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https://doi.org/10.1029/2007GL031426

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Climatic control on the peak discharge of glacier outburst floods

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Received 24 July 2007; revised 21 September 2007; accepted 8 October 2007; published 7 November 2007.

1. Introduction

Jökulhlaups are floods with high discharge (up to 10^7 m³ s⁻¹ today) and days-to-weeks duration caused by unstable water release by ice-dammed lakes [Björnsson, 2004; Haeberli, 1983; Post and Mayo, 1971; Roberts, 2005]. More than a hundred contemporary lakes are known to have produced jökulhlaups, and it is also thought that in the past mega-jökulhlaups from ice-dammed lakes worldwide, regional warming will also promote higher-impact jökulhlaups by raising the likelihood of warm weather during their occurrence, unless other factors reduce lake volumes at flood initiation to outweigh this effect. Citation: Ng, F., S. Liu, B. Mavlyudov, and Y. Wang (2007), Climatic control on the peak discharge of glacier outburst floods, Geophys. Res. Lett., 34, L21503, doi:10.1029/2007GL031426.

2. Theoretical Controls on Flood Peak Discharge, Q_max

In Nye-type models [Nye, 1976; Spring and Hutter, 1981; Clarke, 1982, 2003; Ng, 1998; Evatt et al., 2006], a single subglacial tunnel is envisaged as carrying the flood discharge Q(t) from the lake. Here t is time. Tunnel enlargement (ice melting by frictional heat of water flow) fuels rapidly-growing Q in the rising flood stage by positive feedback, but drawdown of the lake-water level and pressure offsets this process simultaneously by promoting tunnel closure (viscous ice deformation). Although simulations of Q(t) can be tuned to fit an observed hydrograph, genuine hydrograph forecasting is difficult because the non-linear flood physics make Q(t) sensitive to the initial conditions [Ng and Björnsson, 2003], which are usually uncertain.

Nevertheless, Nye’s physics indicate that the peak discharge should increase with (1) the initial lake volume, V_i, (2) the lake-water temperature, T_i, and (3) the rate of meltwater supply to the lake, Q_i. Controls (1) and (2) result from the corresponding increase in potential and thermal energy driving tunnel enlargement, and have been studied before [Spring and Hutter, 1981; Clarke, 1982]. Control (3) is more subtle. It recognises that besides determining how fast the lake refills between floods, the supply rate Q_i (which is generally weather-dependent) can also modulate Q_max during a flood. An increased meltwater supply slows the fall of lake level and pressure, thus slowing tunnel closure and causing faster growth of Q, as the lake drains,
producing a larger flood peak. (This effect is greatest early in the flood when $Q \sim Q_i$.) In this paper we refer to controls (2) and (3) collectively as ‘meltwater modulation’, because other processes link $T_i$ to $Q_i$ at Merzbacher Lake (Section 3.1). Our hypothesis is that by incorporating meteorological factors, these controls can cause significant variability in $Q_{\text{max}}$, explaining why a simple $(Q_{\text{max}}, V_i)$ relation due to control (1) is not observed.

3. Weather-Induced Modulation of Merzbacher Jökulhlaups

3.1. Model and Data Source

[6] To test this idea, one could reconstruct each flood to match its observed hydrograph and compare $Q_i$ required for this with in-situ measurements. But since neither hydrographs obtained directly from Merzbacher Lake, nor sufficient record of the controls, are available, here we use $(Q_{\text{max}}, V_i)$ data instead to constrain our reconstructions and look for expected correlations between $Q_i$ and hydro-meteorological data as evidence of meltwater modulation. For simplicity we assume $Q_i$ to be constant during each flood (thus ignoring diurnal variations).

[7] Our interest in Merzbacher Lake (Figure 1a) stems from its long flood series and impact on cross-border water issues. Named after the scientist who discovered it over a century ago [Merzbacher, 1905], the lake is roughly 80 m deep and 4 km$^2$ in area when full. Although it lies in the Kyrgyz Republic, its jökulhlaups debouch almost annually onto China’s Silk Road. River gauging records from Xiehela hydrological station identify > 40 outbursts between 1958 and 2002, our study period [Liu, 1992; Liu et al., 1998; Shen et al., 2007]. Thirty-nine of these floods are analysed below.

[8] To find $Q_{\text{max}}$ and $V_i$ for each flood, we first subtracted from the Xiehela discharge its base flow $Q_b$ (Figure 1b) to find a ‘separated’ flood hydrograph. We estimated $Q_b$ by graphical interpolation, or, where the discharge is markedly diurnal (melt-dominated), by using a correlation between diurnal amplitude and smoothed base flow ($Q_b$ minus diurnal) before and after the flood to construct a diurnally-varying $Q_b$ that minimises diurnal changes on the flood discharge after subtraction. The separated Xiehela hydrograph (Figure 1c) is an altered version of $Q(t)$ due to river hydraulics, but water conservation equates its area to the flood volume $V_i$. We also equated its peak value to $Q_{\text{max}}$, as the high river gradient (averaging 7.5 × 10$^{-5}$) and slow rise-time of the floods suggest that floodwaves propagate down-river with negligible diffusion [e.g., Ponce, 1989]. Figure 1d shows the resulting $(Q_{\text{max}}, V_i)$ dataset. Reliable data were obtained for 18 floods, whereas difficult base flow estimation led to more uncertain data for the other 21 floods.

[9] Each flood is reconstructed by solving a simplified Nye model for discharge $Q(t)$ and lake level $h(t)$. As detailed in Appendix A, our method involves numerically integrating a pair of differential equations for these variables to satisfy three conditions. Because the lake is observed to empty completely in the floods [Mavludyov, 1997], the model flood must end with $h = 0$, and begin at a lake level $h(0)$ consistent with the (volume) requirement $V_i = V_b + Q_i \times$ flood duration; it must also peak at $Q_{\text{max}}$. These conditions constrain the solutions uniquely and allow us to deduce the meltwater supply $Q_i$. In this inversion, $Q_i$ (defined for the flood duration) is relatively insensitive to errors in the input data $Q_{\text{max}}$ and $V_i$ because the forward-time model is sensitive. Also, the initial lake level $h(0)$, a variable computed as part of the solution, represents control (1) and
reflects the flood-initiation threshold selected by dam breach physics.

[10] Our inversions for Merzbacher Lake include a submodel for lake temperature \( T_L \) (control 2) because \( T_L \)-data are scarce. Meltwater supplied to the lake by rivers (at several °C during summer [Dikikh and Kuzmichenok, 2003; Maklyudov, 1997]) is typically warmer than meltwater from the ice-dam and icebergs on the lake (≈0°C), so the former meltwater warms the lake by an amount dependent on \( Q_3 \). We assume the proportionality \( T_L = kQ_3 \) to parameterise this effect, without resolving the lake’s internal thermodynamics. (\( T_L \) is taken to be the mean lake temperature in °C.) Limited warming of lake-surface water by the atmosphere is also possible, but is implicitly described by this parameterisation. A plausible estimate, \( k = 0.02 \), is suggested by the observation that \( T_L \) ranges from 0 to ~2.5°C during summer [Dikikh and Kuzmichenok, 2003] while \( Q_3 \) ranges from ~10 to ~100 m³ s⁻¹. We include sensitivity experiments with \( k \) in later calculations.

3.2. Results

[11] Using \( k = 0.02 \), we compare the modelled meltwater supply rate \( Q_3 \) with two proxy variables that should covary with the actual supply rate. Figure 2 shows these comparisons. Meltwater modulation of the flood peaks is evidenced by tight correlations for the 18 floods with reliable \( (Q_{\text{max}}, V_t) \) data; results from uncertain data do not spoil these trends. In Figure 2a, \( \theta_0 \) is a mean meteorological temperature for the early part of each flood, compiled from \( \theta_{\text{NCEP}} \), the NCEP-NCAR Reanalysis daily mean temperature [Kalnay et al., 1996; Kistler et al., 2001] interpolated to the lake location. We use \( \theta_{\text{NCEP}} \) time series (1948-present) instead of instrumental temperatures from Tyan-Shan Station (see Figure 1a), because these records are well correlated (with \( r^2 = 0.89 \)) where they overlap but the latter becomes patchy before the 1970s. In Figure 2b, the comparison is made with \( \overline{Q_b} \), the mean Xiehela baseflow of each flood, to target runoff factors (e.g., catchment snow/ice cover) besides temperature. The covariation of \( \overline{Q_b} \) and \( Q_3 \) is expected as both are related to weather, although \( \overline{Q_b} \) represents the entire Sary-Djaz basin, ~40 times larger than the lake catchment. These correlations support our hypothesis that air temperature controls the meltwater supply, which in turn modulates flood peak discharge. Our calculations for Merzbacher Lake show that differences in \( Q_3 \) can cause differences in \( Q_{\text{max}} \) that are ~20 times greater.

[12] We have validated these conclusions through sensitivity experiments, by repeating all our flood inversions with different combinations of \( k \) and \( n' \) (Manning roughness see Appendix A) in the conservative ranges \( k = 0.005 \) to 0.05 and \( n' = 0.04 \) to 0.12 m⁻¹ s⁻¹. The upper bound of \( k \) derives from the fact that \( T_L = kQ_3 \) predicts unreasonably warm lake-water if \( Q_3 \approx 100 \) m³ s⁻¹ and \( k > 0.05 \). The lower bound is inferred from our model, where reducing \( k \) raises \( Q_3 \) by weakening control (2) relative to control (3), and where \( Q_3 \) becomes unrealistically large (>200 m³ s⁻¹) if \( k < 0.005 \). In each experiment, neither \( k \) nor \( n' \) is changed between floods. These experiments support the significance of our correlations because \( r^2 \) in Figures 2a and 2b never dropped below 0.69 and 0.52, respectively.

4. Discussion

[13] These findings clarify why repeated jökulhlaups from an ice-dammed lake can form complicated series of \( Q_{\text{max}} \) and \( V_t \). After a flood ends, the lake’s (variable) meltwater supply and its (uncertain) flood-initiation threshold together set its refilling period, and the timing of the next flood. In turn, the weather coinciding with this next flood is now understood to control its \( Q_{\text{max}} \) and \( V_t \) (and the lake volume at the start of the next filling cycle, if the lake does not drain completely). The sizes of successive floods are thus interlinked, dependent on their precise timing relative to weather fluctuations.

[14] This has implications for how predictable is the resulting pattern of floods. Since the flood-recurrence timescale at many lakes (~annual at Merzbacher Lake) depends on the melt cycle tied to the annual temperature cycle, climate change over the long term will influence the timing of the outbursts as well as their \( Q_{\text{max}} \) and \( V_t \). However, on shorter timescales, each flood outcome will also be sensitive to ‘noise’ in weather and in the flood-initiation threshold,
implying considerable difficulty in predicting the outburst pattern beyond a flood cycle—this is especially because dam-breath physics are poorly understood. Nevertheless, accurate lake-level monitoring in the early hours of a flood could yield sufficient data for Nye-type models to forecast its subsequent evolution.

Meanwhile, if \( Q_{\text{max}} \) and \( V_t \) carry climatic information at flood events, then it may be possible to use their series to make empirical projection of flood characteristics for the future. Earlier studies [Liu et al., 1998; Shen et al., 2007] have associated rising trends in the peak discharge and volume of the Merzbacher floods with an observed 0.01–0.02°C/yr warming trend [Liu et al., 1998; Aizen et al., 1997, Shi et al., 2002] in the Tian Shan over the second half of the twentieth century, and they caution that continued warming may worsen flood impact. But these studies [Liu et al., 1998; Shen et al., 2007] derived \( Q_{\text{max}} \) and \( V_t \) from the Xiehela discharged record without subtracting its base flow, which itself increased under warming (Figure 3a). Removing this bias shows that the rising trend of \( Q_{\text{max}} \) is only 6.2 m³ s⁻¹/yr rather than 19.0 m³ s⁻¹/yr for \( Q_{pk} \) as inferred previously, while \( V_t \) decreases (Figures 3a and 3b). Because falling flood volumes did not result as long-term reduction in \( Q_{\text{max}} \), meltwater modulation must have had a dominant control on \( Q_{\text{max}} \) as the weather coinciding with the floods warmed (Figure 3c); the mean flood duration must also have shortened.

This analysis raises concern for jökulhlaup risks globally, as it suggests that a similar mechanism may operate in other lakes experiencing regional warming (which increases the chance of floods during warm weather if their distribution within the year is statistically unchanged). Specifically, our model predicts that more hazardous floods will occur unless regional warming is accompanied by a shift of outburst timing to cooler months, or by a fast reduction in initial lake volumes. Glacier thinning could shift of outburst timing to cooler months, or by a fast refilling might lower initial lake levels, as seems to be the case at Merzbacher Lake, or increase them, as suggested by a recent theory [Fowler, 1999]. Proposals for hydroelectric dam construction downstream from Merzbacher Lake (D. Mammatkanov, Institute of Water Problems and Hydropower, Kyrgyz Republic, 2007) should provide new impetus for study from both sides of the border.

Appendix A: Mathematical Model

[17] The surface topography of South Inylchek Glacier’s 15 km-long lower tongue is known, but bed topography data are lacking except near Merzbacher Lake [Macheret et al., 1993; Mavlyudov, 1997]. Therefore, instead of using Nye’s [1976] original model, which incorporates spatial variation along the subglacial tunnel, we adopt a time-dependent-only model approximation [Clarke, 1982; Ng, 1998; Ng and Björnsson, 2003] where differential equations for tunnel evolution and lake-water conservation are, respectively:

\[
\frac{3}{4} \frac{dQ}{dt} = \frac{1}{\rho_i} \left( \frac{\Psi}{F_1} \right)^{1/3} \frac{Q^{1/3}}{m - K (\rho_W g)^n} Q (h_F - h)^n, \tag{A1}
\]

\[
A(h) \frac{dh}{dt} = Q_i - Q. \tag{A2}
\]

Besides variables already introduced, \( m \) is tunnel meltrate, \( A \) is lake area, \( \Psi = 290 \text{ Pa m}^{-1} \) is mean hydraulic gradient, and \( h_F = 109 \text{ m} \) is the estimated flotation lake level. Physical constants are ice density \( \rho_i \), water density \( \rho_W \), gravity \( g \), Glen’s exponent \( n = 3 \), and closure constant \( K = 5 \times 10^{-25} \text{ Pa}^{-3} \text{ s}^{-1} \) (for ice at 0°C). Our standard flood
inversion (Section 3 and Figure 2) assumes Manning roughness $n^2 = 0.075 \text{ m}^{1/3} \text{s}$ in the friction parameter $F_1 = (4\pi)^{2/3} \rho_\text{w} g h^{2/3}$ for cylindrical tunnel. Assuming a stable ice-dam front position [Mavlyudov, 1997], we specify lake hypsometry $A(h)$ from a map by Kuzmichenok [1984] based on aerial photogrammetry. In each numerical flood inversion, we define $t = 0$ when the lake achieves ‘highstand’ shortly after it begins to leak (flood initiation). An inner loop integrates equations (A1) and (A2) from discharge $Q(0) = Q_i$ and highstand lake level $h(0)$, where $Q_i$ is chosen by a shooting algorithm that matches the simulated peak discharge to $Q_{\text{max}}$. An outer loop then finds $h(0)$ iteratively (Newton-Raphson) to satisfy the condition $V_h = V_1 - (Q_i \times \text{simulated flood duration})$, where $V_h$ denotes the initial (highstand) lake volume.

Our equation for the meltwater (spatially averaged along the tunnel) is

$$mL = (1 - \alpha)Q\Psi + \alpha F_0 \frac{\Psi}{F_1} \frac{T_1}{Q_1} \frac{Q_1}{Q_2}^{3/2},$$

where $L$ is latent heat and $F_0 = 0.2055(2\rho_\text{w}/\eta\sqrt{\pi})^{0.8}$ is a heat transfer constant ($\eta$ and $\kappa$ are viscosity and thermal conductivity of water, respectively). Equation (A3) differs from Clarke’s [1982] equation (10) in the way in which it apportions potential energy ($1 \times Q^2\Psi$ on its right-hand side) and thermal energy (remaining terms on the right) for melting. Because water temperature $T$ equilibrates from $T_L$ at the lake towards a different temperature down the tunnel, $m$ is not what it would be if $T = T_L$ everywhere, so the coefficient $\alpha < 1$ (dependent on $Q$). By solving the corresponding pseudo-steady temperature problem we find

$$\alpha = 1 - e^{-\beta} \quad \text{where} \quad \beta = \frac{F_0 d_0}{\rho_\text{w} c_\text{w} \sqrt{Q}} \frac{\Psi}{F_1} \frac{3/20}{T_1}.$$

$I_0$ is the tunnel length (15 km) and $c_\text{w}$ is specific heat capacity of water.

[19] Acknowledgments. We thank Howard Conway, Andy Hodson, and Christian Schoof for comments, Richard Hindmarsh and Roger LeB. Hooke for their careful and constructive reviews, Shunde Wang for Kyrgyz meteorological data, Ian Harris for providing NCEP-NCAR Reanalysis data, and Valeriy Kuzmichenok for passing us his paper on lake hypsometry.

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