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Climatic control on the peak discharge of glacier outburst floods

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[1] Lakes impounded by natural ice dams occur in many glacier regions. Their sudden emptying along subglacial paths can unleash ~1 km^3 of floodwater, but predicting the peak discharge of these subglacial outburst floods ('jökulhlaups') is notoriously difficult. To study how environmental factors control jökulhlaup magnitude, we use thermo-mechanical modelling to interpret a 40-year flood record from Merzbacher Lake in the Tian Shan. We show that the mean air temperature during each flood modulates its peak discharge, by influencing both the rate of meltwater input to the lake as it drains, and the lake-water temperature. The flood devastation potential thus depends sensitively on weather, and this dependence explains how regional climatic warming drives the rising trend of peak discharges in our dataset. For other subaerial 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.

1. Introduction

[2] Jökulhlaups are floods with high discharge (up to 10^5 m^3 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 sheets delivered freshwater to the oceans, triggering climatic changes [*Clarke et al.*, 2003; *Evatt et al.*, 2006]. Yet the problem of forecasting the volume (V_t) and the peak discharge (Q_{max}) of these floods is unsolved. Although studies of data from multiple lakes [*Clague and Mathews*, 1973; *Walder and Costa*, 1996] show that these flood parameters follow a statistical power-law trend Q_{max} = constant × $V_t^{0.67}$, the physical basis of this relation is uncertain [*Ng and Björnsson*, 2003]. Moreover, floods from the same lake often differ markedly in size, with no consistent trend in their (Q_{max}, V_t) data [*Ng and Björnsson*, 2003]. Deciphering these variations is an important step towards jökulhlaup prediction, not least because it may be possible to predict an imminent flood's peak discharge from its available lake-water volume.

[3] Here we study this problem by examining the controls on flood evolution. Flood hydrographs have been simulated with *Nye*'s [1976] theory. We direct this theory to investigate how the lake volume at flood initiation conspires with flood mechanics to determine Q_{max}, paying attention to the possible effect of weather or climate on this process; thus we focus on marginal rather than subglacial lakes. Merzbacher Lake in Kyrgyzstan is used as case study. This lake empties completely in jökulhlaups, so each flood can be reconstructed using (Q_{max}, V_t) data as constraint. Specifically, our modelling recovers not only the flood hydrograph but also an estimate for the rate of water supply to the lake (Q_i) during the flood. For the cold regions being considered, this supply typically comes from snow- and ice-melt near the lake, and in some cases also precipitation. We then show that modelled Q_i-values correlate with weather conditions, implying that although the peak discharge Q_{max} depends on the initial lake volume, it is also controlled externally by weather through the lake's meltwater input. We use this result at the end of the paper to consider long-term change in flood behaviour.

2. Theoretical Controls on Flood Peak Discharge, Q_{max}

[4] 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 the 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 nonlinear flood physics make $Q(t)$ sensitive to the initial conditions [*Ng and Björnsson*, 2003], which are usually uncertain.

[5] Nevertheless, Nye's physics indicate that the peak discharge should increase with (1) the initial lake volume, V_h, (2) the lake-water temperature, T_L, 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,

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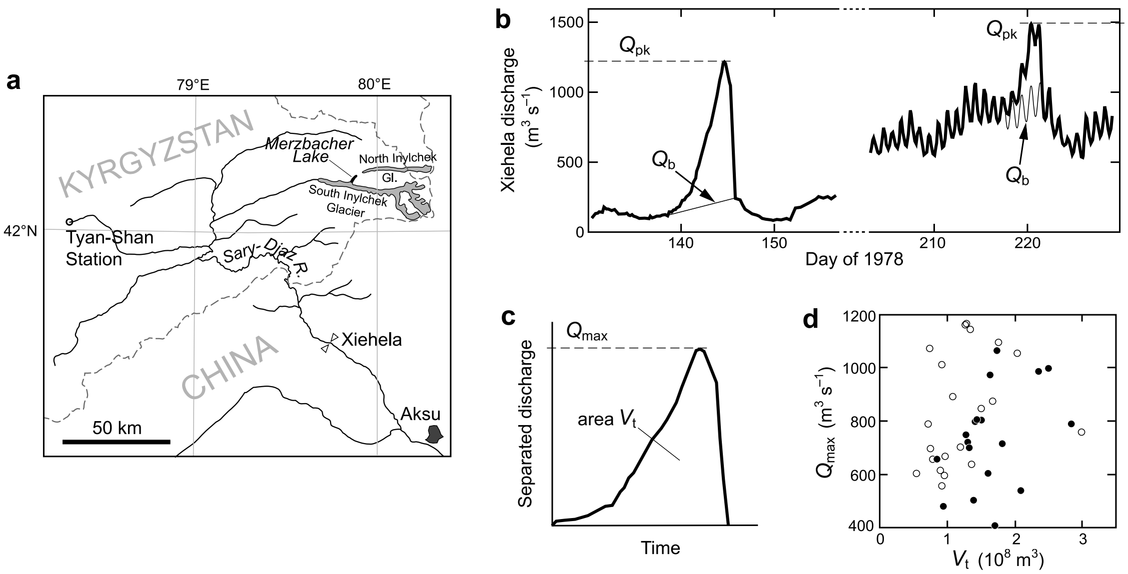

Figure 1. Anatomy of Merzbacher-Lake jökulhlaups. (a) Map of Central Tian Shan showing Merzbacher Lake (42°12′N, 79°50′E), South Inylchek Glacier (whose lowest 15 km dams the lake), Sary-Djaz River, Tyan-Shan weather station, Xiehela hydrological station, and approximate international borders. (b) Xiehela records of two 1978 floods and our definitions of Xiehela peak discharge Q_{pk} and base flow Q_b. (c) The 'separated' hydrograph, and definitions of flood peak discharge Q_{max} and flood volume V_t. (d) Q_{max} versus V_t for 39 floods from 1958 to 2002. Solid points derive from reliable hydrograph separation. Open points suffer uncertainty due to pronounced diurnal signals (>33% of flood signal) on the Xiehela discharge that make base flow estimation difficult (as exemplified by the second 1978 flood).

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_L 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_{max}, explaining why a simple (Q_{max}, V_t) 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_{max}, V_t) 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_{max} and V_t 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_t. We also equated its peak value to Q_{max}, as the high river gradient (averaging 7.5×10^{-3}) 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_{max}, V_t) 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 [*Mavlyudov*, 1997], the model flood must end with $h = 0$, and begin at a lake level $h(0)$ consistent with the (volume) requirement $V_t = V_h + Q_i \times$ flood duration; it must also peak at Q_{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_{max} and V_t 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

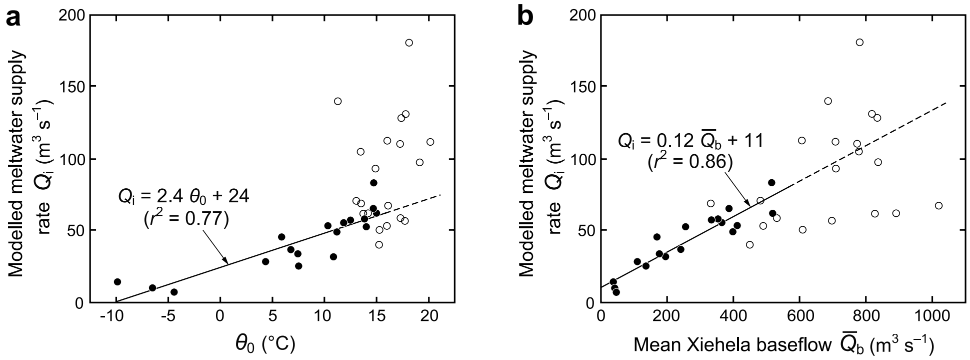

Figure 2. Correlation of modelled meltwater supply rate Q_i to Merzbacher Lake with (a) θ_0, the NCEP-NCAR Reanalysis daily surface temperature (θ_{NCEP}) averaged over the first-third duration of each flood, and with (b) Q_b averaged over each flood. θ_{NCEP}-variations track meteorological temperature variations near the lake, and Q_b represents background runoff in the Sary-Djaz River. Regression lines pertain to solid points (derived from reliable data in Figure 1d) and neglect open points (derived from data relating to uncertain base flow). Not surprisingly, the latter points are associated with warm weather and high base flow of summer. Results for the June-1963 flood are missing because its large volume implies lake levels above flotation, precluding application of Nye's model.

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; *Mavlyudov*, 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_i. We assume the proportionality $T_L = kQ_i$ 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_i ranges from ~10 to ~100 $m^3 s^{-1}$. 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_i 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_{max}, V_t) data; results from uncertain data do not spoil these trends. In Figure 2a, θ_0 is a mean meteorological temperature for the early part of each flood, compiled from θ_{NCEP}, the NCEP-NCAR Reanalysis daily mean temperature [*Kalnay et al.*, 1996; *Kistler et al.*, 2001] interpolated to the lake location. We use θ_{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 base flow of each flood, to target runoff factors (e.g., catchment snow/ice cover) besides temperature. The covariation of $\overline{Q}_b$ and Q_i 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_i can cause differences in Q_{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 $m^{-1/3}$ s. The upper bound of k derives from the fact that $T_L = kQ_i$ predicts unreasonably warm lake-water if $Q_i \approx 100\ m^3 s^{-1}$ and $k > 0.05$. The lower bound is inferred from our model, where reducing k raises Q_i by weakening control (2) relative to control (3), and where Q_i becomes unrealistically large (>200 $m^3 s^{-1}$) 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_{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_{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_{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,

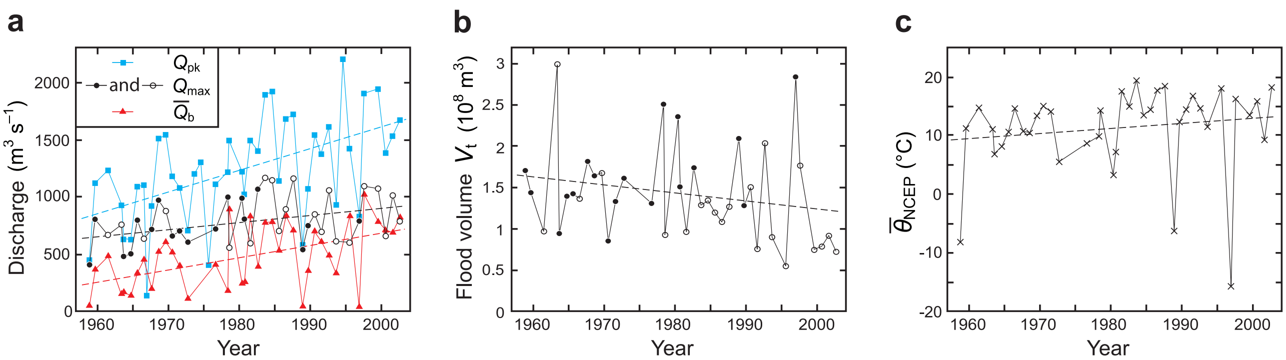

Figure 3. Merzbacher flood characteristics from 1958 to 2002. (a) Xiehela peak discharge Q_{pk}, mean Xiehela base flow $\overline{Q}_b$, and flood peak discharge Q_{max}. Linear trends for Q_{pk}, $\overline{Q}_b$, Q_{max} are +19.0, +10.5, +6.2 $m^3\ s^{-1}$/yr, respectively. (b) Flood volume V_t; linear trend $-1.0 \times 10^6\ m^3$/yr. (c) Mean θ_{NCEP} for each flood; linear trend +0.087°C/yr. In Figures 3a and 3b, solid and open circles indicate data reliability following convention in Figure 1. We do not have Xiehela record for floods in 1973, 1974, 1975, 1994, and December 1966, but Figure 3a includes the Q_{pk} data for these years from *Liu* [1992], *Liu et al.* [1998], and *Shen et al.* [2007].

implying considerable difficulty in predicting the outburst pattern beyond a flood cycle—this is especially because dam-breach 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.

[15] Meanwhile, if Q_{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_{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_{max} is only 6.2 $m^3\ s^{-1}$/yr rather than 19.0 $m^3\ s^{-1}$/yr (for Q_{pk}) as inferred previously, while V_t decreases (Figures 3a and 3b). Because falling flood volumes did not result in long-term reduction in Q_{max}, meltwater modulation must have had a dominant control on Q_{max} as the weather coinciding with the floods warmed (Figure 3c); the mean flood duration must also have shortened.

[16] 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 favour the latter scenario. For instance, *Mavlyudov* [1997] has associated falling initial levels at Merzbacher Lake with ice-dam thinning over the last century [*Glazyrin and Popov*, 1999], although (as shown above) this was too slow to overcome the melt-inflow effect. How the interplay of factors governs flood characteristics at other lakes awaits investigation. In cases where peak discharges show a falling trend, these discharges could have been smaller without meltwater modulation. Research must also elucidate how environmental change affects flood magnitude and timing through glacier and lake processes. Outstanding questions relate to dam-breach physics, including whether rapid lake-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. Mamatkanov, 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:

$$\frac{3}{4}\frac{dQ}{dt} = \frac{1}{\rho_I}\left(\frac{\Psi}{F_1}\right)^{3/8} Q^{1/4} m - K(\rho_W g)^n Q (h_F - h)^n, \qquad \text{(A1)}$$

$$A(h)\frac{dh}{dt} = Q_i - Q. \qquad \text{(A2)}$$

Besides variables already introduced, m is tunnel meltrate, A is lake area, Ψ = 290 Pa m^{-1} is mean hydraulic gradient, and h_F = 109 m is the estimated flotation lake level. Physical constants are ice density ρ_I, water density ρ_W, gravity g, Glen's exponent $n = 3$, and closure constant $K = 5 \times 10^{-25}$ Pa^{-3} s^{-1} (for ice at 0°C). Our standard flood inversion (Section 3 and Figure 2) assumes Manning roughness $n' = 0.075$ m$^{-1/3}$ s in the friction parameter $F_1 = (4\pi)^{2/3}\rho_W g n'^2$ 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_{max}. An outer loop then finds $h(0)$ iteratively (Newton-Raphson) to satisfy the condition $V_h = V_t - (Q_i \times$ simulated flood duration), where V_h denotes the initial (highstand) lake volume.

[18] Our equation for the meltrate (spatially averaged along the tunnel) is

$$mL = (1-\alpha)Q\Psi + \alpha F_0 \left(\frac{\Psi}{F_1}\right)^{3/20} T_L Q^{1/2}, \qquad \text{(A3)}$$

where L is latent heat and $F_0 = 0.205\kappa(2\rho_W/\eta\sqrt{\pi})^{0.8}$ is a heat transfer constant (η and κ 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 ($1 \times Q\Psi$ 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

$$\alpha = \frac{1-e^{-\beta}}{\beta} \quad where \quad \beta = \frac{F_0 l_0}{\rho_W c_W \sqrt{Q}}\left(\frac{\Psi}{F_1}\right)^{3/20}. \qquad \text{(A4)}$$

l_0 is the tunnel length (15 km) and c_W is specific heat capacity of water.

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