaup hypothesis, despite its questionable physics, has survived for decades [*Shaw*, 2002]. The 1996 jökulhlaup provides both conceptual impetus and empirical data for a fresh appraisal of tunnel valley genesis by floodwater. Moreover, extraordinary floods reviewed here dispel the view that sudden, voluminous meltwater release is normally nonerosive and therefore incapable of gouging subglacial sediments [cf. *Ó Cofaigh*, 1996].

5. EXTRAORDINARY JÖKULHLAUPS AND HEINRICH EVENTS?

[53] Sediment cores from the North Atlantic Ocean reveal a series of six layers that are rich in detrital carbonates and dropstones over distances of 10^2 km [see *Hemming*, 2004, and references therein]. *Heinrich* [1988] was the first to postulate that each of these rapidly deposited sediment

pulses originated from the melting of carbonate-bearing icebergs, cast adrift in huge quantities from the Laurentide Ice Sheet. Heinrich events therefore symbolize a sudden marine transfer of copious, spatially extensive, ice-rafted debris (IRD) from the Laurentide Ice Sheet because of cataclysmic, cyclic episodes of iceberg disgorgement. Despite overriding progress on the sedimentology and mineralogical provenance of North Atlantic IRD [*Andrews and Maclean*, 2003], there is a scarcity of understanding about the glaciological processes capable of embedding of the order of 100 km^3 of subglacially sourced debris synchronously within a $\sim 2 \times 10^6$ km^2 sector of the Laurentide Ice Sheet [*Clarke et al.*, 1999]. Moreover, little is known about the glaciological trigger responsible for a cascading release of ice-packaged debris from the grounding line of the Laurentide Ice Sheet. Hence a remarkable paradox remains: How do massive amounts of basal debris become embedded within a significant fraction of the ice column, thus poised for transport across an extensive ice shelf, and concentrated to the extent that IRD is traceable over the floor of the North Atlantic [*Clarke et al.*, 1999; *Andrews and Maclean*, 2003; *Hemming*, 2004]?

[54] In my view the glaciological enigma of Heinrich events can be resolved by a better geophysical understanding of debris entrainment processes confined to extraordinary jökulhlaups. In sections 4.1 and 4.2 I explained that the most violent, linearly rising jökulhlaups engender a subglacial response capable of projecting turbid floodwater to high elevations in the ice column, the corollary being englacial sedimentation due to hydraulic supercooling. Such events provide the necessary kinetics required to shatter a large area of the Laurentide Ice Sheet, while simultaneously ramming and hydraulically freezing comminuted sediment high into englacial hydrofractures. As an example I estimate from known hydrofracture dimensions [*Roberts et al.*, 2000] that ≥ 0.03 km^3 of glaciofluvial sediment were deposited within Skeiðarárjökull during the 1996 jökulhlaup.

[55] A jökulhlaup model for Heinrich events exists [*Johnson and Lauritzen*, 1995], but it is seen by some researchers as inadmissible. In any case, the prevailing view is that Laurentide ice-dammed lakes dumped floodwater episodically to the North Atlantic during glacial time [*Shoemaker*, 1992b; *Barber et al.*, 1999; *Teller et al.*, 2002; *Broecker*, 2003; *Clarke et al.*, 2004]. A jökulhlaup lasting 6 months with a Q_{MAX} of $\sim 5 \times 10^6$ m^3 s^{-1} [*Clarke et al.*, 2004] could prime the Laurentide Ice Sheet with englacial debris [*Roberts et al.*, 2002], force irreversible retreat of the grounding line [*Shoemaker*, 1992b], and discharge copious icebergs loaded with jökulhlaup sediments (Figure 12). Qualitatively, the jökulhlaup hypothesis helps to reconcile the spatial variance of North Atlantic IRD [*Hemming*, 2004]; that is, (1) a fluviomarine plume of floodwater causes glacier-proximal erosion at the ocean floor, (2) hyperpycnal flows transform to turbidity currents, and (3) far-field sedimentation arises from the subaqueous melting of debris-packed hydrofractures, preserved in slab-shaped fragments of ice (Figure 12). To help redress the lack of understanding about the glaciological causes of Heinrich events and to confront the cursory hypothesis outlined here, I advocate that greater interdisciplinary discussions take place between jökulhlaup and Quaternary scientists.

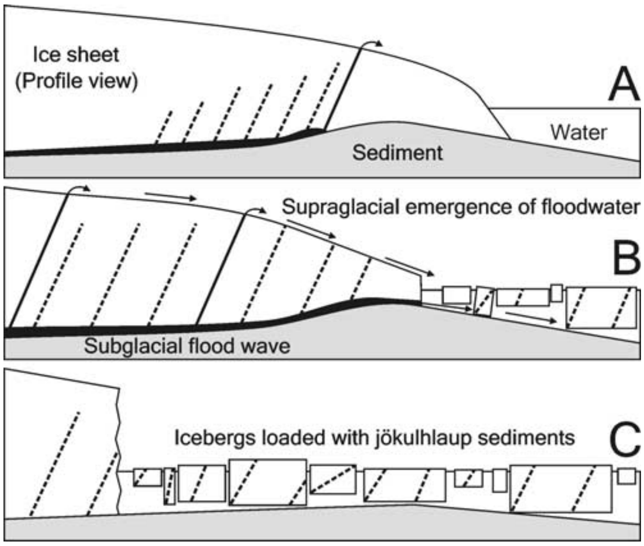

Figure 12. Conceptualization of the glaciological processes behind Heinrich events. (a) Subglacial propagation of a flood wave and concomitant debris entrainment within the Laurentide Ice Sheet. (b) Retreat of the grounding line by hydromechanical erosion and initial release of debris-freighted ice to the marine environment. (c) Flood wave having passed through the ice sheet and flotillas of debris-freighted ice adrift in the North Atlantic.

6. CONCLUDING REMARKS

[56] The purpose of this review was threefold: (1) to reassess how floodwater flows through glaciers, (2) to develop a qualitative view of subglacial processes during jökulhlaups, and (3) to suggest research priorities for future jökulhlaup studies. Regardless of the generic applicability of my linearly rising postulate for jökulhlaups, the review findings (Figure 13) make it plain that some jökulhlaups cause pervasive hydraulic perturbations wholly inconsistent with the most liberal theoretical views of intraglacial hydrodynamics. Monotonically accelerating discharge >0.1 m^3 s^{-2} from conduit outlets signifies nonlinear power law behaviors unrepresentative of traditional linear theories of jökulhlaup hydrodynamics. In most instances reported here, disparity between observation and theory was significant, thus suggesting that rapid, linearly rising jökulhlaups make up a class of phenomena strikingly different from those embodied in the path-breaking work of *Nye* [1976], *Spring and Hutter* [1981, 1982], and *Clarke* [1982].

[57] Glaciological studies of jökulhlaups have flourished on the utility of steady state hydrodynamic theory; in light of observations here, such unity now seems contradictory. New mathematical models must therefore be more circumspect by anticipating a potential richness of nonlinear hydraulic response within intraglacial drainage (Figure 13). Once defined mathematically, these coupled hydraulic responses will provide outstanding theoretical insight into

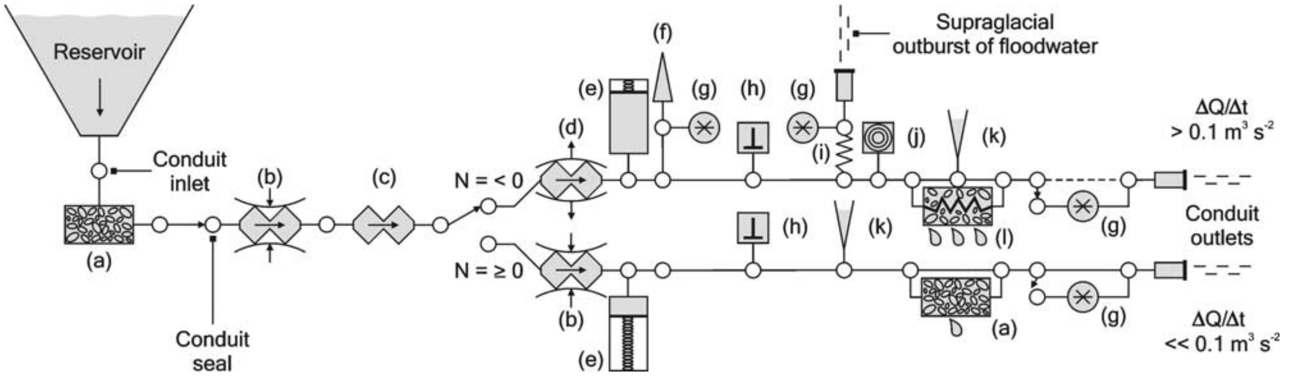

Figure 13. Qualitative model of the physical processes governing floodwater flow through glaciers. The model follows the logic and morphological representations of *Clarke* [1996]. The symbolic elements of the subglacial circuit are labeled as follows: a, Darcian filter; b, R-channel resistor; c, enthalpy extractor; d, distension wave generator; e, glaciodynamic closed storage; f, glaciostatic closed storage; g, frazil ice generator; h, oscillatory ice valve; i, hydrofracture offshoot; j, pump for hydraulic transience; k, supraglacial or subaerial hydrostatic storage; and l, subterranean hydrofracturing. Depending on the rate of $\Delta Q/\Delta t$ at the conduit seal and floodwater temperature at point c, floodwater can exhibit either linearly rising (upper circuit) or exponentially rising (lower circuit) hydrodynamics, as defined by effective pressure (N). Symbolic elements are located arbitrarily along the flood tract. Note that processes g to l could occur in different sequences and that most processes could occur individually.

the dynamic, hydromechanical reactions of glacier ice to pervasive hydraulic transience; such a situation can only act as a stepping stone toward more accurate predictions of jökulhlaup timing and intensity.

[58] Deficiencies in our theoretical understanding of jökulhlaups are due mainly to a systematic lack of pertinent, high-quality observational data. Apart from the work of *Clarke* [1982], *Björnsson* [1992], and *Anderson et al.* [2003], jökulhlaup scientists seldom make observations over the entire flood tract. Field studies that treat jökulhlaups as a system are likely to make significant contributions to jökulhlaup science; additionally, controlled laboratory experiments can facilitate excellent mechanistic understanding [e.g., *Catania and Paola*, 2001]. Instead of overly relying on calibrated parameters and adjustable coefficients, theoretical models should be recast with crucial glaciological data relevant to the jökulhlaup system in question. The deployment of water temperature profilers, hydrometric gauging equipment, borehole pressure transducers, and geodetic equipment during some jökulhlaups has allowed excellent insight into jökulhlaup physics, but novel, precursory applications of such instruments are necessary.

7. FUTURE RESEARCH PRIORITIES

[59] Future theoretical models of jökulhlaup hydrodynamics need to include the following realizations: (1) The location of the conduit inlet and the conduit seal need not be contiguous in time and space. (For linearly rising jökulhlaups the primary flow constriction exists at the tip of the propagating flood wave.) (2) In-phase variations in dynamic P_w and Q are probable at the outset of jökulhlaups, thus engendering hydraulic impulses in subglacial drainage. (3) Subglacial flood waves impart massively heterogeneous variations in hydrodynamic stress to the glacier base, resulting in the mechanical creation of englacial and supraglacial circuits and stores for overpressurized floodwater. (4) Hydraulically driven mechanisms are a primary method for the initial excavation and spatial expansion of a subglacial flood tract over unconsolidated sediments. (5) Wide, shallow subglacial conduits are spatially prevalent during jökulhlaups over unconsolidated sediments. (This realization will facilitate Q derivations based on accurate channel cross sections and realistic values of relative conduit roughness.) (6) Dynamic, nonlinear spatial and temporal changes in P_w force structural adjustments to the geometry and configuration of intraglacial drainage (on a timescale of minutes to hours) during jökulhlaups of any type, magnitude, or scale. (For linearly rising jökulhlaups this realization is intrinsic and applicable to the entire water catchment of the affected glacier.) (7) The spatial pattern of ϕ is usually highly variable over the subglacial flood tract; consequently, ϕ should be determined dynamically from composite measurements of the slope of the ice surface and glacier bed. (This realization will allow the width of the subglacial flood path to be determined more precisely, thus preventing unmodified use of the Gauckler-Manning equation as an estimate for the relationship between Q and ϕ.) (8) Thermodynamic empiricisms extracted from engineering literature are a poor surrogate for the unique heat transfer problems posed by jökulhlaups. (9) Pulses of sediment-laden floodwater induce thermodynamically significant rates of ice abrasion and cavitation owing to the production of ice shards. (This realization has implications for the volumetric contribution of ice melt attributed to viscous dissipation of heat.) (10) Freshwater density assumptions are unrepresentative of jökulhlaups. (Subglacial floodwater can have an unequal mass distribution with respect to volume.) (11) When active, hydraulic supercooling causes drastic changes to the hydrodynamic and thermodynamic properties of jökulhlaups. (This realization is important because accretionary freezing of floodwater heightens the constriction rate of ice-roofed con-

duits, and frazil ice alters floodwater rheology.) (12) Finally, exponentially rising and linearly rising jökulhlaups are hydrodynamically distinctive and theoretically separable, with the defining characteristics of linearly rising jökulhlaups arising because of the explanatory shortcomings of contemporary theory.

GLOSSARY

Effective pressure: Difference between hydrostatic and glaciostatic pressure, as defined by $N = P_i - P_w$. When P_w is zero, N is asymptote; conversely, if P_w equals P_i, then N is zero.

Englacial: Zone above the glacier base but below the glacier surface, i.e., within the body of a glacier.

Glacier bed: Spatial interface that supports a glacier and facilities glacier sliding due to the presence of meltwater. Bed composition can be both lithologically and sedimentologically diverse as well as spatially and temporally heterogeneous [see *Clarke*, 2005].

Glacier terminus: Lowermost front of a glacier. Variant phases include glacier toe, glacier tongue, and glacier snout.

Glaciohydraulic: Generic term used to encompass hydrodynamic processes operating in subglacial, englacial, and supraglacial zones.

Hydraulic supercooling: Occurs when subglacial meltwater is out of thermodynamic balance with local glaciostatic stress. As subglacial meltwater moves horizontally toward areas of lower glaciostatic stress (and thus comparatively warmer ice), viscous dissipation of latent and frictional heat keeps meltwater in thermodynamic balance with ambient glaciostatic stress [*Shreve*, 1985; *Shoemaker*, 1987; *Alley et al.*, 1998]. However, if the total hydraulic potential of meltwater increases rapidly, then a deficit in heat production can occur, allowing meltwater temperature to fall (i.e., supercool) below the ambient, pressure-determined melting point. It is convenient to visualize hydraulic supercooling as the rapid displacement of meltwater isotherms to areas of higher ice temperature.

Intraglacial: Pertaining to both subglacial and englacial environments.

Outlet: A discrete vent for meltwater formed at the glacier terminus in zones of low hydraulic potential.

Pressure-determined melting point: Temperature at which meltwater freezes at a particular glaciostatic pressure. The pressure dependence of the melting point (θ_i) follows the linearized Clausius-Clapeyron equation $\theta_i = -C_t P_i$, where C_t is proportional to -7.5×10^{-5} °C kPa^{-1}. Note that solutes in meltwater can affect the pressure dependence of the melting point.

Proglacial: Subaerial zone next to a glacier terminus that contains landforms, fluvial processes, and sedimentary assemblages exclusive to glacial environments. Synonyms include outwash plain and sandur. (The latter phase is Icelandic, literally meaning sand; plural is sandar.)

Subglacial: Spatial interface between the glacier bed and the glacier base [see *Clarke*, 2005].

Supraglacial: Pertaining to the glacier surface.

Tunnel valley: A large, long, and sinuous trench formed partly in the base and bed of a temperate glacier. Remnant tunnel valleys often cut into bedrock and unconsolidated sediments. Active tunnel valleys are observable as ice surface depressions that extend up glacier from high-capacity outlets.

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