The effects of interannual climate variability on the moraine record

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ABSTRACT

Glacial moraines are commonly used to infer mean climate conditions (annual precipitation and melt-season temperature) at the time of moraine formation. However, recent research has demonstrated that substantial fluctuations in glacier length also occur in response to stochastic, year-to-year variability in mass balance. When interpreting moraine sequences it is important to differentiate between moraines that reflect the signal of actual climate changes versus those that may reflect the noise of interannual climate variability. We address this issue for the glaciers of the Colorado Front Range, USA. Using a linear glacier model that allows for a 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) glaciers smaller than 6 km² were likely transient features, coming and going due to interannual variability.
INTRODUCTION

Glacial to interglacial changes in the long-term averages of annual precipitation ($P$) and melt-season temperature ($T$) are sufficient, in many places, to drive large changes in glacier length, as recorded by the emplacement of terminal moraines. But the natural, year-to-year (interannual), fluctuations of $P$ and $T$ have also been shown to force kilometer-scale glacial length fluctuations, due solely to the random alignment of years of negative and positive mass balance, even in steady climates (e.g., Oerlemans, 2001; Reichert et al., 2002; Roe and O’Neal, 2009). A steady climate implies constant long-term averages ($= \bar{P}, \bar{T}$) and, importantly, constant standard deviations ($= \sigma_P, \sigma_T$). Interannual variability is due to the stochastic fluctuations of weather and internal modes of variability such as the North Atlantic, the Pacific/North American, and the El Niño-Southern Oscillations. The amplitude of interannual variability varies with location and climate state, but it is always present. If we are to accurately constrain the values of past $\bar{T}$ and $\bar{P}$ using glacier moraines we must understand the effects of this unavoidable year-to-year climate 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 $\bar{P}$ but that also includes interannual variability. The glacier of case (b) will produce a terminus history that will fluctuate around the same $\bar{L}$ as glacier (a) (Fig. 1B; Fig. 2B). 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) glacial length maxima represent potential moraine-forming locations, ii) terminal moraines can be formed on 10-20 year timescales (see DR.3), iii) terminal moraines do not limit the extent of subsequent advances, and iv) all moraines that are overrun by subsequent advances
are removed, then the maximum excursion from $\bar{L}$, will form the furthest terminal moraines (see DR.3 for a detailed discussion). Estimates of average climate (i.e., $\bar{P}$, $\bar{T}$) should be based on $\bar{L}$ rather than $L_{\text{max}}$ (Fig. 1B). Thus, we face the challenge of estimating $\bar{L}$ knowing only the glacier geometry preserved by the $L_{\text{max}}$ advance, while also accounting for the substantial uncertainties in the mass balance, geometry, and glacier parameters that pertained in past climates.

We focus on the last glacial maximum (LGM) moraine record in the Colorado Front Range, USA, because of well-dated, nested sets of near-LGM moraines, diverse glacier geometries, and the availability of long-term, high-elevation modern meteorological records. We first employ a dynamic flowline model to confirm, in accordance with prior work in other settings, that interannual variability can drive significant length fluctuations. The focus however is on using a linearized model to more efficiently explore a wide range of parameter uncertainty.

MODEL DESCRIPTIONS

Flowline Model

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

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

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, $f_d = 1.9 \times 10^{-24} \text{Pa}^3 \text{s}^{-1}$ and $f_s = 5.7 \times 10^{-20} \text{Pa}^3 \text{m}^2 \text{s}^{-1}$. 

Linearized Model

The linear model used in this study considers the conservation of mass, neglects ice dynamics, and, in a steady climate, considers length variations as departures from a mean length $\bar{L}$ that are small enough that the system is linear (Roe and O’Neal, 2009). Using finite differences to discretize the governing equation and using a time step $\Delta t$ of 1 year, the response of the glacier to stochastic climate variability is

$$L_{t+\Delta t} = L_t \left(1 - \frac{\Delta t}{\tau}\right) + \beta \sigma_p v_t + \alpha \sigma_t \lambda_t,$$

where subscript $t$ denotes the year; $v_t$ and $\lambda_t$ are independent normally distributed random processes; and $\alpha$ and $\beta$ are coefficients that are functions of glacier geometry:

$$\alpha = -\frac{\mu A_T \sigma_0}{wH}, \quad \beta = \frac{A_{	ext{rot}}}{wH},$$

where,

$$A_{T>0} = A_{abl} + \frac{P_W}{\mu \Gamma \tan \phi}.$$

Parameter definitions are given in Table 1. The characteristic timescale $\tau$ for the linear model is

$$\tau = \frac{wH}{\mu \Gamma \tan \phi A_{abl}},$$

the time over which the glacier ‘remembers’ past $T$ and $P$ states. Derivation of these governing equations is given in the appendix of Roe and O’Neal (2009).

Climate Data and Parameter Selection

Meteorological data were extracted from the longest-running high-elevation weather station in North America on Niwot Ridge, CO (see Figure 3B; Station D1; 1952-2010; 3743m a.s.l.). Assuming that the melt season runs from June to September, we determined that $(\bar{T}, \sigma_T)$ is (6.3, 1.3) °C and for annual precipitation $(\bar{P}, \sigma_P)$ is (1.2, 0.22) m a⁻¹. Data were linearly
detrended. While the modern summertime near-surface lapse rate in the ice-free Boulder Creek
catchment is -5.6 °C km\(^{-1}\) (Pepin and Losleben, 2002), we consider a range 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 per 1°C
change in \(T\)), based on a global compilation of \(\mu\) for snow (4.5±1.7 mm °Cday\(^{-1}\)) and ice
(7.7±3.2 mm °Cday\(^{-1}\)) (see 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).

We define climate as the statistics of weather, averaged over any interval relevant for the
question at hand (typically longer than 30 years). The duration of the regional LGM climate is
defined by the parameter \(D\), for which we consider a broad range between 500 and 7500 years,
the upper limit being the duration of the LGM sea level low stand (26.5 to 19 ka; 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 climate variability.

**IMPACT OF INTERANNUAL VARIABILITY ON MEAN GLACIER LENGTH**

The flowline model (eq. 1) was integrated with mid-range parameters (Table 1) for the
basal slope and length of Fall Creek glacier (Table 2). The most likely parameter set generated a
standard deviation of length fluctuations (\(\sigma_L\)) of 370 m. Sample model output is shown in Fig.
2C. For the linear model eqs. 3\&4 can be solved exactly for \(\sigma_L\)

\[
\sigma_L = \sqrt{\frac{\tau \Delta t}{2}} \sqrt{\alpha^2 \sigma_T^2 + \beta^2 \sigma_P^2}.
\]  

(7)
For the same parameters as the flowline model, the linear model gives $\alpha_L = 415$ m, consistent with the ~15% overestimate described and explained by Roe (2011). This inter-model difference is much smaller than the uncertainty range of the parameters (Table 1). The outer bounds of $\alpha_L$ for the linear model are 180 and 800 m, representing cases in which all parameters are simultaneously given their extreme values. Beyond its simplicity, a clear value of the linear model, as seen in eq.7, is that parameter uncertainty can be straight-forwardly propagated: a high sensitivity to $T$ or $P$ (i.e., large $\alpha$ or $\beta$), or a long response time leads to a large $\alpha_L$.

Roe (2011) derived excursion statistics for glaciers driven by climate variability (following standard texts, e.g., Vanmarcke, 1983; see also Reichert et al. 2002), which we adapt here. In any given time interval $D$ the maximum excursion, $L_{\text{max}}$ cannot be known exactly, but it is described by a probability distribution. Roe (2011) showed that the most likely $L_{\text{max}}$ can be related to the average position, $\bar{L}$, by

$$
\bar{L} = L_{\text{max}} - \sigma_L \sqrt{2 \ln \left( \frac{D\sigma_L}{2\pi\sigma_L \ln(2)} \right)},
$$

(8)

where $\sigma_L$ is the standard deviation of the time rate of change of glacier length (see eq. 9 and A8 in Roe (2011) with $p = 0.5$). Equation 8 is quite general, and holds provided the probability distribution of glacier fluctuations is normally distributed (Fig. 1B), and has been shown to govern the variability of terminus position in flowline glacier models (see Reichert et al., 2002; Roe, 2011). Roe (2011) further demonstrated that setting $\sigma_L = \sigma_L \sqrt{2/\psi \tau \Delta t}$ emulated the behavior of a standard numerical flowline model, where $\psi$ ($\approx 10$) is a factor introduced because high frequencies are damped more strongly in a numerical flowline model than is predicted by the linear model. Because $\psi$ is passed through two square roots and a natural log in eq. 8, the
effect of varying this parameter is minimal. A doubling or halving of $\psi$ results in a $\pm 0.6\%$
change in $\bar{L}$ when $D$ is larger than the glacier response time, $\tau$.

Mean Length and Signal-to-Noise Results

Table 2 shows the linear model mean-length results for eleven LGM glaciers from the
Colorado Front Range. For glaciers with area greater than 6 km$^2$, the most likely mean glacier
length is 10-15% back from $L_{\text{max}}$. Bounds on the potential location of the mean length—3% to
50% — represent cases in which all parameters are simultaneously given their extreme values.
As this is unlikely, the resulting range should be considered very much an outer bound on the
mean length.

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; eq. 8) and a noise
component (climate noise $\sigma_P, \sigma_T$ driving length fluctuations, $\alpha_L$) (Fig. 1B; eq. 7). Comparison of
the $\bar{L}$ and the $\alpha_L$ components for modeled LGM glaciers with area greater than 6 km$^2$ shows that
it is likely that larger glaciers persisted throughout the local LGM, as their $\bar{L}$ was much larger
than their $\alpha_L$ (Table 2). Glaciers with area less than 6 km$^2$ likely flickered in and out of existence
due to the interannual variability of the (colder) LGM climate, as $\bar{L}$ was of comparable size to
$\alpha_L$.

DISCUSSION

Moraines reflect maximum advances, and our results suggest that climate noise can drive
such advances substantially beyond what the mean climate conditions would support (Fig 1B;
Fig. 2D). Climate estimates derived from maximum glacier geometries do not represent the local
LGM-mean climate. Rather, they have a one-sided bias due to stochastic excursions away from
the mean. Equilibrium-line-altitudes (ELAs) and climate change estimates 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. Because moraines reflect maximum advances, dating the maximum terminal moraine provides a minimum estimate of when a climate change initiated. Furthermore, constraining the timing of retreat from the estimated mean length provides an estimate of when a climate change terminated (e.g., Young et al., 2011).

The role of interannual variability in driving stochastic glacier length fluctuations is important to consider for two additional reasons: the formation of LGM nested moraine sequences in individual valleys; and the asynchronous timing of LGM advances across the western US. Broad LGM terminal moraine complexes (sets of closely spaced or nested terminal moraines in a single valley; Fig. 1B; Fig. 2C; Fig. 3C; DR.4) possibly represent cycles of kilometer-scale retreat and re-advance that are independent of true climate change. It follows that such a glacier likely formed one moraine, then left the terminal moraine complex and returned, forming a second moraine. Terminal moraines between $\bar{L}$ and $L_{\text{max}}$ are therefore not likely ‘recessional’ moraines in the classic sense. If we expect 10-15% advances from the mean glacial length, we should also expect 10-15% retreats (Fig. 1B; Fig. 2C) that are unlikely to form moraines. The potential for exposure of bedrock between $L_{\text{max}}$ and $L_{\text{min}}$ during these cycles could impact the interpretation of cosmogenic radionuclide-derived bedrock exposure ages.

It is also possible that the natural variation of glacial length about its mean could explain the spread of ages derived from LGM maximum terminal moraines across the Western US (Fig. 2D; see Young et al., 2011; Thackray, 2008; Licciardi et al., 2004 for other interpretations), and even globally (Schäfer et al., 2006). Incoherent patterns of interannual variability from region-to-
region could have resulted in glacier advances and retreats at different times around the Western US during the regional LGM. Modern snowpack snow-water-equivalent anomalies support the above hypothesis, as they are coherent over a several hundred-kilometer length scale that is significantly smaller than the glaciated area within the Western US (Cayan, 1996; Huybers and Roe, 2009).

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, $T$, $\sigma_P$, 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.
CONCLUSIONS

We have assessed the effects of interannual variability on glacier length and its associated uncertainty in a continental climate during the LGM. Interannual variability inevitably results in glacial length fluctuations around a mean glacier 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%. Glacier length fluctuations forced by interannual variability can be viewed as a smaller scale example of the moraine survival problem, e.g. Gibbon et al. (1984), with consequences for the interpretation of moraine complexes. Both standard flowline and linear model results support the fundamental conclusions of this study. Glacier length fluctuations due to year-to-year climate variability should therefore be included in the interpretation of the moraine record in order to realize the record’s full potential.

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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 (grey 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 grey dashed ice profiles. Given enough time, the terminus position fills a normal distribution centered at $\bar{L}$ (the signal; eq. 8) and is characterized by $\sigma_L$ (the noise; eq. 7), 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. Eq. 8 gives the most likely mean length from this pdf. (B) An example melt-season climatology developed from a normally distributed random process, 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. The bold-dashed line is the most likely mean length from eq. 8. The grey
dashed line shows the mean length from the example terminus history derived from flowline model output using the same parameters as in (A). (D) Glacier length normalized to the LGM maximum terminal moraine with terminal moraines in-board from the maximum extent shown as triangles for western US LGM glacial 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 US mountain 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. $L_{\text{mean}}$, the most likely estimate of mean length, and $L_{\text{max}}$, 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 and are derived from Young et al. (2011) and references within. Numbers near moraines denote the number of distinct ice advances recorded in the pro-$L$ area.
TABLE 1. LINEAR MODEL PARAMETERS AND GEOMETRY INPUTS

<table>
<thead>
<tr>
<th>Name</th>
<th>Units</th>
<th>Description</th>
<th>Min.</th>
<th>Mean</th>
<th>Max.</th>
</tr>
</thead>
<tbody>
<tr>
<td>(\mu)</td>
<td>(m (C) yr(^{-1}))</td>
<td>Melt factor</td>
<td>.5</td>
<td>.7</td>
<td>.9</td>
</tr>
<tr>
<td>(\Gamma)</td>
<td>(C km(^{-1}))</td>
<td>On-ice near-surface lapse</td>
<td>3.5</td>
<td>5</td>
<td>6.5</td>
</tr>
<tr>
<td>AAR</td>
<td></td>
<td>Accumulation-area ratio</td>
<td>.5</td>
<td>.65</td>
<td>.8</td>
</tr>
<tr>
<td>(\bar{P})</td>
<td>(m)</td>
<td>Mean annual precipitation</td>
<td>.6</td>
<td>1.2</td>
<td>2.4</td>
</tr>
<tr>
<td>(c_T)</td>
<td>(\degree C)</td>
<td>Std. of summertime temp.</td>
<td>1.0</td>
<td>1.3</td>
<td>1.6</td>
</tr>
<tr>
<td>(c_P)</td>
<td>(m a(^{-1}))</td>
<td>Std. of annual precipitation</td>
<td>.11</td>
<td>.22</td>
<td>.44</td>
</tr>
<tr>
<td>(D)</td>
<td>(yr)</td>
<td>Duration of climate change</td>
<td>500</td>
<td>4000</td>
<td>7500</td>
</tr>
</tbody>
</table>

**Geometry Inputs**

- \(A_{\text{tot}}\): Total area of the glacier
- \(A_{\text{abl}}\): Ablation area of the glacier
- \(\tan\theta\): Slope of the glacier bed
- \(w\): Width of the ablation zone
- \(H\): Thickness of the glacier

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

<table>
<thead>
<tr>
<th>Glacier name</th>
<th>Area (km(^2))</th>
<th>Slope</th>
<th>Width (km)</th>
<th>Height (km)</th>
<th>(L_{\text{max}}) (km)</th>
<th>(\Gamma_{\text{max}}) % to (L_{\text{max}})</th>
<th>(\Gamma_{\text{mean}}) % to (L_{\text{max}})</th>
<th>Mean Signal-to-Noise</th>
<th>(\Gamma_{\text{max}}) % to (L_{\text{max}})</th>
<th>Mean Signal-to-Noise</th>
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<tbody>
<tr>
<td>1. Middle Boulder</td>
<td>56.62</td>
<td>0.031</td>
<td>1.30</td>
<td>0.22</td>
<td>18.5</td>
<td>0.97</td>
<td>40.22</td>
<td>0.86</td>
<td>16.68</td>
<td>0.62</td>
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<tr>
<td>2. North Saint Vrain</td>
<td>45.89</td>
<td>0.050</td>
<td>1.16</td>
<td>0.19</td>
<td>15.2</td>
<td>0.97</td>
<td>45.19</td>
<td>0.88</td>
<td>19.63</td>
<td>0.68</td>
</tr>
<tr>
<td>3. Bear Lake</td>
<td>31.19</td>
<td>0.055</td>
<td>1.56</td>
<td>0.16</td>
<td>12.6</td>
<td>0.97</td>
<td>43.11</td>
<td>0.87</td>
<td>17.70</td>
<td>0.63</td>
</tr>
<tr>
<td>4. North Boulder</td>
<td>26.00</td>
<td>0.091</td>
<td>1.32</td>
<td>0.11</td>
<td>12.4</td>
<td>0.97</td>
<td>52.51</td>
<td>0.89</td>
<td>22.94</td>
<td>0.71</td>
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<td>5. Fall Creek</td>
<td>14.57</td>
<td>0.078</td>
<td>0.58</td>
<td>0.14</td>
<td>10.5</td>
<td>0.96</td>
<td>41.52</td>
<td>0.86</td>
<td>17.86</td>
<td>0.65</td>
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<tr>
<td>6. Hunter's Creek</td>
<td>6.25</td>
<td>0.138</td>
<td>0.81</td>
<td>0.09</td>
<td>6.19</td>
<td>0.96</td>
<td>39.80</td>
<td>0.85</td>
<td>16.12</td>
<td>0.59</td>
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<td>7. Mill Creek</td>
<td>5.80</td>
<td>0.089</td>
<td>0.52</td>
<td>0.11</td>
<td>5.94</td>
<td>0.95</td>
<td>27.17</td>
<td>0.79</td>
<td>10.27</td>
<td>0.41</td>
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<tr>
<td>8. Roaring Fork</td>
<td>4.11</td>
<td>0.178</td>
<td>0.51</td>
<td>0.08</td>
<td>5.72</td>
<td>0.96</td>
<td>42.22</td>
<td>0.86</td>
<td>17.69</td>
<td>0.63</td>
</tr>
<tr>
<td>9. Silver Creek</td>
<td>1.67</td>
<td>0.131</td>
<td>0.45</td>
<td>0.04</td>
<td>2.07</td>
<td>0.83</td>
<td>7.51</td>
<td>0.28</td>
<td>1.11</td>
<td>-1.19</td>
</tr>
<tr>
<td>10. Rainbow Creek</td>
<td>1.42</td>
<td>0.120</td>
<td>0.44</td>
<td>0.04</td>
<td>2.92</td>
<td>0.87</td>
<td>9.95</td>
<td>0.43</td>
<td>2.11</td>
<td>-0.77</td>
</tr>
<tr>
<td>11. Horseshoe</td>
<td>1.41</td>
<td>0.137</td>
<td>0.35</td>
<td>0.05</td>
<td>2.86</td>
<td>0.90</td>
<td>13.24</td>
<td>0.57</td>
<td>3.64</td>
<td>-0.30</td>
</tr>
</tbody>
</table>

*Signal-to-noise ratio \(\equiv \frac{\Gamma}{\sigma_L}\)

\(^1\)GSA Data Repository item 2013xxx, Sections DR.1-DR.5 are available online at www.geosociety.org/pubs/ft2009.htm, or on request from editing@geosociety.org or Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301, USA.
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Leif S. Anderson, Gerard H. Roe, and Robert S. Anderson

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 (=measurements only on the glacier surface), near-surface (=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>
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<tr>
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<th>Classification</th>
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<th>Lapse Rate °C km$^{-1}$</th>
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Valley Glacier Mean: 4.7 ± 2.2 w/ 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 B). 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 B shows that the most likely positive degree day factor for ice: is $7.7 \text{ m } \text{oC}^{-1} \text{ a}^{-1}$ with a $1\sigma$ value of $3.2 \text{ m } \text{oC}^{-1} \text{ a}^{-1}$ with extreme values of $20 \text{ m } \text{oC}^{-1} \text{ a}^{-1}$; Van de Wal (1992) and $2.6 \text{ m } \text{oC}^{-1} \text{ a}^{-1}$; Zhang et al. (2006); and the most likely positive degree day factor for snow is $4.5 \text{ m } \text{oC}^{-1} \text{ a}^{-1}$ with a $1\sigma$ value of $1.7 \text{ m } \text{oC}^{-1} \text{ a}^{-1}$ with extreme values of $11.6 \text{ m } \text{oC}^{-1} \text{ a}^{-1}$; Kayastha et al. (2000) and $1.4 \text{ m } \text{oC}^{-1} \text{ a}^{-1}$ Howat et al. (2007). It is important to note that our parameter combinations produce mass balance values that are reasonable for continental climates.

<table>
<thead>
<tr>
<th>Ice Sheet and Ice Cap</th>
<th>Location</th>
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<td>Marshall et al., 2007</td>
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**Ice Sheet and Ice Cap Mean:** 5.1±1.2

* Debris-covered glacier

*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
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<td>Summer 1963</td>
<td>Braithwaite, 1981</td>
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TABLE B. GLOBAL COMPILATION OF POSITIVE DEGREE-DAY MELT FACTORS
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<th>Duration</th>
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<td>Aug 1982-1983</td>
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<td>5.9</td>
<td>5</td>
<td>4800</td>
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<td>26 Jun-11 Jul 1982</td>
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<td>13.2</td>
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<td>12</td>
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<td>5.9</td>
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<td>36°N</td>
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<td>3.6</td>
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<td>36°N</td>
<td>14 Jun-27 Jun 1981</td>
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<td>Xiaodongkemadi, Tanggula, China</td>
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<td>7.2</td>
<td>5405-5475</td>
<td>39°N</td>
<td>Jul-Aug 1993</td>
<td>Zhang et al., 2005</td>
</tr>
<tr>
<td>Qiyi, Qilian Shan, China</td>
<td>9</td>
<td>9</td>
<td>5700-6000</td>
<td>39°N</td>
<td>Jul-Aug 2002</td>
<td>Zhang et al., 2005</td>
</tr>
<tr>
<td>Kangwure, Himalaya, China</td>
<td>5.2</td>
<td>5.2</td>
<td>8.4</td>
<td>42°N</td>
<td>Summer</td>
<td>Cui, 2009</td>
</tr>
<tr>
<td>Location</td>
<td>Snow</td>
<td>Ice</td>
<td>Elevation (m)</td>
<td>Latitude</td>
<td>Duration</td>
<td>Reference</td>
</tr>
<tr>
<td>----------------------------------</td>
<td>------</td>
<td>-----</td>
<td>---------------</td>
<td>----------</td>
<td>---------------------------</td>
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<tr>
<td>Gooseberry Creek, Utah, USA</td>
<td>2.5</td>
<td>2650</td>
<td>~38°N</td>
<td>23 Apr-9 May 1928</td>
<td>Clyde, 1931</td>
<td>Hock, 2003</td>
</tr>
<tr>
<td>Weissfluhjoch, Switzerland</td>
<td>4.5</td>
<td>2540</td>
<td>46°48’N</td>
<td>Several seasons</td>
<td>Zingg, 1951</td>
<td>Hock, 2003</td>
</tr>
<tr>
<td>3 basins in USA</td>
<td>2.7</td>
<td></td>
<td></td>
<td>C. of Engineers, 1956</td>
<td>Hock, 2003</td>
<td></td>
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<tr>
<td>3 basins in USA</td>
<td>4.9</td>
<td></td>
<td></td>
<td>C. of Engineers, 1956</td>
<td>Hock, 2003</td>
<td></td>
</tr>
<tr>
<td>Former European USSR</td>
<td>5.5</td>
<td>7</td>
<td>1800-3700</td>
<td>Several seasons</td>
<td>Kuzmin, 1961, p. 117</td>
<td>Hock, 2003</td>
</tr>
<tr>
<td>Non-glaciated site means:</td>
<td>4.0±1.2</td>
<td>7.0</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Mean melt factor for all examples in the literature: 4.5±1.7 7.7±3.2

Non-glaciated Sites

Central Asia means: 5.4±2.3 7.9±3.6
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 (<20 years; a time scale supported by flowline modeling (see Roe, 2011 Figure 4)), we present a review of the moraine sedimentological literature (Table C), which shows that the majority of moraines with constrained formation periods form over periods less than 20 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. 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 (<20 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).

<table>
<thead>
<tr>
<th>Region</th>
<th>Time period</th>
<th>Type</th>
<th>Sub-Category</th>
<th>Time of formation</th>
<th>Height</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>Iceland</td>
<td>Modern</td>
<td>Push</td>
<td>Imbricate</td>
<td>1 year</td>
<td>3-5m</td>
<td>Humlum, 1985</td>
</tr>
<tr>
<td>Iceland</td>
<td>Modern</td>
<td>Push</td>
<td>Lodgement freeze on</td>
<td>Seasonal</td>
<td>.3 -.7m</td>
<td>Krüger, J., 1995</td>
</tr>
<tr>
<td>Iceland</td>
<td>LIA</td>
<td>Push</td>
<td>Fold and thrust</td>
<td>2-6 days</td>
<td></td>
<td>Bennediktsson et al., 2010</td>
</tr>
<tr>
<td>Iceland</td>
<td>LIA</td>
<td>Push</td>
<td>Fold and thrust</td>
<td>12yr at terminus</td>
<td></td>
<td>Bennett et al., 2000</td>
</tr>
<tr>
<td>Iceland</td>
<td>LIA</td>
<td>Push</td>
<td>Single large nappe and faulting</td>
<td>&lt;3yr likely 1 or 2 yrs</td>
<td>8m</td>
<td>Bennett et al., 2004</td>
</tr>
<tr>
<td>Iceland</td>
<td>Modern</td>
<td>Push</td>
<td></td>
<td>Seasonal</td>
<td>1-2m</td>
<td>Boulton, 1986</td>
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<tr>
<td>Norway</td>
<td>LIA and modern</td>
<td>Push</td>
<td>Bulldozing and thrusting</td>
<td>1-10years</td>
<td>3-8m</td>
<td>Burki et al., 2009</td>
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<tr>
<td>Argentina</td>
<td>Modern</td>
<td>Push</td>
<td>Polygenetic push</td>
<td>Seasonal</td>
<td>.4-5.25m</td>
<td>Sharp, 1984</td>
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<tr>
<td>Iceland</td>
<td>LIA</td>
<td>Push/Thrust</td>
<td>1890 Surge</td>
<td>1 day</td>
<td>5m</td>
<td>Benediktsson et al., 2008</td>
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<td>Scotland</td>
<td>Younger Dryas</td>
<td>Ice Contact Fan and push</td>
<td>Debris flow and alluvium</td>
<td>3-9 or 7-19 years</td>
<td></td>
<td>Benn and Lukas, 2006</td>
</tr>
<tr>
<td>Rocky Mtns</td>
<td>LGM</td>
<td>Ice Contact Fan</td>
<td>Debris flow and alluvium</td>
<td>&lt;20 years</td>
<td></td>
<td>Johnson and Gillham, 1995</td>
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<tr>
<td>Iceland</td>
<td>Modern</td>
<td>Ice Contact Fan</td>
<td>Glacioluvial outwash fans/ sandur</td>
<td>&lt;10years</td>
<td>&lt;10m</td>
<td>Boulton, 1986</td>
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<td>Modern</td>
<td>Ice Contact Fan then push</td>
<td>Glacier is oscillating seasonally</td>
<td>&lt;21yrs</td>
<td></td>
<td>Boulton, 1986</td>
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<tr>
<td>Patagonia</td>
<td>Modern</td>
<td>Ice cored/ Push</td>
<td>Folding and Thrusting</td>
<td>&lt;13 years</td>
<td>15-20m</td>
<td>Glasser and Hambrey, 2002</td>
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<td>Alaska</td>
<td>Modern</td>
<td>Ice Cored/ Push</td>
<td>Readvance in a Surging glacier</td>
<td>1 year</td>
<td>3m</td>
<td>Johnson, 1972</td>
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<tr>
<td>Svalbard</td>
<td>LIA</td>
<td>Ice Cored</td>
<td>Retreating from LIA maximum</td>
<td>Formed upon retreat</td>
<td>25-30m</td>
<td>Lyså and Lønné, 2001</td>
</tr>
<tr>
<td>Svalbard</td>
<td>Neoglacial</td>
<td>Thrust</td>
<td>Melt out thrust</td>
<td>Formed upon retreat</td>
<td>&lt;20m</td>
<td>Bennett et al., 1996</td>
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<td>Modern</td>
<td>Dump</td>
<td>Minor push component</td>
<td>&lt;20 years</td>
<td>4-7m</td>
<td>Krüger et al., 2002</td>
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<tr>
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<td>Modern</td>
<td>Basal Freezing</td>
<td></td>
<td>1 year</td>
<td>1.5-2.5m</td>
<td>Krüger, 1993</td>
</tr>
</tbody>
</table>

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 C). Further sedimentological and stratigraphic 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, the Andes, and the New Zealand Alps. 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 unlikely to be found 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 conglomerate 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 A. 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.](image)
**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:

\[ R = \frac{A_{r>0} \mu \sigma_T}{A_{tot} \sigma_P}. \]

The \( R \) values for Front Range glaciers greater than 6 km² 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.

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