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# The effects of interannual climate variability on the moraine record

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## ABSTRACT

Valley glacier moraines are commonly used to infer mean climate conditions (mean annual precipitation and mean melt-season temperature). However, recent research has demonstrated that, even in steady climates, multi-decadal, kilometer-scale fluctuations in glacier length occur in response to stochastic, year-to-year variability in mass balance. When interpreting moraine sequences it is important to include the effect of interannual weather variability on glacier length; moraines record advances that are forced either by interannual variability or by a combination of climate change and interannual variability. We address this issue for the LGM glaciers of the Colorado Front Range, USA. Using a linear glacier model that allows thorough exploration of parameter uncertainties, supplemented by a shallow-ice flowline model, our analyses suggest that i) nested LGM moraine sequences are often confined to a range of glacier lengths that can be attributed to interannual climate variability; ii) mean glacier lengths are ~10–15% up-valley from maximum glacier lengths; and iii) individual LGM terminal

moraines were formed by a combination of a climate change and interannual variability forced advances.

## INTRODUCTION

Glacial to interglacial changes in the long-term averages of annual precipitation ($P$) and mean melt-season temperature ($T$) are sufficient, in many places, to force significant changes in glacier length (10s of km). But even in steady climates, year-to-year (interannual) variations in $P$ and $T$ have also been shown to force multi-decadal, kilometer-scale length fluctuations in valley glaciers, due solely to the random alignment of years of negative and positive mass balance (e.g. Reichert et al., 2002; Roe and O'Neal, 2009; DR.8). A steady climate implies constant long-term averages ($\equiv \overline{P}, \overline{T}$) and, importantly, constant standard deviations ($\equiv \sigma_P, \sigma_T$). All climates, steady or transient, include interannual variability. It is often incorrectly assumed that glaciers average away all interannual climate variability, and respond only to more persistent climate fluctuations. However, glaciers act as low-pass filters, producing multi-decadal (for the glaciers discussed in this paper) length fluctuations even if the climate forcing it is not correlated from year-to-year (white noise), which is almost always the case (Fig. 1B; Burke and Roe, 2013). Interannual variability is a result of the stochastic fluctuations of weather (climate noise) and the internal modes of variability in the climate system, such as the North Atlantic, the Pacific/North American, and the El Niño-Southern Oscillations (see DR.8 also). The amplitude of interannual variability varies with location and climate state, but it is always present. To constrain the values of past $\overline{T}$ and $\overline{P}$ using glacier moraines and to correctly interpret the moraine record we must understand the effects of year-to-year weather variability on glacier length and moraine emplacement.

To illustrate the problem, consider two glaciers: (a) a glacier subject to constant $\bar{T}$ and $\bar{P}$ forms a steady ice-surface profile that terminates at a steady, mean length, $\bar{L}$ (Fig. 1A); and (b) a glacier subject to a climate with the same $\bar{T}$ and $\bar{P}$ that also includes interannual variability. The glacier of case (b) will produce a terminus history that will fluctuate on multi-decadal scales (red noise) around the same $\bar{L}$ as glacier (a) (Fig. 1B; Fig. 2C). However, for glacier (b) $\bar{L}$ is a theoretical length with no expression in the landscape, and is the location around which the terminus fluctuates. If we assume that: (i) terminal moraines up to 40 m in height can be formed on timescales less than 50 years ii) moraines do not significantly impede subsequent advances, and iii) all moraines that are overrun by subsequent advances are removed, then the maximum excursion from $\bar{L}$ will form the furthest terminal moraines. Assumption (i) is supported by a compilation of 45 terminal moraine formation timescales and moraine heights, which shows that 25 m ice contact/dump moraines as well as 50 m push/glaciotectonic moraines form in less than 50 years (DR.3). Assumption (ii) is primarily a concern when latero-frontal dump moraines $> 50$ m in height are present; these are common in tectonically active regions such as the Himalaya, Andes, or Southern Alps (e.g. Benn and Evans, 1998; DR.3). Estimates of average climate (i.e., $\bar{P}, \bar{T}$) should be based on $\bar{L}$ rather than $L_{max}$ (Fig. 1B). Thus, we face the challenge of estimating the mean length $\bar{L}$ while knowing only the glacier geometry preserved by the maximum advance, $L_{max}$, and recognizing the substantial uncertainties in physical parameters.

We focus on the last glacial maximum (LGM) moraine record in the Colorado Front Range, USA to establish the effect of interannual variability on the moraine record. We use a shallow-ice-approximation flowline model (using standard techniques; DR.6) to confirm, in accordance with prior work in modern maritime, Alpine, and continental settings, that

interannual variability can force significant multi-decadal length fluctuations (e.g., Oerlemans, 2001). We primarily focus on constraining the effect of parameter uncertainty on the magnitude of length fluctuations forced by interannual variability using a linearized glacier model for the eleven-modeled glaciers.

## MODEL DESCRIPTION

### Linearized Model

The linear model considers glacier length variations, $L'$, as departures from a mean length $\bar{L}$ that are small enough that the equations are linear. Linear models use simplified glacier geometries to improve efficiency while still honoring the essence of glacier length change. They have been applied to a variety of glaciological problems (e.g., Johanneson et al., 1989; Oerlemans, 2001). Its use in this study is essential for efficient uncertainty analysis. Roe and O'Neal (2009) present a complete model description. If time is discretized into increments of Δt = 1 yr, then

$$L'_{t+\Delta t} = L'_t\left(1-\frac{\Delta t}{\tau}\right)+\beta P'_t+\alpha T'_t, \qquad (1)$$

The first term on the right-hand side represents the glacier's dependence on its previous state. The last two terms are the forcing due to a given year's anomaly in precipitation, $P'$, and melt-season temperature, $T'$, for which we use white noise with standard deviations of $\sigma_P$ and $\sigma_T$. The coefficients α and β are functions of glacier geometry: $\alpha = -\mu A_{T>0}\Delta t/(wH)$, $\beta = A_{tot}\Delta t/(wH)$, $A_{T>0} = A_{abl} + Pw/\mu\Gamma\tan\phi$, and τ is the characteristic time scale (also response time) over which the glacier responds to past forcing: $\tau = wH/(\mu\Gamma\tan\phi A_{abl})$ (see Table 1 for definitions).

### Climate Data and Parameter Selection

Meteorological data were extracted from the longest-running high-elevation weather station in North America on Niwot Ridge, CO (Figure 3B; Station D1; 1952–2010; 3743 m a.s.l). Assuming that the melt season runs from June to September, we determined that melt-season temperature ($\bar{T}$ , $\sigma_T$) is (6.3,1.3) °C and annual precipitation ($\bar{P}$ , $\sigma_P$) is (1.2, 0.22) m $a^{-1}$. Data were linearly detrended. We consider a range of near-surface lapse rates based on a global compilation of summer on-ice, near-surface lapse rates which for valley glaciers has a mean of 4.0 ± 2.1 °C $km^{-1}$ (1σ) (Table 1; DR.1). We constrain the melt-factor, $\mu$ (i.e., ablation rate per 1 °C change in *T*), based on a global compilation of $\mu$ for snow (4.5 ± 1.7 mm $day^{-1}$ $°C^{-1}$) and ice (7.7 ± 3.2 mm $day^{-1}$ $°C^{-1}$; DR.2). We also consider a relatively broad range of accumulation-area ratios (*AAR*), the ratio of the accumulation area to the total glacier area, from 0.5 to 0.8 (Meier and Post, 1962).

## IMPACT OF INTERANNUAL VARIABILITY ON MEAN GLACIER LENGTH

The flowline model (DR.6) was integrated with mid-range parameters (Table 1) for the basal slope and length of a mid-sized LGM Front Range glacier (Fall Creek glacier; Table 2; Fig. 2C). The most likely parameter set generated a standard deviation of length fluctuations ($\sigma_L$) of 370 m. For the linear model $\sigma_L$ can be solved exactly:

$$\sigma_L = \sqrt{\frac{\tau \Delta t}{2}} \sqrt{\alpha^2 \sigma_T^2 + \beta^2 \sigma_P^2} \, . \qquad (2)$$

For the same parameters as the flowline model, the linear model suggests $\sigma_L$ = 415 m, consistent with the ~15% overestimate of the flowline model results described by Roe (2011). This inter-model difference is much smaller than the parameter uncertainty (Table 1). The outer bounds of the linear model $\sigma_L$ are 180 and 800 m. A high sensitivity to *T* or *P* (i.e., large α or β), or a long response time leads to a large $\sigma_L$.

We use excursion statistics for glaciers driven by climate variability (after Roe, 2011; Reichert et al. 2002). In any given time interval *D* the mean glacier length, $\bar{L}$, cannot be known exactly, but it is described by a probability distribution (Fig. 2A). Roe (2011) showed that the most likely $\bar{L}$ can be related to the maximum glacial length, $L_{max}$, by

$$\bar{L} = L_{\max} - \sigma_L \sqrt{2 \ln\left( \frac{D\dot{\sigma}_L}{2\pi\sigma_L \ln(2)} \right)}, \qquad (3)$$

where $\dot{\sigma}_L$ is the standard deviation of the time rate of change of glacier length (see Equation 9 and A8 in Roe (2011) with $p = 0.5$). Equation 3 is quite general, and holds provided the probability distribution of glacier fluctuations is normally distributed (Fig. 1B). It has been shown to govern the variability of terminus position in flowline glacier models (Reichert et al., 2002; DR.7). Roe (2011) further demonstrated that setting $\sigma_{\dot{L}} = \sigma_L \sqrt{2/\psi\tau\Delta t}$ emulated the behavior of a standard numerical flowline model, where $\psi$ (≈10) is a factor introduced because high frequencies are damped more strongly in a numerical flowline model than is predicted by the linear model. The effect of varying $\psi$ is minimal: doubling or halving of $\psi$ results in a ± 0.6% change in $\bar{L}$ when *D* is larger than the glacier response time, τ. We consider a broad range of *D* (the duration of the climate of interest) values between 500 and 7500 years, the upper limit being the duration of the LGM sea-level low stand (Clark et al., 2009). Alternatively, a natural choice for *D* is the time interval separating two dated moraines we wish to attribute to either climate change or interannual weather variability.

**Mean Length and Signal-to-Noise Results**

For glaciers with areas between 56 and 4 km$^2$, the most likely mean glacier length is 10%–15% up valley from $L_{max}$ (Table 2, column 8). Bounds on the potential location of the mean

length—3% to 50% up valley from $L_{max}$ — represent cases in which all parameters are simultaneously given their extreme values (Table 2, columns 7 and 9). As this is unlikely, this range should be considered an outer bound on $\bar{L}$. Interpreting the cause of glacier length changes requires that we discern the competing influences of a signal (a change in $\bar{P}$ or $\bar{T}$ leading to a change in $\bar{L}$; Fig. 1B; Eqn. 3) and a noise component (climate noise $\sigma_{P,}$ $\sigma_T$ driving length fluctuations, $\sigma_L$; Fig. 1B; Eqn. 2). Glaciers with area less than 2 km$^2$ have a $\sigma_L$ comparable to their $\bar{L}$ (Table 2), and so perhaps flickered in-and-out of existence or hung on as small stagnant ice bodies during periods of the local LGM.

## DISCUSSION

Historical records of advance and retreat show that 1-20 m moraines can form from interannual variability forced advances (Table 3; references in supplemental). They also support the notion that valley glaciers fluctuate on multi-decadal timescales in response to interannual variability. It is often assumed that long glacial standstills (century scale) were required to form large LGM terminal moraines (often >10 m in height). Our results suggest that century-scale glacial standstills (terminus within 50m of the same location) did not occur for the Colorado Front Range glaciers we modeled. Rather, our numerical model shows that the longest standstills lasted ~50 years (glaciers with longer response times have a propensity for longer standstills) (Fig. 2C; DR.4; Johnson and Gillam, 1995).

Glacier length fluctuations forced by interannual variability can be viewed as a smaller scale example of the moraine survival problem, e.g., Gibbons et al. (1984). Broad LGM terminal moraine complexes (Fig. 3D), which are often conglomerates of many moraines formed by different advances, may represent cycles of kilometer-scale retreat and re-advance that are

independent of true climate change. The glacier likely formed one moraine, then left the terminal moraine complex and returned, forming a second moraine (Fig. 2C; Fig. 3D). Terminal moraines between $\bar{L}$ and $L_{max}$ are therefore not necessarily 'recessional' moraines in the classic sense.

Because maximum moraines reflect advances formed sometime after the glacier reached $\bar{L}$, dating the maximum moraine provides a minimum estimate of when a climate change initiated. Constraining the timing of retreat from $\bar{L}$ provides an estimate of when $\bar{P}$ and $\bar{T}$ changed (e.g., Young et al., 2011; Fig. 1B). Care should be taken to insure that the same advance formed the deposits sampled for cosmogenic radionuclide exposure (CRN) ages (Fig. 3C; 3D). Sampling the highest ridge in a moraine complex may not provide dates from the oldest advance (Fig. 3D). The potential for sporadic exposure of bedrock between $L_{max}$ and $L_{min}$ could also impact the interpretation of CRN concentrations in bedrock for either exposure ages or production rates.

Incoherent patterns of interannual variability from region-to-region (coherence can be expected over an ~500 kilometer length scale; Letreguilly and Reynaud, 1989) could have resulted in glacier advances and retreats at different times around the Western U.S. during the regional LGM. This effect could potentially explain the spread of ages derived from LGM maximum terminal moraines across the Western U.S (Fig. 2D; see Young et al., 2011; Thackray, 2008; Licciardi et al., 2004 for other interpretations), and even globally (Schäfer et al., 2006).

Moraines reflect maximum advances, and our results suggest that climate noise (weather) is likely to force kilometer-scale advances beyond what the mean climate conditions support. Climate estimates derived from maximum glacier geometries do not represent the local LGM-mean climate. Rather, they have a one-sided bias due to glacial length excursions down valley from $\bar{L}$. Equilibrium-line-altitudes (ELAs) and climate change estimates derived from glacier

models directly reconstructed from maximum moraines will therefore overestimate the climate change. In our setting, the central parameter range suggests this is a 10%–15% effect for LGM moraines.

## CONCLUSIONS

Interannual variability results in decadal-scale glacial length fluctuations around a mean length. The maximum excursion from the mean length is responsible for the formation of LGM terminal moraines that are, in the Colorado Rockies, most likely 10%–15% down valley from the mean LGM glacier length, with maximum bounds of 50%–3%. Interannual variability is present in all climates. We should therefore expect it to play an important role in kilometer-scale length fluctuations and moraine formation in the past and present as well as in maritime, Alpine, and continental settings (e.g. Oerlemans, 2001). Glacier response times and the magnitude of $\sigma_P$ and $\sigma_T$ will determine the expected timescale and magnitude of length fluctuation. Several modeling efforts, historical glacier extent records, and documentation of modern moraine formation support our conclusions. Glacier length fluctuations due to year-to-year climate variability should therefore be included in the interpretation of the moraine record.

## ACKNOWLEDGMENTS

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**FIGURE CAPTIONS**

Figure 1. (A) A glacier forced only by $\bar{P}$ and $\bar{T}$ (bold dashed line) over time period, *D*, leads to a steady ice profile (bold black line) at $\bar{L}$. The maximum terminal moraine forms at $\bar{L}$. (B) A glacier forced by the same $\bar{P}$ and $\bar{T}$ as in (A) but with interannual variability included (gray white noise in the inset panel, which fills a normal distribution centered at $\bar{P}$ or $\bar{T}$) results in a terminus position that fluctuates around $\bar{L}$ and is shown by the gray dashed ice profiles. Given enough time, the terminus position fills a normal distribution centered at $\bar{L}$ (the signal; Equation 3) and is characterized by $\sigma_L$ (the noise; Equation 2), the standard deviation of glacier length perturbations around $\bar{L}$. A change in $\bar{L}$ would occur with a change in $\bar{P}$ or $\bar{T}$ but not a change in $\sigma_P$ or $\sigma_T$.

Figure 2. An illustration of the relationship between the mean glacier length and moraine formation. (A) The probability density function (pdf) of possible mean lengths derived from the most likely parameter set normalized by the maximum extent of the glacier. Equation 3 gives the most likely $\bar{L}$ from this pdf. (B) An example melt-season white-noise climatology, which is used to produce the example glacier terminus history in (C). (C) An example terminus history with potential moraine-forming locations indicated by triangles. (D) Glacier length normalized to the LGM maximum terminal moraine with terminal moraines in-board from the maximum extent shown as triangles for western U.S. LGM valleys. Note the scatter in LGM maximum terminal moraines ages (ka) shown on the right (see supplemental for citations).

Figure 3. (A) Map of western U.S. glaciers during the LGM with generalized glacier outlines after Porter et al. (1983). (B) Shaded relief map showing the aerial extent of the eleven-modeled glaciers (Table 2) and the location of the weather station used in this study. $\bar{L}_{mean}$, the most likely estimate of mean length, and $\bar{L}_{max}$, the maximum possible mean length, are shown for

each modeled glacier. (C) LiDAR-derived hillshade of the terminal region of the LGM Middle Boulder Creek glacier. Cosmogenic radionuclide dates are in ka (see supplemental for citations). Note that two distinct LGM advances could have been dated leading to the spread of CRN ages. Numbers near moraines denote the number of distinct ice advances recorded in the pro-$\bar{L}$ area. (D) LiDAR hillshade of Glacier Gulch, WY LGM terminal moraine complex. Notice that 4 represents a re-advance that crosscuts advances 1-3. The 14 ice marginal features preserved support the highly mobile nature of glaciers during the LGM in the Western US.

[1]GSA Data Repository item 2013xxx, Sections DR.1–DR.8 are available online at www.geosociety.org/pubs/ft2013.htm, or on request from editing@geosociety.org or Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301, USA.

Figure 1
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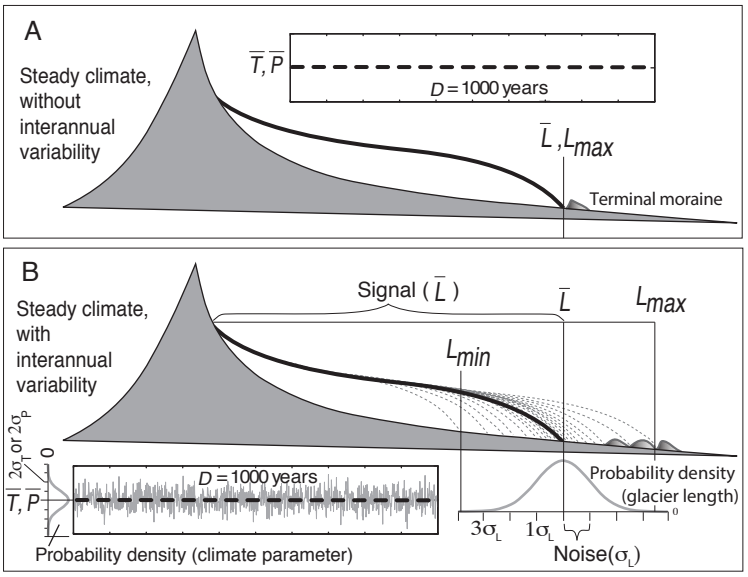

Figure 2

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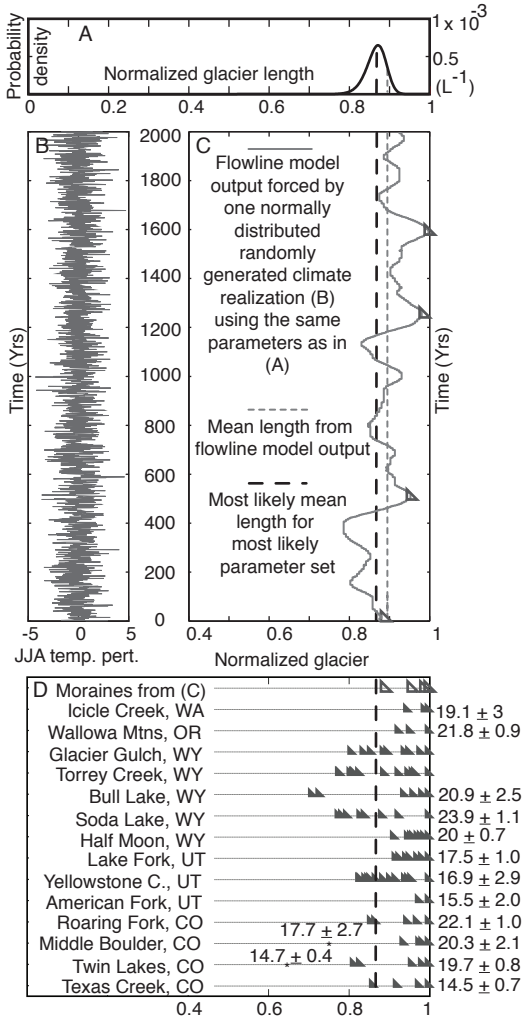

Figure 3
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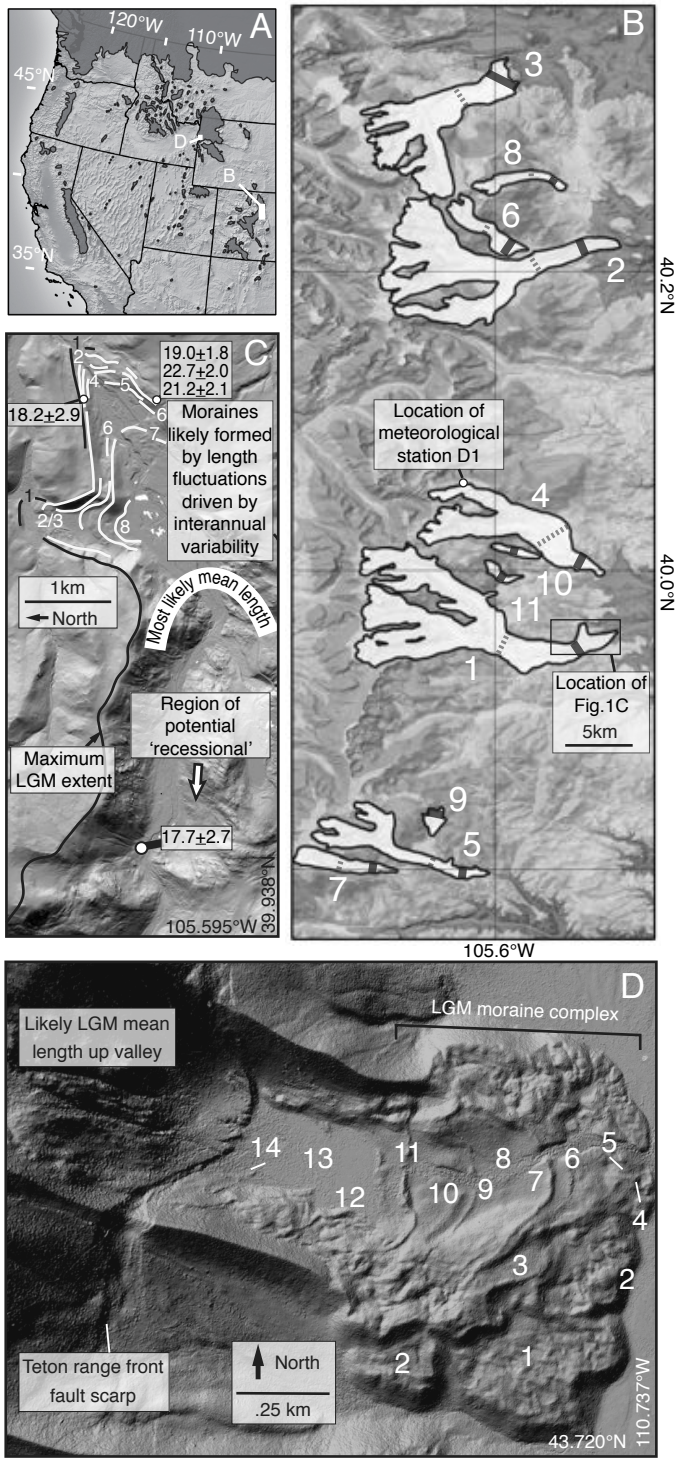

Table 1

TABLE 1. LINEAR MODEL PARAMETERS AND GEOMETRY INPUTS

| Name | Units | Description | Min. | Mean | Max. |
|---|---|---|---|---|---|
| $\mu$ | (m °C$^{-1}$ yr$^{-1}$) | Melt factor | .5 | .7 | .9 |
| $\Gamma$ | (°C km$^{-1}$) | On-ice near-surface lapse | 3.5 | 5 | 6.5 |
| *AAR* | | Accumulation-area ratio | .5 | .65 | .8 |
| $\overline{P}$ | (m) | Mean annual precipitation | .6 | 1.2 | 2.4 |
| $\sigma_T$ | (°C) | Std. of summertime temp. | 1 | 1.3 | 1.6 |
| $\sigma_P$ | (m a$^{-1}$) | Std. of annual precipitation | .11 | .22 | .44 |
| *D* | (yr) | Duration of climate change | 500 | 4000 | 7500 |
| Geometry Inputs | | | | | |
| $A_{tot}$ | | Total area of the glacier | | | |
| $A_{abl}$ | | Ablation area of the glacier | | | |
| $tan\phi$ | | Slope of the glacier bed | | | |
| *w* | | Width of the ablation zone | | | |
| *H* | | Thickness of the glacier | | | |

Table 2

TABLE 2. GLACIER GEOMETRY INPUTS, MEAN LENGTHS, AND SIGNAL-TO-NOISE RATIOS

| Glacier name | Area ($km^2$) | Slope | Width (km) | Height (km) | $L_{max}$ (km) | $\bar{L}_{min}$ % to $L_{max}$ | $\bar{L}_{mean}$ % to $L_{max}$ | $\bar{L}_{max}$ % to $L_{max}$ | Mean Signal-to-Noise | Response time ($\tau$) |
|---|---|---|---|---|---|---|---|---|---|---|
| 1. Middle Boulder | 56.62 | 0.031 | 1.30 | 0.22 | 18.55 | 97 | 86 | 62 | 16.68 | 133.44 |
| 2. North Saint Vrain | 45.89 | 0.050 | 1.16 | 0.19 | 15.25 | 97 | 88 | 68 | 19.63 | 77.95 |
| 3. Bear Lake | 31.19 | 0.055 | 1.56 | 0.16 | 12.61 | 97 | 87 | 63 | 17.70 | 118.99 |
| 4. North Boulder | 26.00 | 0.091 | 1.32 | 0.11 | 12.47 | 97 | 89 | 71 | 22.94 | 50.10 |
| 5. Fall Creek | 14.57 | 0.078 | 0.58 | 0.14 | 10.55 | 96 | 86 | 65 | 17.86 | 57.75 |
| 6. Hunter's Creek | 6.25 | 0.138 | 0.81 | 0.09 | 6.19 | 96 | 85 | 59 | 16.12 | 69.00 |
| 7. Mill Creek | 5.80 | 0.089 | 0.52 | 0.11 | 5.94 | 95 | 79 | 41 | 10.27 | 91.33 |
| 8. Roaring Fork | 4.11 | 0.178 | 0.51 | 0.08 | 5.72 | 96 | 86 | 63 | 17.69 | 45.59 |
| 9. Silver Creek | 1.67 | 0.131 | 0.45 | 0.04 | 2.07 | 83 | 28 | –119 | 1.11 | 67.22 |
| 10. Rainbow Creek | 1.42 | 0.120 | 0.44 | 0.04 | 2.92 | 87 | 43 | –77 | 2.11 | 83.50 |
| 11. Horseshoe Creek | 1.41 | 0.137 | 0.35 | 0.05 | 2.86 | 90 | 57 | –30 | 3.64 | 73.85 |

*Signal-to-noise ratio $= \bar{L}$

Table 3

TABLE 3. HISTORICAL EXAMPLES OF MORAINES FORMED BY INTERANNUAL VARIABILITY FORCED ADVANCES

| Glacier | Time | Relief | Citation |
|---|---|---|---|
| Engabreen, NO | <20yr | 15m | Worsely & Alexander, 1976 |
| Nigardsbreen, NO | <10yr | <10m | Nussbaumer et al., 2011 |
| Six glaciers, NO | <10yr | <3m | Winkler and Matthews, 2010 |
| U. Grindelwald, CH | <10yr | 10m | Zumbühl et al., 2008 |
| Des Bossons, FR | ~10yr | 20m | Nussbaumer & Zumbühl, 2012 |
| Mer de Glace, FR | ~10yr | <10m | Zumbühl et al., 2008 |

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Supplemental material

***Section DR.1 On-ice near surface lapse rates***

The selection of the lapse rate, $\Gamma$, for glaciological purposes must be made with care because reported summertime on-ice (ᵕ measurements only on glacier surface), near-surface (ᵕ measurements made 2 m above the surface) lapse rates vary by nearly a factor of eight (Table A). Since $\Gamma$ governs how ablation changes with elevation, much of the uncertainty in the results arises from this parameter. We argue that observed on-ice, summertime lapse rates provide a better approximation of the relevant paleo lapse rates than either the standard moist adiabatic lapse rate, observed free atmospheric lapse rates, or observed off-ice (measurements made off-glacier) near-surface lapse rates— even in those in the Front Range.

Lapse rates in the free atmosphere are determined by atmospheric vertical mixing and moisture. However, surface lapse rates are controlled surface radiative transfer and by the near surface environment (surface albedo, roughness, topographic aspect, and local meteorological effects, e.g., Marshall and Sharp, 2007). While the use of modern summer near-surface temperature lapse rates from the Front Range is likely more appropriate than the often-used 6.5 °C $km^{-1}$ moist adiabatic lapse rate, modern environmental conditions obviously differ greatly from likely summer conditions on an LGM glacier (presence of ice, reduced roughness, different elevation and topography due to the presence of the glacier). We therefore used modern on-ice near-surface temperature lapse rates to guide our uncertainty analysis. Table A shows that the most likely mean summer on-ice near surface lapse rate is 4.9 °C $km^{-1}$ with a 1σ value of 1.7 °C $km^{-1}$ . Extreme mean values are 1.1 °C $km^{-1}$ (Pasterze Glacier, Austria) and 7.9 °C $km^{-1}$ (Greenland Ice Sheet).

Table DR.1 COMPILATION OF ON-ICE NEAR SURFACE LAPSE RATES

| **Valley Glaciers** | | | | | |
|---|---|---|---|---|---|
| Location | Classification | Latitude | Lapse Rate °C $km^{-1}$ | Period of Averaging | Reference |
| Pasterze Glacier, Austria | Valley | 47°N | 1.1 | June, July | Greuell and Smeets, 2001 |
| John Evans Glacier, Canada | Valley | 80°N | 1.1 | Summer | Arendt and Sharp, 1999 |
| Pasterze Glacier, Austria | Valley | 47°N | 1.4 | June, July | Greuell and Smeets, 2001 |
| Haut Glacier, Switzerland | Valley | 45.97°N | 2.0 | Summer | Strasser et al., 2004 |
| John Evans Glacier, Canada | Valley | 80°N | 3.1 | Summer | Gardner et al., 2009 |
| Pasterze Glacier, Austria | Valley | 47°N | 3.5 | June, August | Denby and Greuell, 2000 |
| Franz Josef Glacier, New Zealand | Valley | 43.49°N | 4.8 | Annual | Anderson et al., 2006 |
| Keqicar Glacier, Tien Shan | Valley | 41.75°N | 5.0 | July | Li et al, 2011 |
| South Glacier, Yukon | Valley | 60.8°N | 5.3 | Annual | MacDougall and Flowers, 2011 |
| Storglaciären, Sweden | Valley | 67.9°N | 5.5 | Annual | Hock and Holmgren, 2005 |
| North Glacier, Yukon | Valley | 60.8°N | 6.0 | Annual | MacDougall and Flowers, 2011 |
| Keqicar Baqi, Tien Shan, China* | Valley | 41.75°N | 6.0 | Summer | Zhang et al., 2007* |
| South Cascade Glacier, WA | Valley | 48.35°N | 6.5 | Summer | Anslow et al., 2008 |
| Juncal Norte, Chile | Valley | 32.6°S | 6.5 | Summer | Petersen and Pellicciotti, 2011 |
| Miage glacier, Italy* | Valley | 45.5°N | 6.7 | Summer | Brock et al., 2010* |

| | | | | | |
|---|---|---|---|---|---|
| Baltoro Glacier, Pakistan* | Valley | 35.7°N | 7.5 | Summer | Mihalcea et al., 2006* |
| Miage glacier, Italy* | Valley | 45.5°N | 8.0 | Summer | Brock et al., 2010* |
| | *Valley Glacier Mean:* | | *4.7±2.2* | *w/ debris-cover* | |
| | | | *4.0±2.1* | *w/o debris-cover* | |

**Ice Sheet and Ice Cap**

| Location | Classification | Latitude | Lapse Rate °C km$^{-1}$ | Period of Averaging | Reference |
|---|---|---|---|---|---|
| Vatnajökull, Iceland | Ice Cap | 64.1°N | 3.6 | Summer | Oerlemans et al., 1999 |
| Prince of Whales Icefield, Canada | Ice Cap | 78°N | 3.7 | JJA | Marshall et al., 2007 |
| Prince of Whales Icefield, Canada | Ice Cap | 78°N | 4.3 | Summer | Marshall et al., 2007 |
| Prince of Wales Ice Field, Canada | Ice Cap | 78°N | 4.4 | Summer | Marshall and Sharp, 2009 |
| Prince of Whales Icefield, Canada | Ice Cap | 78°N | 4.6 | Summer | Gardner et al., 2009 |
| Devon Ice Cap, Canada | Ice Cap | 75.2°N | 4.8 | Summer | Mair et al. 2005 |
| Prince of Whales Icefield, Canada | Ice Cap | 78°N | 4.8 | JJA | Marshall et al., 2007 |
| Devon Ice Cap, Canada | Ice Cap | 75.2°N | 4.9 | Summer | Gardner et al., 2009 |
| Prince of Whales Icefield, Canada | Ice Cap | 78°N | 5.3 | JJA | Marshall et al., 2007 |
| Langjökull, Iceland | Ice Cap | 64.5°N | 5.6 | JJA | Gudmundsson, et al, 2003 |
| Vestari Hagafellsjökull, Iceland | Ice Cap | 64.5°N | 5.7 | Summer | Hodgkins et al., 2012 |
| King George Island, Antarctica | Ice Cap | 62.3°S | 6 | Summer | Braun and Hock, 2004 |
| Aggasiz Ice Cap, Canada | Ice Cap | 80.2°N | 6.4 | Summer | Gardner et al., 2009 |
| Greenland Ice Sheet (>1000m a.s.l) | Ice Sheet | ~67°N | 2.4 | Summer | Oerlemans and Vugts, 1993 |
| Greenland Ice Sheet | Ice Sheet | 60- | 4 | June | Steffen and Box, 2001 |
| NE Greenland Ice Sheet | Ice Sheet | 70- | 4 | June, August | Boggild et al., 1994 |
| Greenland Ice Sheet (<1000m | Ice Sheet | 60- | 4.3 | Summer | Hanna et al., 2005 |
| Greenland Ice Sheet | Ice Sheet | 60- | 5 | June, July | Box and Rinke, 2003 |
| Greenland Ice Sheet (<1000m | Ice Sheet | ~67°N | 5 | Summer | Oerlemans and Vugts, 1993 |
| West Greenland Ice Sheet | Ice Sheet | ~67°N | 5.8 | Mean | van den Broeke et al., 2011 |
| Greenland Ice Sheet (<1000m | Ice Sheet | ~67°N | 6.3 | Summer | Oerlemans and Vugts, 1993 |
| West Greenland Ice Sheet | Ice Sheet | ~67°N | 7.4 | Mean | van den Broeke et al., 2011 |
| Greenland Ice Sheet (>1000m | Ice Sheet | 60- | 7.9 | Summer | Hanna et al., 2005 |
| | *Ice Sheet and Ice Cap Mean:* | | *5.1±1.2* | | |

| | | | |
|---|---|---|---|
| * Debris-covered glacier | Mean of all cited lapse rates: | *4.9±1.7* | *w/ debris-cover* |
| | Mean of all cited lapse rates: | *4.7±1.6* | *w/o debris-cover* |

### *Section DR.2 Melt-factors*

The melt factor, $\mu$, employed in our ablation parameterization is a simplified form of the often used positive-degree-day model that relates mean summer temperatures to vertical surface mass loss. The melt factor $\mu$ is converted from published positive-degree-day factors by assuming a melt season covering the months of June, July, and August (Table DR.2). The selection of $\mu$ must be made with care as positive degree-day factors for snow can vary by nearly a factor of ten, and for ice by a factor of six. We combine and supplement several previous compilations of snow and ice melt-factors for modern glaciers and mountainous regions. Table DR.2 shows that the most likely positive degree day factor for ice: is 7.7 mm day$^{-1}$ °C$^{-1}$ with a 1$\sigma$ value of 3.2 mm day$^{-1}$ °C$^{-1}$ with extreme values of 20 mm day$^{-1}$ °C$^{-1}$; Van de Wal (1992) and 2.6 mm day$^{-1}$ °C$^{-1}$; Zhang et al. (2006); and the most likely positive degree day factor for snow is 4.5 m °C$^{-1}$ a$^{-1}$ with a 1$\sigma$ value of 1.7 mm day$^{-1}$ °C$^{-1}$ with extreme values of 11.6 mm day$^{-1}$°C$^{-1}$; Kayastha et al. (2000) and 1.4 mm day$^{-1}$ °C$^{-1}$ Howat et al. (2007). It is important to note that our parameter combinations produce mass balance values that are reasonable for continental climates.

TABLE DR.2 GLOBAL COMPILATION OF POSITIVE DEGREE-DAY MELT FACTORS ($mm \cdot day^{-1} C^{-1}$)

| **Greenland** | | | | | | | |
|---|---|---|---|---|---|---|---|
| Location | Snow | Ice | Elevation | Latitude | Duration | Reference | Cited in |
| Thule Ramp, Greenland | | 12 | 570 | 76°25'N | 1 Jul - 31 Jul 1954 | Schytt, 1955 | Hock, 2003 |
| Thule Ramp, Greenland | | 7 | 570 | | 1 Aug-31 Aug 1954 | Schytt, 1955 | Hock, 2003 |
| Camp IV-EGIG, Greenland | | 18.6 | 1013 | 69°40'N | Melt season 1959 | Ambach, 1988a | Hock, 2003 |
| GIMEX profile, Greenland | | 8.7 | 341 | 67°06'N | 10 Jun-31 Jul 1991 | Van de Wal, 1992 | Hock, 2003 |
| GIMEX profile, Greenland | | 9.2 | 519 | 67°06'N | 15 Jun-6 Aug 1991 | Van de Wal, 1992 | Hock, 2003 |
| GIMEX profile, Greenland | | ***20*** | 1028 | 67°04'N | 15 Jun-6 Aug 1991 | Van de Wal, 1992 | Hock, 2003 |
| Qamanârssûp sermia, Greenland | 2.8 | 7.3 | 370-1410 | 64°28'N | 1979-1987 | Johannesson et al., 1995 | Hock, 2003 |
| Qamanârssûp sermia, Greenland | 2.9 | 8.2 | 790 | 64°28'N | 512 days (1980-86) | Braithwaite, 1995 | Hock, 2003 |
| Nordboglacier, Greenland | 3.7 | 7.5 | 880 | 61°28'N | 415 days (1979-83) | Braithwaite, 1995 | Hock, 2003 |
| Kronprins Christian Land, Greenland | | 9.8 | 380 | 79°54'N | 8 Jul - 27 Jul 1999 | Braithwaite et al., 1998 | Hock, 2003 |
| Hans Tausen Ice Cap, Greenland | | 5.9 | 540 | 82°49'N | 2 Jul-5 Aug 1994 | Braithwaite et al., 1998 | Hock, 2003 |
| Qamanârssûp sermia, Greenland | 2.5 | 7.7 | ~800 | 64°28'N | Summer | Braithwaite, 1989 | Braithwaite and Zhang, 2000 |
| Qamanârssûp sermia, Greenland | | 7.9 | 790 | 64°28'N | May-Sep 1980-1986 | Braithwaite, 1993 | Braithwaite and Zhang, 2000 |
| *Greenland means:* | *3.0±0.5* | *10.0±4.4* | | | | | |
| **Europe/Americas/NZ** | | | | | | | |
| Location | Snow | Ice | Elevation | Latitude | Duration | Reference | Cited in |
| Aletshgletscher, Switzerland | | 11.7 | 2220 | 46°270' N | 2 Aug - 27 Aug,1965 | Lang, 1986 | Hock, 2003 |
| Ålfotbreen, Norway | 4.5 | 6 | 850-1400 | 61°45'N | 1961-1990 | Laumann and Reeh, 1993 | Hock, 2003 |
| Ålfotbreen, Norway | 3.5 | 5.5 | 1450-2200 | 61°34'N | 1961-1990 | Laumann and Reeh, 1993 | Hock, 2003 |
| Ålfotbreen, Norway | 4 | 5.5 | 300-2000 | 61°41'N | 1961-1990 | Laumann and Reeh, 1993 | Hock, 2003 |
| Nigardsbreen, Norway | 4.4 | 6.4 | 300-2000 | 61°41'N | 1964-1990 | Johannesson et al., 1995 | Hock, 2003 |
| Storglaciären, Sweden | 3.2 | | 1550 | 67°55'N | 5 Jul-7 Sep 1993 | Hock, 1999 | Hock, 2003 |
| Storglaciären, Sweden | | 6 | 1370 | 67°55'N | 5 Aug - 12 Aug 1993 | Hock, 1999 | Hock, 2003 |
| Storglaciären, Sweden | | 6.4 | 1370 | 67°55'N | 19 Jul-27 Aug 1994 | Hock, 1999 | Hock, 2003 |
| Storglaciären, Sweden | | 5.4 | 1250 | 67°55'N | 9 Jul-4 Sep 1994 | Hock, 1999 | Hock, 2003 |
| Vestfonna, Spitzbergen | | 13.8 | 310-410 | ~80°N | 26 Jun - 5 Aug 1958 | Schytt, 1964 | Hock, 2003 |
| Satujo□kull, Iceland | 5.6 | 7.7 | 800-1800 | ~65°N | 1987-1992 | Johannesson et al., 1995 | Hock, 2003 |
| Aletshgletscher, Switzerland | 5.3 | | 3366 | 46°270' N | 3 Aug-19 Aug, 1973 | Lang, 1986 | Hock, 2003 |
| John Evans Glacier, Canada | 5.5 | | 260 | 79°40' N | 27 Jun-29 Jun 1996 | Arendt and Sharp, 1999 | Hock, 2003 |
| John Evans Glacier, Canada | 4.1 | | 820 | 79°40' N | 19 Jun - 14 Jul 1996 | Arendt and Sharp, 1999 | Hock, 2003 |
| John Evans Glacier, Canada | 3.9 | | 820 | 79°40' N | 23 May - 1 Jul 1998 | Arendt and Sharp, 1999 | Hock, 2003 |
| John Evans Glacier, Canada | 3.9 | | 1180 | 79°40' N | 25 Jun-19 Jul 1996 | Arendt and Sharp, 1999 | Hock, 2003 |
| John Evans Glacier, Canada | 2.7 | | 1180 | 79°40' N | 31 May - 19 Jul 1996 | Arendt and Sharp, 1999 | Hock, 2003 |
| John Evans Glacier, Canada | | 7.6 | 260 | 79°40' N | 4 Jul - 16 Jul 1996 | Arendt and Sharp, 1999 | Hock, 2003 |
| John Evans Glacier, Canada | | 8.1 | 820 | 79°40' N | 15 Jul - 19 Jul 1996 | Arendt and Sharp, 1999 | Hock, 2003 |
| John Evans Glacier, Canada | | 5.5 | 820 | 79°40' N | 2 Jul -19 Jul 1998 | Arendt and Sharp, 1999 | Hock, 2003 |
| Weissfluhjoch, Switzerland | 4.2 | | 2540 | 46°48'N | 28 year record | de Quervain, 1979 | Braithwaite and Zhang, 2000 |
| Franz Josef Glacier, New Zealand | 3 | 6 | 122 | 43°28'N | Summer | Woo and Fitzharris, 1992 | Braithwaite and Zhang, 2000 |
| Saint Supphellebreen, Norway | | 6.3 | | 61°30'N | Summer | Orheim, 1970 | Braithwaite and Zhang, 2000 |
| Glacier de Sarennes, France | 3.8 | 6.2 | ~3000 | 45°10'N | Summer | Vincent and Vallon, 1997 | Braithwaite and Zhang, 2000 |
| Griesgletscher, Switzerland | | 8.9 | 2287 | 46°39'N | 112 summer days | Braithwaite, 2000 | Braithwaite and Zhang, 2000 |
| Australian Alps | 2.9 | | 1250 | 36°30'S | 1966-1985 | Whetton, et al., 1996 | Brugger, 2010 |
| Blo□ndujo□kull, Kv´ıslajökull, Iceland | 4.5 | 5 | 115 | 64°50'N | Summer | Johennesson, 1997 | Brugger, 2010 |
| Illvirajökull, Iceland | 5.6 | 7.6 | 115 | 64°50'N | Summer | Johennesson, 1997 | Brugger, 2010 |
| Glacier Upsala, Patagonia | | 7.1 | 350 | 49°58S | Summer 1993-1994 | Naruse et al., 1997 | Brugger, 2010 |
| South Cascade Glacier, USA | | 6.2 | 1980 | 48°21'N | Summer | Tangborn,1999 | Brugger, 2010 |
| Rabots Glacier, Sweden | 4.7 | 6.8 | ~1300 | 67°55'N | Summer | Refsnider, 2001 | Brugger, 2010 |
| Sverdrup Glacier, Canada | | 4 | 300 | 75°N | Summer 1963 | Braithwaite, 1981 | |

| | | | | | | | |
|---|---|---|---|---|---|---|---|
| Andrews Glacier, USA | 4.3 | | | | Summer | Outcalt and MacPhail, 1965 | Lauman & Reeh, 1993 |
| Storsteinsfjellbreen, Norway | 5,6 | 7.5 | | 68°15'N | Summer | Pytte and Liestol, 1966 | Lauman & Reeh, 1993 |
| Storbreen, Norway | | 5.5 | | 61°34'N | Summer 1949-1965 | Liestol, 1967 | Lauman & Reeh, 1993 |
| White Glacier, Canada | | 4.9 | 210 | 79°N | Summer 1960-1962 | Braithwaite, 1981 | |
| Alfotbreen, Norway | 5.3 | 7.5 | | 61°45'N | Summer 1965 | NVE, 1965 | Lauman & Reeh, 1993 |
| Various Swiss glaciers | | 6 | | ~46°30'N | Summer | Kasser, 1959 | Braithwaite and Zhang, 2000 |
| Fillefjell, Norway | 3.9 | | | 61°10'N | Summer 1967-1964 | Furmyr and Tollan, 1975 | Lauman & Reeh, 1993 |
| Moreno glacier, Argentina | | 7 | 330 | 50°28'S | 1993 -1994 | Takeuchi et al., 1996 | Hock, 2003 |
| Martial Este Glacier, Argentina | 4.7 | 9.4 | 990 | 54°47'S | Dec 2005 - Feb 2006 | Buttstadt et al., 2009 | |
| Haut Glacier d'Arolla, Switzerland | 7.7 | 10.8 | ~2900 | 45°58'N | May - Sep 2001 | Pellicciotti et al., 2005 | |
| Mount Shasta, Cascade Range, USA | 1.6 | 6.9 | 2600 | 41°12'N | May - Nov 2002 | Howat et al., 2007 | |
| Mount Shasta, Cascade Range, USA | ***1.4*** | 5.5 | 3000 | 41°12'N | May - Nov 2002 | Howat et al., 2007 | |
| Hansbreen, Svalbard | 6 | 8.3 | 180 | 77°05'N | JJA, 2008 | Grabiec et al., 2012 | |
| Franz Josef Glacier, New Zealand | 4.6 | 7.1 | | 43°28'N | Summer | Anderson, 2004 | |
| Hansbreen, Svalbard | | 6.8 | 316 | 77°05'N | 1994-1995 | Szafraniec, 2002 | |
| Gran Campo Nevado Ice Cap, Chile | | 7.6 | 450 | 53°S | Feb - Apr 2000 | Schneider et al., 2007 | |
| Tasman Glacier, New Zealand | | 4.5 | 1360 | 43°37'S | 1985-1986 | Kirkbride, 1995 | |
| Tasman Glacier, New Zealand | | 5 | 1360 | 43°37'S | 1986-1987 | Kirkbride, 1995 | |
| Tasman Glacier, New Zealand | | 3.9 | 960 | 43°37'S | 1985-1986 | Kirkbride, 1995 | |
| Tasman Glacier, New Zealand | | 3.6 | 960 | 43°37'S | 1986-1987 | Kirkbride, 1995 | |
| Glacier de Saint-Sorlin, France | 4 | 6.4 | 2760 | 45°N | 21 Jul- 31 Jul 2006 | Six et al., 2009 | |
| Koryto Glacier, Kamchatk, Russia | 4.7 | 7 | 810 | 54°50'N | 7 Aug-12 S. 2000 | Konya et al., 2004 | |
| *Europe/America/NZ means:* | *4.3±1.3* | *6.8±2.0* | | | | | |

**Central Asia**

| Location | Snow | Ice | Elevation | Latitude | Duration | Reference | Cited in |
|---|---|---|---|---|---|---|---|
| Urumqi glacier,Tien Shan, China | 6.3 | 8.5 | 3831-3945 | ~42°N | 1986-1993 | Liu et al., 1996 | Zhang_etal 2006 |
| Urumqui glacier,Tien Shan, China | | 7.3 | 3754-3898 | ~42°N | 1986-1988 | Liu et al., 1996 | Zhang_etal 2006 |
| Urumqui glacier,Tien Shan, China | 3.1 | | 4048 | ~42°N | 1986-1993 | Liu et al., 1996 | Zhang_etal 2006 |
| Keqicar Baqi, Tien Shan, China | | 4.5 | 3347 | ~42°N | 28 Jun- 12 Sep 2003 | Zhang et al., 2005 | Zhang_etal 2006 |
| Keqicar Baqi, Tien Shan, China | | 7 | 4216 | ~42°N | 11 Jul-13 Sep 2003 | Zhang et al., 2005 | Zhang_etal 2006 |
| Qiongtailan glacier, Tien Shan, China | | 4.5 | 3675 | ~42°N | 17 Jun- 14 Aug 1978 | Zhang et al., 2006 | |
| Qiongtailan glacier, Tien Shan, China | | 7.3 | 4100 | ~42°N | 25 Jun-14 Aug 1978 | Zhang et al., 2006 | |
| Qiongtailan glacier, Tien Shan, China | | 8.6 | 4200 | ~42°N | 21 Jun-31 Jul 1978 | Zhang et al., 2006 | |
| Qiongtailan glacier, Tien Shan, China | 3.4 | | 4400 | ~42°N | 21 Jun- 11 Aug 1978 | Zhang et al., 2006 | |
| Hailuogou, Hengduan mtns, China | | 5 | 3301 | ~30°N | Aug 1982- Aug 1983 | Zhang et al., 2006 | |
| Baishuihe Hengduan mtns, China | | 13.3 | 4600 | ~30°N | 23 Jun- 30 Aug 1982 | Zhang et al., 2006 | |
| Baishuihe,Hengduan mtns, China | 5.9 | | 4800 | ~30°N | 26 Jun- 11 Jul 1982 | Zhang et al., 2006 | |
| Dagongba glacier, Hengduan, China | | 13.2 | 4540 | ~30°N | Sep 1982- Sep 1983 | Zhang et al., 2006 | |
| Xiaogongba glacier, China | | 12 | 4550 | ~30°N | Jul 1982- Jul 1983 | Zhang et al., 2006 | |
| Batura, Karakoram, China | | 3.4 | 2780 | ~36°N | Jun-Aug 1975 | Zhang et al., 2006 | |
| Teram Kangri, Karakoram, China | | 5.9 | 4630 | ~36°N | 25 Jun- 7 Sep 1987 | Zhang et al., 2006 | |
| Teram Kangri, Karakoram, China | | 6.4 | 4650 | ~36°N | 24 Jun- 7 Sep 1987 | Zhang et al., 2006 | |
| Qirbulake, Karakoram, China | | ***2.6*** | 4750 | ~36°N | 6 Jun- 30 Jul 1960 | Zhang et al., 2006 | |
| Yangbulake, Karakoram, China | | 4.3 | 4800 | ~36°N | 1Jul - 5 Jul 1987 | Zhang et al., 2006 | |
| Meikuang, Kunlun Shan, China | | 3 | 4840 | ~36°N | 7 May- Sep 1989 | Zhang et al., 2006 | |
| Halong, Kunlun Shan, China | | 4.7 | 4616 | ~36°N | 15 Jun- 28 Jun 1981 | Zhang et al., 2006 | |
| Halong, Kunlun Shan, China | | 3.6 | 4900 | ~36°N | 14 Jun 27 Jun 1981 | Zhang et al., 2006 | |
| Xiaodongkemadi, Tanggula, China | | 13.8 | 5425-5475 | ~32°30'N | Jul- Aug 1993 | Kayastha et al., 2003 | Zhang_et al., 2006 |
| Qiyi, Qilian Shan, China | | 7.2 | 4305-4619 | ~39°N | Jul- Aug 2002 | Kayastha et al., 2003 | Zhang_et al., 2006 |
| Kangwure, Himalaya, China | | 9 | 5700-6000 | | 20 Jul-25 Aug 1993 | Zhang et al., 2006 | |
| Urumqi Glacier, Tien shan, China | 5.2 | 8.4 | | ~42°N | Summer | Cui, 2009 | Xianzhong, et al., 2010 |

| | | | | | | | |
|---|---|---|---|---|---|---|---|
| Urumqi Glacier, Tien shan, China | 3.1 | 7.1 | | ~42°N | Summer | Cui, 2009 | Xianzhong, et al., 2010 |
| Urumqi Glacier, Tien shan, China | 5.2 | 7.1 | | ~42°N | Summer | Cui, 2009 | Xianzhong, et al., 2010 |
| Urumqi Glacier, Tien shan, China | | 4 | | ~42°N | Summer | Cui, 2009 | Xianzhong, et al., 2010 |
| Baishui Glacier, Hengduan, China | | 4.92 | 4200 | 26°00'N | 26 Jun-11 Jul 1982 | Liu, 1996 | Xianzhong, et al., 2010 |
| Baishui Glacier, Hengduan, China | | 10.3 | 4600 | 26°00'N | Sept 2008 | Xianzhong, et al., 2010 | |
| Baishui Glacier, Hengduan, China | | 13.6 | 4700 | 26°00'N | Sept 2008 | Xianzhong, et al., 2010 | |
| Baishui Glacier, Hengduan, China | | 14.1 | 4800 | 26°00'N | Sept 2008 | Xianzhong, et al., 2010 | |
| Baishui Glacier, Hengduan, China | 2.4 | | 4400 | 26°00'N | 13 May-6 Jun 2009 | Xianzhong, et al., 2010 | |
| Baishui Glacier, Hengduan, China | 2.8 | | 4500 | 26°00'N | 13 May-6 Jun 2009 | Xianzhong, et al., 2010 | |
| Baishui Glacier, Hengduan, China | 4.6 | | 4600 | 26°00'N | 5 May - 6 Jun 2009 | Xianzhong, et al., 2010 | |
| Baishui Glacier, Hengduan, China | 5.2 | | 4700 | 26°00'N | 13 May-6 Jun 2009 | Xianzhong, et al., 2010 | |
| Baishui Glacier, Hengduan, China | 5.8 | | 4800 | 26°00'N | 13 May - 6 Jun 2009 | Xianzhong, et al., 2010 | |
| Dokriani Glacier, Himalaya | 5.9 | | 4000 | 31°45' N | 4 Jun-6Jun 1995 | Singh and Kumar, 1996 | Hock, 2003 |
| Dokriani Glacier, Himalaya | 5.7 | 7.4 | 4000 | 31°45' N | 4 days (1997-98) | Singh et al., 2000a,b | Hock, 2003 |
| Glacier AX010, Himalaya | 7.3 | 8.1 | 4956 | 27°45' N | Jun-Aug 1978 | Kayastha et al., 2000a | Hock, 2003 |
| Glacier AX010, Himalaya | 8.7 | 8.8 | 5072 | 27°45' N | Jun-Aug 1978 | Kayastha et al., 2000a | Hock, 2003 |
| Glacier AX010, Himalaya | ***11.6*** | | 5245 | 27°45' N | 1 Jun-31 Aug 1978 | Kayastha et al., 2000a | Hock, 2003 |
| Khumbu Glacier, Himalaya | | 16.9 | 5350 | 28°00'N | 21 May-1 Jun 1999 | Kayastha et al., 2000b | Hock, 2003 |
| Rakhiot Glacier, Himalaya | | 6.6 | 3350 | 35°22'N | 18 Jul-6 Aug 1986 | Kayastha et al., 2000b | Hock, 2003 |
| Yala Glacier, Himalaya | | 9.3 | 5120 | 28°14'N | 1 Jun-31 Jul 1996 | Kayastha, 2001 | Hock, 2003 |
| Yala Glacier, Himalaya | | 10.1 | 5270 | 28°14'N | 1 Jun-31 Jul 1996 | Kayastha, 2001 | Hock, 2003 |
| *Central Asia means:* | *5.4±2.3* | *7.9±3.6* | | | | | |
| **Non-glaciated Sites** | | | | | | | |
| Location | Snow | Ice | Elevation | Latitude | Duration | Reference | Cited in |
| Gooseberry Creek, Utah, USA | 2.5 | | 2650 | ~38°N | 23 Apr-9 May 1928 | Clyde, 1931 | Hock, 2003 |
| Weissfluhjoch, Switzerland | 4.5 | | 2540 | 46°48'N | Snowmelt season | Zingg, 1951 | Hock, 2003 |
| 3 basins in USA | 2.7 | | | | Several seasons | C. of Engineers, 1956 | Hock, 2003 |
| 3 basins in USA | 4.9 | | | | Several seasons | C. of Engineers, 1956 | Hock, 2003 |
| Former European USSR | 5.5 | 7 | 1800-3700 | | | Kuzmin, 1961, p. 117 | Hock, 2003 |
| 12 sites in Finland | 3.9 | | | ~60-68°N | 1959-1978 | Kuusisto, 1980 | Hock, 2003 |
| *Non-glaciated site means:* | *4.0±1.2* | *7.0* | | | | | |
| *Mean meltfactor for all examples in the literature:* | ***4.5±1.7*** | ***7.7±3.2*** | | | | | |

### *Section DR.3 Discussion of terminal moraine assumptions*

In order to support our assumption that terminal moraines can form during advances driven by interannual variability without long term terminus standstills (< 50 years; a time scale supported by flowline modeling (see Roe, 2011 Figure 4)), we present a review of the moraine sedimentological literature (Table DR.3), which shows that the majority of moraines with constrained formation periods form over periods less than 50 years. The development of a universal model for the timescale of moraine formation has been hampered by the complexity of formational processes, the abundance of unconstrainable variables and initial conditions. But it is important to note that all moraine formation timescales found in the literature were less than 50 years. The length of time needed to form terminal moraines is dependent on the process of formation and can be constrained only crudely. Ice marginal indicators are typically divided into glaciotectonic, push, hummocky, drop moraines, and ice-contact fans but composite moraines are common (Benn and Evans, 1998). For the purposes of justifying the short timescale of ice marginal deposit formation (<50 years), we further divide the indicators into those that are independent of terminus standstills (glaciotectonic, push and hummocky moraines) and those that are dependent on terminus standstills (drop moraines and ice-contact fans). Note the dominance of push moraines in the table. The authors made no attempt to bias the type of moraines presented in this table. Rather more research has been focused on push moraines or push moraines are more common. We use the broad, continuum definition of push moraines used in Bennett (2001).

TABLE DR.3 COMPILATION OF MORAINE FORMATION TIMES.

| Region | Time period | Type | Sub-Category | Formation Time | Height | Reference |
|---|---|---|---|---|---|---|
| ***Moraines independent of terminus standstills*** | | | | | | |
| Argentina | Modern | Glaciotectonic | Folding and Thrusting | <13 years | 15-50m | Glasser & Hambrey, 2002 |
| Svalbard | Modern | Push | Surge | Likely <5yrs | 30-40m | Boulton et al., 1999 |
| Svalbard | LIA | Push | Englacial thrusts | | 45m | Hambrey & Huddart, 1995 |
| Chile | Neoglacial | Push | Formed of subglacial clasts | | 20-40m | Glasser et al., 2006 |
| Svalbard | Neoglacial | Thrust | Melt out thrust | Formed upon retreat | 40m | Bennett et al., 1996 |
| Iceland | LIA | Glaciotectonic | Fold and thrust | 2-6 days | 10-40m | Benediktsson et al., 2010 |
| Baffin I., Canada | Neoglacial | Push | Pushed outwash gravels | 1 yr | 40m | Boulton et al., 1986 |
| Iceland | LIA | Push | Single large nappe and faulting | <39 likely 1 or 2 yrs | 8m, 35m | Bennett et al., 2004 |
| Svalbard | LIA | Push/Thrust | Surge | <1 yr | >30m | Hart & Watts, 1997 |
| Svalbard | LIA | Ice Cored | Retreating from LIA maximum | Formed upon retreat | 25-30m | Lyså & Lønne, 2001 |
| Iceland | LIA | Glaciotectonic | Fold and thrust | 12 yrs at terminus | 25-30m | Bennett et al., 2000 |
| Norway | Modern | Push? | 2 year advance | 2 years | 20m | Benedict et al., 2013 |
| Svalbard | LIA | Push | 1882/1886 Surge | <1 yrs | 1-20m | Boulton et al., 1996 |
| New Zealand | LGM | Push/Thrust | | Likely <30 yrs | 10-15m | Hart, 1996 |
| Iceland | Modern | Push | Imbricate | 1 yr | 1-10m | Humlum, 1985 |
| Alaska | Modern | Push | | Sustained adv. | 10m | Motyka & Echelmeyer, 2003 |
| Norway | LIA/ modern | Push | Bulldozing and thrusting | 1-10 yrs | 3-8m | Burki et al., 2009 |
| Yukon,Canada | LIA/ modern | Ice Cored | Debris thickness reported | | 1-6m | Johnson, 1972 |
| Iceland | Modern | Push | Polygenetic push | Seasonal | .4-5.25m | Sharp, 1984 |
| Iceland | LIA | Push/Thrust | 1890 Surge | 1 day | 5m | Benediktsson et al., 2008 |
| Iceland | Modern | Push | Annual moraines | Seasonal | 4m | Sharp, 1984 |
| Iceland | Modern | Push/Basal Freezing | | 1 year | 3.5-4m | Krüger, 1993 |
| Iceland | Modern | Push | | 1 year | 3.5m | Krüger, 1993 |
| Norway | Modern | Push | From six separate glaciers | 1 year sustained adv | 1-3m | Winkler & Matthews, 2010 |
| Alaska | Modern | Ice Cored/ Push | Readvance in a surging glacier | 1 year | 3m | Johnson, 1971 |
| Iceland | Modern | Ice Cored | Debris thickness reported | 1 year | 1-3m | Krüger & Kjær, 2000 |
| Argentina | Modern | Push | | 1 yr | 2.5m | Rabassa et al., 1979 |
| Iceland | Modern | Push | | Seasonal | 1-2m | Boulton, 1986 |
| Switzerland | Modern | Push | Annual winter advances | Seasonal | <1.5m | Lukas, 2012 |
| Norway | Modern | Push | Annual winter advances | Seasonal | <1m | Benedict et al., 2013 |
| Iceland | Modern | Push | Lodgement freeze on | Seasonal | .3 -.7m | Krüger, 1995 |
| New Zealand | Modern | Ice Cored | Debris thickness reported | 1-2 years | .4m | Brook & Paine, 2011 |
| Region | Time period | Type | Sub-Category | Formation Time | Height | Reference |
| ***Moraines dependent on terminus standstills*** | | | | | | |
| Colorado, USA | LGM | Ice Contact Fan | Debris flow and alluvium | <20 years | 25m | Johnson and Gillham, 1995 |
| France | LIA | Ice Contact Fan/Dump | Formed over 3 advances | ~10 years | 20m | Nussbaumer & Zumbühl, 2012 |
| Scotland, UK | Younger Dryas | Ice Contact Fan/push | Debris flow and alluvium | 3-9 or 7-19 years | 15m | Benn & Lukas, 2006 |

| | | | | | | |
|---|---|---|---|---|---|---|
| Iceland | Modern | Ice Contact Fan | Outwash fans/ sandar | <10years | 10m | Boulton, 1986 |
| Iceland | Modern | Dump/push | Initially push | 7 years sustained adv. | 4-7m | Krüger et al., 2002 |

* LIA refers to the Little Ice Age
** LGM refers to the Last Glacial Maximum

*Moraines independent of glacial standstills*

The most rapidly forming moraines require the propagation of debris in front of an advancing ice front (e.g. Krüger, 1995; Benediktsson, et al., 2010; Benediktsson et al., 2008; Boulton, 1986; Humlum, 1985). Because the material is bulldozed or thrust in front of the glacier, the moraine can form during any advance and retreat cycle irrespective of time spent in standstill. The formation of glaciotectonic and push moraines is more dependent on the availability of sediment or permafrost in the foreland than it is on the glaciological conditions (Bennett, 2001). *Glaciotectonic Moraines* are formed when the stress imposed by an advancing glacier excavates and elevates (associated with thrusting and folding) proglacial bedrock and/or quaternary sediments. *Push Moraines* are formed by the bulldozing of proglacial sediment and typically have steep proximal and gentle distal slopes. Advances over long distances can result in formation of a more extensive set of moraine ridges. *Hummocky and ice-cored moraines* form when heavily debris-covered ice is dynamically separated from an active, retreating glacier (Lyså and Lonne, 2001). As these moraines are in place as soon as the ice is dynamically separated from the active glacier, all that remains is for the ice core to waste away. Ice-cored and hummocky moraines do not require a glacial standstill to form (Glassner and Hambrey, 2002; Johnson, 1972) and their identification in the moraine record implies that the moraine was emplaced instantaneously for the timescales of interest for this study.

*Moraines dependent on glacial standstills*

Latero-frontal fans and dump moraine sizes are dependent on the debris flux off the glacier and the length of time the glacier terminus remains stationary. A glacier that advances and retreats without a terminus standstill will not likely form an ice-contact fan or a dump moraine, although there are reported occurrences in the literature. One of these potential influences is thick supraglacial debris-cover, which can slow terminus oscillations and provide the debris fluxes to create large moraines. *Ice-contact fans* form by the coalescence of debris fans and glaciofluvial processes at the glacier terminus. Although latero-frontal fans can form over short periods and even in a single short-lived advance, these fans are typically on the order 10 meters in height whereas fans that limit subsequent ice advances are typically 100s of meters in height (Benn and Lukas, 2006; Benn and Evans, 1998). *Dump Moraines* are formed by the delivery of supraglacial material derived from rockfall onto the glacier or the melt out of basal debris septa that flows or falls off the terminal ice slope. Paleoglacier valleys with large ice-contact fans (>100 m in height) or dump moraines should be treated with more caution than moraines that are independent of glacial standstills. Nearly all documented terminal moraine formation durations are less than 20 years (Table DR.3). Further sedimentological and stratagraphic investigation of LGM terminal moraines is needed to constrain the importance of moraine formation timescale on paleoclimate reconstruction (e.g., Johnson and Gillam, 1995).

*Terminal moraines do not limit subsequent advances*

We have assumed that the furthest length excursion from the mean glacier length forms the maximum terminal moraine. In effect, this requires that that previously formed moraines do not limit the extent of subsequent advances. The only moraine types that have been shown to limit subsequent advances are large latero-frontal moraines or scree aprons; these are common in tectonically active regions such as the Himalaya or the Andes. These moraines can become sufficiently massive to dam glacier ice and cause subsequent glacial advances to terminate at the same location (Lliboutry, 1977; Thorarinsson, 1956). This effect is especially apparent where large lateral moraines are deposited outside of cirques and steep valleys and are therefore less susceptible to paraglacial processes (Thorarinsson, 1956). Cases where latero-terminal moraines could have limited ice extent are easily identifiable by the height and extent of the latero-frontal moraines. These situations are uncommon in LGM terminal moraines in the Western US.

*Overridden terminal moraines are destroyed*

Moraines can be overrun by subsequent advances and still be identifiable upon retreat (Karlén, 1973; Bennett et al., 2000). Overrun moraines may be differentiated from moraines that haven't been overrun by their subdued topography compared to moraines down valley, the presence of fluted till overriding the moraine, and the presence of lateral continuations of the moraine that have not been overridden that exhibit a sharper morphology (Karlén, 1973). Preservation of overrun moraines is rare and the potential for preservation depends on the local bedrock topography and the amount of time the overrun moraine is subjected to subglacial processes. An overrun moraine could potentially pose a problem for paleoclimatic or mean glacial length reconstruction only if a moraine is overrun and there is no indication of the maximum extent of the overriding glacier. The overrun moraine would then be interpreted as the maximum extent of the glacier for the time period of interest and could produce substantial error. This situation is unlikely for LGM moraines, as any overrun moraine would have been smoothed by the overriding glacier and then subjected to at least 10 thousand years of diffusional surface process that would further obliterate the morainal form.

***Section DR 4. LGM moraine complexes***

LGM 'terminal moraines' in the western US are often composed of a conglomeration of moraines formed during numerous glacier advances. We call these clusters of moraines, terminal moraine complexes keeping in mind that it is possible that these clusters of ridges were formed by a single advance and the individual moraines interpreted as terminal moraines are actually fault bend folds from a glaciotectonic push moraine. Below in figure B we present a LiDAR hillshade of the Teton Glacier LGM terminal moraine and our interpretation of distinct terminal moraines and the subjective limits of the terminal moraine complex. This hillshade allows us to define many more ice marginal features than possible without detailed field surveys. The terminal moraines defined in figure B are likely formed between the LGM mean length and the maximum terminal moraine (labeled 1) and are therefore likely candidates for moraines formed by glacier length fluctuations driven by interannual variability.

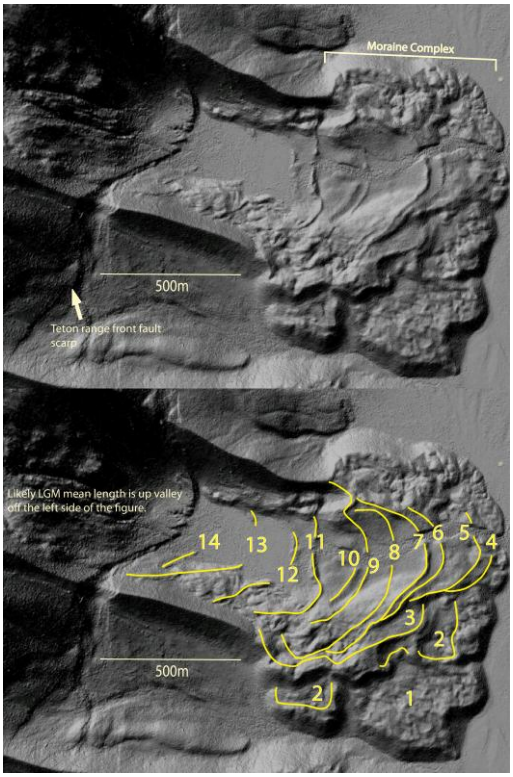


Figure DR.4 LiDAR of the LGM Teton glacier terminal moraines. In the bottom panel we show what we interpret to be 14 distinct ice margins revealed by the LiDAR. The LiDAR is courtesy of OpenTopography.

***DR.5 Relative sensitivity of length fluctuations due to temperature and precipitation variability***
Roe and O'Neal (2009) show that the relative sensitivity of a glacier's fluctuations to temperature vs. precipitation variability is given by:

$$R = \frac{A_{T>0}\mu\sigma_T}{A_{tot}\sigma_P} \quad .$$

The $R$ values for Front Range glaciers greater than 4 km$^2$ range between 2.2 and 2.9 with a mean of 2.5, suggesting that year-to-year variations in summer temperature were two to three times as important for driving length perturbations as were variations in annual precipitation. This dominant sensitivity to summertime temperature variation is expected in continental climates.

***Section DR.6 Flowline model description***

We follow standard equations for the shallow-ice-approximation incorporating glacier sliding (e.g., Oerlemans, 2001):

$$\frac{dH(x)}{dt} = \dot{b}(x) - \frac{dF(x)}{dx}\ ;\ F(x) = \rho^3 g^3 \left( f_d H(x)^2 + f_s \right) H(x)^3 \left( \frac{dz_s}{dx} \right)^3 , \qquad (1)$$

where $H(x)$ is ice thickness at position $x$, $\dot{b}(x)$ is the local net mass balance, $F(x)$ is the depth-integrated ice flux, $\rho$ is ice density, $g$ is the acceleration due to gravity, $dz_s/dx$ is the local ice surface slope, $f_d$ = 1.9 x $10^{-24}$ Pa$^3$ s$^{-1}$ and $f_s$ = 5.7 x $10^{-20}$ Pa$^3$ m$^2$ s$^{-1}$ (the coefficients of deformation and sliding).

***Section DR.7 Model discussion***

By exploring a very wide parameter space, we have constrained the effects of interannual variability on glacial length and moraine formation over extreme bounds. The range of parameter uncertainty could be better constrained by examining how the climate parameters vary in space from the LGM to the present. The most uncertain climate parameters, $\Gamma$, $\sigma_P$, and $\sigma_T$, could be better constrained by using atmospheric circulation model output, and better minimum estimates of $D$ could be obtained by reducing the uncertainty in moraine-derived dates. It should also be determined if using higher order ice physics models changes the effects of interannual variability on glacier length, although we anticipate that parameter uncertainty will swamp any differences between models. In the climate forcing presented here, we have assumed that $T$ and $P$ are uncorrelated from year-to-year (white noise), as is generally the case for centennial-scale instrumental observations of $T$ and $P$ and glacier mass balance records (e.g., Burke and Roe, 2013); on longer time-scales, paleoclimate records show that a degree of persistence (correlation from year-to-year) does exist (e.g., Huybers and Curry, 2006). Even a small degree of persistence can substantially increase the magnitude of fluctuations (e.g. Reichert et al., 2001). For this reason and others outlined in Roe and O'Neal (2009), we feel that our estimates of the fluctuation of glacier length about the mean length are conservative.

### *Section DR. 8 Explanation of Interannual Variability*

Interannual climate variability refers to changes in the mean value of climate parameters (air temperature, precipitation, etc.) from year-to-year. Think of last year's mean summer temperature compared to this year's mean summer temperature (same can be done for total winter precipitation or annual precipitation). The variation from one year to the next is what we refer to as interannual variability. When long records of mean summer temperature (or annual precip) are tested for year-to-year correlation (if we have a warm summer relative to the long term mean this year are we more likely to have a warm summer next year?) there is little evidence of correlation (Burke and Roe, 2013). Interannual records of summer temperature show very little or no correlation from one year to the next and are best represented as white noise (equal power at all frequencies). The change in weather from year-to-year is not considered a climate change so glacier fluctuations forced by interannual variability are independent of climate change.

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