1 **Intermittent glacial sliding velocities explain variations in long-**
2 **timescale denudation**

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14 **Abstract:**

15 Quantifying controls on glacial erosion over geologic timescales is central to
16 understanding the role of Cenozoic climate change on the development of modern
17 mountain belts, yet the mechanisms that produce the distinct relief and topography
18 visible in glaciated regions remain poorly constrained. We test the hypothesis that
19 commonly assumed glacial sliding parameterizations control denudation rates over
20 geologic timescales. We do this by modeling glacier dynamics over a glacial-
21 interglacial cycle and compare with a dense dataset of (U-Th)/He
22 thermochronometer derived denudation rates from the southern Coast Mountains,
23 BC. Results indicate zones of rapid Quaternary erosion correspond to locations
24 where the model predicts the highest averaged sliding velocities. The results are
25 consistent with the hypothesis that sliding influences the rate of glacial erosion.
26 Regression between sliding predicted by the model and erosion rates show a

statistically significant correlation ($r^2=0.6$). The coefficient of the regression (10^{-5}) is smaller than previous estimates based on data from much shorter timescales. The model results also reveal that for a specific location, active subglacial sliding, and hence erosion, occurs for only ~ 10 - 20% of a glacial-interglacial cycle, suggesting high temporal variations in erosion rates. This intermittency of erosion requires instantaneous erosion rates to be greater than long term averages when active, explaining how timescale averaging can impact estimates of glacial erosion rates.

1.1 Introduction

In many mountainous landscapes Late Cenozoic climate change caused a transition from fluvial to glacial erosion processes (Brozović et al., 1997; Egholm et al., 2009; Herman et al., 2013). Understanding and quantifying this response is a focus of many glacial landscape evolution studies (Pedersen and Egholm, 2013; Tomkin, 2009; Yanites and Ehlers, 2012); however, quantifying the controls on glacial erosion remains elusive (Koppes et al., 2015). This results from a number of issues that obviate direct measurements of glacial erosion, including limited observations beneath active glaciers and a paucity of techniques for measuring rates of erosion in glacially modified terrain.

Empirical and theoretical evidence suggests that glacial sliding controls subglacial erosion, but testing this claim is limited to short timescales (Herman et al., 2015; Humphrey and Raymond, 1994), arguments based on simple topographic relationships such as the height of hanging valleys (Amundson and Iverson, 2006), or inferences from topographic reconstructions based on rock-cooling histories in

New Zealand (Shuster et al., 2011). Koppes et al. (2015), found that mean annual temperature, as opposed to glacial sliding, is the dominant control on century to decadal sediment flux out of temperate glaciers. Climate variability further confounds a quantitative analysis of sliding velocity and erosion rates as glaciers have consistently expanded and retreated in response to changing climate. The findings of previous studies combined with the variability of climate over the timescale of landscape evolution require an integrated modeling-data approach towards assessing the controls on glacial erosion over geologic timescales.

We utilize a natural experiment in the glaciated southern Coast Mountains, British Columbia, Canada (Figure 1) where spatial and temporal variations in Quaternary denudation have been documented over a 60 x 60 km region (Fig. 1b) (Densmore et al., 2007; Ehlers et al., 2006; Olen et al., 2012; Shuster et al., 2005) . Our approach involves quantifying and averaging glacial sliding through an entire glacial-interglacial cycle (both 40 ky and 100 ky) using a glacial model and comparing these results with long-term rates of Quaternary erosion. This comparison provides an evaluation of glacial erosion models used in modern landscape evolution models of mountain topography development.

2.1 Field area

The glacially sculpted landscape of the southern Coast Mountains of British Columbia provides a natural laboratory to explore the long-term (10^5 - 10^6 yrs) controls on glacial landscape evolution (Densmore et al., 2007; Ehlers et al., 2006; Shuster et al., 2005). During the Last Glacial Maximum in British Columbia, ice

expanded from confined alpine glaciers to form the Cordilleran Ice Sheet (Clague and James, 2002). The maximum ice sheet thickness occurred at ~14 ka, reaching thicknesses up to 2km. Although constraints on the western extent of the ice sheet are limited due to submergence of terminal moraines off the coast, there is enough evidence to reconstruct the general distribution of ice and flow directions during the LGM (Clague and James, 2002). Such an ice sheet likely existed in prior glacial periods, but constraints on the specific extent and thickness of these events are missing. Nonetheless, expanding and retreating alpine and ice sheet glaciers have carved deep glacial valleys into this region over the last 2 million years.

A dense dataset of 79 apatite (U-Th)/He thermochronology ages collected within a 60 x 60 km region provide constraints on long-term patterns of glacial erosion. Apatite (U-Th)/He data record rock cooling from ~60-90°C to the Earth's surface (Ehlers and Farley, 2003) and correspond to shallow crustal depths of ~2-4 km. These data record the latest phase of denudation in the Coast Mountains, which is largely glacially driven (Shuster et al., 2005). Thermochronometers provide a spatially averaged denudation rate over lateral distances akin to the closure temperature depth (e.g. ~2-4 km for apatite (U-Th)/He) meaning that individual samples record the denudation history of the area surrounding each sample. Topographic warping of isotherms can limit the utility of using a simple age as a quantitative marker for erosion (Olen et al., 2012). Using a 3-D thermokinematic model and linear inverse technique, Olen et al. (Olen et al., 2012) constrain the denudation rates across the Mt. Waddington region, accounting for the warping and advection of isotherms. The thermochronometer ages and denudation rates of this

region (Fig 1b) exhibit a large spatial variability in mineral cooling ages. This variability provides an opportunity to explore the controls on glacial landscape evolution and test the hypothesis that glacial sliding largely drives erosion. Because the erosion of the landscape occurs throughout a glacial-interglacial period, we use a glacial model to assess if sliding velocities, integrated over the entire cycle, can explain the spatial variability in denudation calculated from thermochronometer data.

3.1 Methods

3.1.2 Model

The distribution of glacial sliding velocity through a glacial-interglacial period is modeled using a shallow-ice approximation (SIA) for glacial mechanics with a constriction factor to account for topographic effects on ice flow (Braun et al., 1999). The ICE model (Herman and Braun, 2008) was used and adapted (Yanites and Ehlers, 2012) to include processes important to glacial mass balance and mechanics in the southern Coast Mountains including orographic precipitation (Roe et al., 2003), snow avalanching (Kessler et al., 2006), and iceberg calving (Van Der Veen, 1996). The model is run at time steps of 0.01 yr. Model inputs are topography, periodic climate variations, and glacial mechanical properties. For topography, we use a DEM in which estimates for the thickness of modern glacial ice in the region of Mt. Waddington has been removed (Clarke et al., 2009). Observations of exposed bedrock in many valley bottoms suggest that interglacial alluvial fill is minimal in the study area (Figure 1B). Furthermore, the exhumation indicated by the

thermochronology suggests that any accumulated sediment is evacuated by advancing ice to allow for bedrock erosion (Densmore et al., 2007; Ehlers et al., 2006; Shuster et al., 2005). The DEM used in this study was combined with bathymetry from the Canadian Hydrographic Service and resampled to ~700 m resolution (Fig. 1a) for the model simulations.

The effect of orographic precipitation on the glacial mass balance in southern British Columbia were accounted for with a moist-advection precipitation model (Roe et al., 2003). The model uses the integrated Clausius-Clapeyron relationship that incorporates a wind component that accounts for the horizontal transport of hydrometeors (Roe et al., 2003)

$$-\nabla \cdot F = (\alpha_0 + \alpha_1 \bar{v} \frac{dz}{dx}) e_{sat}(T_s) \quad (1)$$

where $-\nabla \cdot F$ is the vertical convergence of water moisture, α_0 is the convergence in the absence of topography, $\bar{v}$ is the wind velocity, dz/dx is the surface slope, and the coefficient α_1 partitions the moisture between the air column and intercepting topography (Roe et al., 2003). The term $e_{sat}(T)$ is the saturation vapor pressure which is a function of surface temperature. Climate parameters are chosen to match modern precipitation and temperature magnitudes (Figure 2a). Precipitation in the model falls as snow where the mean annual temperature (MAT) is below freezing. As can be seen in Figure 2, this approach captures the pattern and magnitude of high annual precipitation along the western coast as well as a dry interior plateau revealed by reanalysis data (New et al., 1999). The pattern and precipitation shift during the glacial maximum (Figure 2b), which results in an ice sheet with a similar

141 extent to the LGM (Fig. 3). To test for sensitivity to the pattern of precipitation, we
142 also ran a model with uniform precipitation equal to the mean rate in the area
143 around Mt. Waddington (1.5 m/yr) to make sure the same magnitude of
144 precipitation is falling in the different model runs.

145 Surface temperatures for the orographic and glacial models are chosen such
146 that they are consistent with reconstructions of the Last Glacial Maximum (LGM)
147 (Annan and Hargreaves, 2013) and produce glacial extents similar to the LGM
148 (Clague and James, 2002). Previous glacial periods likely had a different ice extent
149 than the LGM, but without constraints on these time periods, we cannot assess the
150 model ability to capture those events. Because of these uncertainties, we avoid
151 quantitative comparisons between the modeled and observed LGM ice extent. In the
152 results, we show that variations from LGM and modern ice extent do not
153 significantly alter the quantitative comparison between sliding and erosion, and we
154 are therefore confident that our results provide robust insight into the sliding
155 patterns averaged over the multiple glacial-interglacial periods. Finally, we note that
156 this region of the Coast Mountains served as an ice divide through most of the last
157 glacial (Clague and James, 2002), and this is reproduced by our models.

158 Glacial- interglacial conditions are driven by varying temperature along a
159 sine wave with minimum and maximum temperatures of -2 and 8°C. The chosen
160 interglacial temperature is consistent with modern temperatures at sea-level in this
161 region of Canada and produces a glacier extent consistent with the modern (Figure
162 3). The value of glacial temperature (-2°C at sea-level) is a conservative estimate of
163 surface temperature changes at LGM for this region (Annan and Hargreaves, 2013).

Sea-level varies linearly with temperature with a minimum of -125 m and a maximum of 0 m, and thus, for these model runs, there is no lag between temperature change and sea-level. For computational efficiency, the period of the sine function is set at 40 ky but we ran one model at 100 ky and found consistency among the patterns of ice extent and sliding velocities. We note that local sea-level regressions/transgressions were decidedly more complicated due to isostatic effects of the ice sheet (Clague and James, 2002); however, based on model sensitivity tests, the pattern of sliding in the study region is sufficiently isolated from the effects of the ice sheet terminus, though the absolute magnitude of modeled velocities might vary with different sea-level histories.

Snow and ice are melted at a rate (M) dictated by an empirical positive-degree day relationship (Braithwaite, 1995). We model the daily average annual temperature variation with a sine wave (period=365.25 days). We sum the temperatures (T) for days when the mean temperature is above freezing. The product of this total and the empirical ablation coefficient (k_{melt}) estimated by Braithwaite (1995) provides an estimate of the annual mass loss to ablation,

$$M = k_{melt} \sum T_{(T>0)} \quad (2)$$

where a typical value of k_{melt} is ~ 0.008 (Braithwaite, 1995).

Snow avalanching is simulated by assuming a critical slope stability angle. Snow from surfaces steeper than the threshold for slope stability (40 degrees) are avalanched (mass transported downslope) until either the slope is below this threshold or all the snow has been removed (Kessler et al., 2006). This works by successively moving snow downslope for any surface that is $>40^\circ$. The magnitude of

transported snow is such that the new slope will equal 40°. If this value exceeds the amount of snow present at that node, then only the available snow is moved.

Ice that terminates below sea-level is calved at a rate linearly dependent on the depth of the ice below the water surface (H_w) (Brown et al., 1982; Van Der Veen, 1996)

$$M_c = k_c H_w \quad (3)$$

where M_c is the mass of ice loss and k_c is an empirically derived constant (Brown et al., 1982). Sliding velocities are set to zero when the ice is no longer in contact with the bed (i.e. where modeled ice is sufficiently below modeled sea-level such that flotation occurs). Sliding velocities up-glacier of the grounding line are still driven by the driving stress and not directly influenced by this condition.

Next, ice is transported by both internal glacial deformation and basal sliding processes. Both of these transport mechanisms are assumed to be functions of driving shear stress, τ , using a shallow-ice approximation approach (Braun et al., 1999)

$$\tau = \rho g h S_i \quad (4)$$

where ρ is the density of ice, g is the gravitational acceleration, h is ice thickness, and S_i is the ice surface slope (not the bed slope). To account for the effects of valley walls on glacial movement, the model calculates a constriction factor, β to reduce the deformation and sliding velocities.

$$\beta = \frac{1}{1 - \gamma \frac{d^2 z}{dx^2}} \quad (5)$$

where λ is a characteristic length scale for which lateral stresses influence ice deformation and sliding. A λ of $\sim 1\text{km}$ produces flow patterns consistent with higher order models (Egholm et al., 2012).

The integrated deformation velocity, u_d , is calculated as a function of this stress and constriction factor

$$u_d = \beta k_d (\tau)^n \quad (6)$$

where k_d is a coefficient dependent on the ice properties and n is equal to 3, as in the commonly used Glen's flow law. We assume that glacial sliding is also a function of the driving stresses (Fowler, 1981), c , and has a general form

$$c = \beta k_g (\tau)^a \quad (7)$$

where c is the glacial sliding velocity, k_g is a constant (related to water pressure and bed friction), and a is the glacial sliding exponent. The values of k_g and a are poorly understood, so to test for sensitivity, we explore a range of values (Table 2) that still give reasonable instantaneous sliding velocities (~ 100 's of m/yr). Other work exploring this relationship at finer scales than our grid resolution have suggested physical limitations on such a formulation (Schoof, 2005); however, the exploration of a finer scale model incorporating more complex processes is beyond the scope of this work. Instead, our focus is to test the model described by equation 7 that is commonly used in landscape evolution models (Herman and Braun, 2008; MacGregor et al., 2000; Tomkin and Braun, 2002). Sliding velocities are tracked at each time step and averaged over a full glacial-interglacial period. The model also tracks the percent time during which active sliding occurs.

A model domain of $\sim 420 \times 320$ km (Figure 1a) with Dirichlet boundary conditions (e.g. fixed boundary of 0 m of ice) was found to be sufficiently large to remove any boundary effects on the modeled ice thickness and surface slopes in the immediate Mt. Waddington region (Figure 3). Modeled ice extents are of similar extent as both modern and last glacial extents (Figure 3). Comparison to thermochronometer ages was conducted by averaging values of glacial sliding over a 2.5×2.5 km window (the minimum spatial scale cooling ages are sensitive to). We ran two glacial-interglacial cycles and only analyzed the second cycle to limit effects of the model ramp-up.

Model predicted sliding velocities are compared to denudation rates determined from inversion of cooling ages using an inverse transient 3D thermal model (Olen et al., 2012). Olen et al. (2012) inverted thermochronometer ages (Fig. 1B) for the range of denudation rates and paleotopography required to reproduce the ages. In addition to comparing the entire dataset with local valley sliding, we also limit our regression analysis to the interior region (Fig. 1b) of the inversion (Olen et al., 2012) where observations are available. We further constrain the dataset to valley bottoms with relatively wide valleys compared to ice depth (width $>5 \times$ depth). Valley bottoms were identified such that the local elevation was within 200m of the minimum value in a 2km radius on the ice free DEM. This elevation threshold was chosen to best represent locations where it is most likely that *only* glacial erosion has driven mass removal. Above 200 m, interglacial processes such as rock fall and other mass movements occur frequently, and therefore, glacial sliding might not be the dominant driver of local erosion processes even though it is

active when ice is present. Therefore, some samples that would appear to be valley bottom based on the elevation threshold of a modern (i.e. not ice-free) DEM were discarded because the local ice is greater than 200m thick. Two samples taken along the Franklin Glacier were discarded due to this effect. This limits the analysis to wide valleys, ensures that valley widths are >1 pixel for the analysis, and limits this refined analysis to samples directly eroded by subglacial processes.

4.1 Results

Results from the model-data comparison indicate that sliding velocities averaged over a full glacial-interglacial cycle are in agreement with the distribution of denudation rates (Figure 4). For example, zones of high average sliding velocity in the Mt. Waddington area are found in the central Klinaklini Valley and the upper Franklin valley for all glacial sliding models (red areas, Fig. 4a). These valleys coincide with a region of enhanced denudation compared to the area to the northeast and southwest (Fig. 4a,) (Densmore et al., 2007; Ehlers et al., 2006; Olen et al., 2012).

For the majority of a glacial-interglacial cycle no glacial sliding occurs (i.e. all values in Fig. 4b are less than 50%). The extensive periods of no sliding leads to the low average sliding velocities observed in Fig. 4a. The pattern of sliding frequency does not match the distribution of denudation as well as average sliding velocities. For a given point in the landscape, the majority of time (80-90%) is spent either ice-free (~5-10%, Figure 4d) or under cold-based, non-sliding ice (75-80%). As is evidenced by the persistence of glaciers through the interglacial period (Fig. 4C),

much of this upland landscape remains covered by ice through the model run and only becomes wet-based during interglacial periods (Fig. 4C). The intermittent sliding affects the long-term average sliding rate such the values are about an order of magnitude lower than the maximum instantaneous (one model time-step) velocities.

The general pattern of higher erosion rates in regions of faster sliding velocities is consistent when analyzing the model sensitivity to parameter choices (Figure 5). The model sensitivity analyses all reproduce reasonable glacial-interglacial histories (i.e. maximum glacial extent is similar to observations and interglacial extent is similar to modern distribution of glaciers). In general the patterns of sliding are consistent from model to model (Figure 5) although the magnitudes of sliding are different. A simulation with uniform precipitation, equal to the modern mean precipitation, also produces sliding patterns consistent with the pattern of erosion (Figure 5F). This suggests that topography is more important than the precipitation pattern in dictating the distribution of sliding, though precipitation plays a role as well (Herman and Brandon, 2015).

4.2 Calibrated model

The consistency between sliding velocities representative of a glacial-interglacial cycle and denudation rates over million-year time scales can be used to evaluate a glacial erosion model through a simple linear regression analysis (Figure 6). We compare the denudation rates with the glacial sliding velocities in the local valley center, defined as the deepest location in valley cross section for any sample

location (Figure 6A.). Most samples are from high on hillslopes where glacial processes are nonexistent; however, thermochronologic data are sensitive to mass removal over horizontal length scales of kilometers (Ehlers and Farley, 2003). Additionally, erosion in the local valleys will steepen the local hillslope gradients, driving the rate of geomorphic processes on the hillslopes. Therefore, the denudation rates provide a reasonable proxy for the spatially averaged rate of the local, valley glacial erosion. The analysis limited to only the samples that are dominantly eroded by glacial processes (e.g. valley bottom samples) results in an improved correlation ($r^2=0.60$) though the regression coefficients are similar.

Assuming that the average modeled sliding velocity serves as a proxy for the long-term average, the coefficient calculated from the regression can be used to calibrate an erosion model for this lithology and location (Table 2). The slope of the trend is statistically significant and ranges from 2×10^{-5} to 8×10^{-5} for the sensitivity analyses (Figure 7 and Table 2). These values are slightly lower than the 1×10^{-4} values found over annual timescales of a surging glacier (Humphrey and Raymond, 1994). A lower value (i.e. less efficient glacial erosion) in this study would be expected as the approach of Humphrey and Raymond (1994) used sediment flux inventories, some of which were likely sourced from subglacial storage, to constrain instantaneous erosion rates.

Regressions between sliding velocity and erosion rate also provide consistent results, though some model results provide a stronger correlation than others (Figure 7). Model mEXP1 provides the least robust fit. The glacio-mechanic properties of this model (Table 2) lead to sliding velocities on the high-end of

observations (with instantaneous sliding rates on the order of km's per year). The best correlation is found in model mEXP3 and the subset of runs mEXP3_100ky and mEXP3_cold. These models all have a higher exponent (and lower coefficient) than mEXP2. Model mEXP3_100ky provides a test of the cyclicity assumption and shows very little difference between a 40ky and 100ky simulation (Figure 7 A and B). Model mEXP3_cold was conducted with a colder glacial temperature and, again, shows little difference with the regression results in mEXP3 (Figure 7 A and C). Warming during the interglacial generated a climate that completely melted glaciers during this period, which is not consistent with the modern ice extent.

Recent work suggests a non-linear relationship between sliding velocity and erosion (Herman et al., 2015; Koppes et al., 2015). We find that a power-law, rather than linear, relationship between glacial erosion and sliding velocity would also fit the data (Figure 6 and 7). However, the calculated denudation rates span less than one order of magnitude and such an analysis would be unwarranted. A power-law model would circumvent the peculiar result of a non-zero y-intercept in our regression analysis (Figure 6 and 7). We interpret the non-zero intercept to suggest that as the sliding reduces below some threshold value, other erosion processes are responsible for the bulk of erosion (e.g. perhaps interglacial fluvial processes). Low rates of glacial sliding on valley bottoms do occur north of our sample locations (Figure 4a), and estimates of denudation using higher temperature thermochronometers suggest lower rates of erosion to the north (Parrish, 1983), which could help constrain the shape of the lower sliding velocity zone, but the rate over the Quaternary is unconstrained. This analysis provides further support that

the pattern of erosion in the Coast Mountains is driven by spatial variations in glacial sliding velocities; however, the data do not distinguish whether a linear or power-law model is more appropriate.

5.1 Discussion

Our results document that sliding velocity influences subglacial erosion. Our results demonstrate that a linear glacial erosion model based on sliding velocity can explain ~60% of variance ($r^2=0.57$, Figure 6) in the valley bottom denudation rates derived over million year time scales from (U-Th)/He cooling ages. We note the results are also consistent with a power-law model of sliding-erosion (e.g. Herman et al., 2015) though our data are inconclusive on if one provides a better fit (Figure 6 and 7). Sensitivity analyses demonstrate that this relationship holds for a range of glacial-interglacial histories (Figure 7).

We also find that averaging sliding during a full interglacial-glacial period is necessary to capture the pattern in glacial erosion efficiency across the landscape. We suggest that this integrated average sliding rate is similar to the concept of an ‘effective discharge’ in fluvial geomorphology (Wolman and Miller, 1960), whereas a mean rate is one that captures the cumulative impacts of an entire distribution of events. In such cases, a short-timescale measurement of glacial erosion is insufficient for explaining the long-term rates (e.g. Koppes and Montgomery, 2009).

Several other studies support using a simple sliding model in landscape evolution studies to drive glacial erosion (Harbor, 1992; Herman et al., 2015; Herman and Braun, 2008; MacGregor et al., 2000; Shuster et al., 2011; Tomkin and

Braun, 2002). The actual physical processes of glacial mechanics and erosion are much more complicated than encapsulated in the SIA approach (Egholm et al., 2012; Hallet, 1996, 1979; Headley and Ehlers, 2015; Herman et al., 2011). However, based on the results presented here, we suggest that sliding velocity derived erosion rates can capture the erosive potential of glaciers on geologic timescales. Complexities associated with factors such as joint-spacing (Dühnforth et al., 2010), ice dynamics, and hydrology (Egholm et al., 2012) will influence the ability for a simple model to predict glacial erosion over short timescales; however, our results suggest that in systems like the southern Coast Mountains of British Columbia, such factors average out over long timescales and simple sliding models provide appropriate proxies for the geomorphic effectiveness of the system.

Topography has evolved over the timescale of the thermochronology data potentially changing the distribution of sliding velocities (Tomkin and Braun, 2002). However, our correlation between sliding velocity and erosion rate suggests a feedback where locations of erosion drive future ice flow into the lowered regions, maintaining faster rates of erosion (Kessler et al., 2008). We suggest that the original shape of topography dictates a locus of rapid erosion that continues to erode more rapidly, albeit at a potentially lower rate. This assertion is supported by the paleotopographic reconstruction of Olen et al. (2012) which show that valley locations have remain relatively stable over 1.8 Ma with the potential removal of a relatively low relief drainage divide as the only significant drainage rearrangement..

We note that previous work constrained very short, rapid pulses of incision in this area (Shuster et al., 2005), but this pulse occurred over a very short period

and the information from a single locality does not provide the spatial information necessary to quantify erosion differences along a valley. This pulse of erosion is likely the result of the flow of glaciers over a landscape that was previously dominantly sculpted by fluvial processes (e.g. Pedersen et al., 2014). The removal of a drainage divide has also been suggested for the area between the Klinaklini and Franklin Valleys (Ehlers et al., 2006). Reinterpretation of this data with a 3D inversion for transient paleotopography suggests a change in the location of the highest topography (Olen et al, 2012). However, neither study requires a divide to extend across the Klinaklini Valley. In fact, the study of Olen et al. (2012) suggests either a low-relief trough or a low saddle along the Klinaklini Valley at 2 Ma. This zone of rapid of sliding could also be related the migration of a knickpoint-like wave of erosion through the landscape similar to what has recently been suggested for New Zealand (Shuster et al., 2011); however, in such a case, we would expect to find higher denudation rates at the south end of the study area than what is currently observed.

Considering this previous work with the results of this study, we suggest that the area outlined in Figure 4a has been, and continues to be, a locus of rapid erosion set up by early topographic steering of ice. This early steering of ice caused a short-lived, rapid pulse of incision, followed by continued erosion at increased rates due to the feedbacks of ice flow, topography, and climate (Kessler et al., 2008; Shuster et al., 2005). Moreover, while denudation rates vary by ~ 0.6 mm/yr across the study area, localized differences in denudation (e.g. between two nearby points) are on the order of 0.3 mm/yr. Therefore, the maximum relief change between any sample

locations over 2 My is ~600 m (Figure 4). We note that these interpretations rely on the assumption that the pattern recorded by the integrated signal of erosion in the thermochronology data is representative of continued spatial differences in subglacial erosion potential across the landscape.

The prediction that glacial erosion processes only operate on a local scale over relatively short timescales compared to the evolution of the landscape (Figure 4b) has important implications for previous interpretations of erosion rates and glacial processes. Intermittency of erosional processes has long been considered a principal reason for observed variability in landscapes and sedimentation histories (Finnegan et al., 2014; Jerolmack and Sadler, 2007; Sadler, 1981). Compilations of data from glaciated regions show that the estimated rate of erosion depends on the timescale of measurement, with longer timescales resulting in lower rates (Koppes and Montgomery, 2009). If the frozen and un-erosive conditions expand during glacial periods, as predicted by the low percent of time of active glacial sliding (Figure 4b), then modern denudation rates under actively sliding glaciers must be about an order of magnitude greater than the long-term rates. Thus, although glacial erosion can be significantly enhanced for short time periods over localized areas (Koppes and Hallet, 2006), the magnitude of Quaternary climate oscillations, and limited climate window in which glaciers are erosive (Koppes et al., 2015; Yanites and Ehlers, 2012), likely reduces the ability for glaciers to maintain extraordinarily high regional erosion rates.

6.1 Conclusions

The general agreement between the spatial distribution of glacial-interglacial average sliding velocity and Quaternary denudation rates from thermochronometer data suggests that glacial sliding controls subglacial erosion processes. Along the Klinaklini Valley, denudations rates vary from a maximum of ~ 0.2 to ~ 1 mm/yr. The pattern of sliding velocities closely mimics the denudation rate, with absolute magnitudes depending on model parameters. In any case, the relationship between sliding and erosion is on the order of $2\text{--}8 \times 10^{-5}$. These results are consistent with previous studies in the Coast Mountains that suggest glacial erosion, and particularly glacial sliding, is the primary control on denudation process in valley bottoms over the last few million years (Olen et al., 2012; Shuster et al., 2005). Furthermore, around Mt. Waddington, active, subglacial erosion only operates for $\sim 10\text{--}20\%$ of the landscape's history, leading to strong temporal variability in erosion rates at timescales less than a glacial-interglacial cycle. Such intermittency can explain the impacts of timescale averaging on glacial erosion rates.

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FIGURES:

Figure 1: Map of the southern Coast Mountain region in British Columbia, Canada showing combined digital elevation and bathymetric models (A). Black boxes indicate model domain (see Figure 3) as well as study area. (B) Map of the distribution of apatite (U-TH)/He ages in the study area. White outline indicates the locations for the regression analysis in Figure 6. The extent of Fig. 1B is the same for Figs. 4 and 5.

Figure 2: (A) Modeled interglacial precipitation pattern using the model described in the supplementary text for the model run mEXP2. Contours are modern reanalysis data of precipitation from the 30 yr (1961-1990) CRU 0.5° mean monthly climatology (New et al., 1999). Note the scale mismatch between the modeled (700 m) and modern precipitation (0.5°). (B) Modeled precipitation pattern during the glacial maximum. Note that precipitation model runs over the ice surface, such that the shaded relief shown is not represented of the surface the precipitation pattern was modeled on. The pattern essentially mimics the shape of the ice sheet

Figure 3: (A) Modeled ice thickness at glacial maximum for model mEXP2. The lack of ice in the northern part of Vancouver Island and the northwest part of the model

domain is a result of the limited northern extent of the modeled region. Glacier mechanics are unaffected by this in the study area. Red line is the modeled ELA during the glacial maximum. The green line is the estimated LGM ice extent from Ehlers and Gibbard (2004). (B) Modeled interglacial ice cover and thickness (color bar). Green line represents modern ice extent.

Figure 4: Model results in the study area around Mt. Waddington for model run mEXP2. Colored dots represent denudation rates from apatite (U-Th)/He ages. (A) Map of sliding velocity averaged over a glacial-interglacial cycle. Contours are at 2 m/yr intervals. (B) Map of the percent of time through a full glacial cycle that the grid point had sliding ice present. Contours are at 5% intervals. (C) Map of the percent of time through a full glacial cycle that a grid point has ice below the pressure melting point at the base. Contours are 15% intervals. (D) Map of percent time the landscape is ice-free. Contours are 2% intervals.

Figure 5: Sensitivity of the pattern of sliding to model parameters. Although absolute magnitudes vary, the patterns, and therefore the relationship between sliding and erosion (Figure 7), are similar. See Table 1 for differences in model parameters. (A-C) are a higher sliding-shear stress exponent. (B) is the same as A except for a 100ky cycle (versus 40 ky for A). (C) has a colder glacial temperature but is otherwise the same as (A). (D) has a lower sliding-shear stress exponent and (E) is the same as (D) but with a lower sliding-shear stress coefficient. (F) is the run with uniform (1.5 m) annual precipitation.

643

644 Figure 6: Thermochronology data versus valley center and average sliding velocity.

645 Estimates of erosion rate are from thermochronometric and topographic inversion

646 data (Olen et al., 2012) (A) Thermochronology estimated denudation rate vs. local

647 valley center sliding velocity. Symbols are colored by the height of the sample above

648 the local valley center elevation. Uncertainties in ages are propagated to the erosion

649 rate estimates (Olen et al., 2012). (B) Erosion rates from samples that are

650 dominantly eroded by glacial processes based on field observations and height

651 above valley floor (e.g. no ridge top samples) versus the local sliding velocity

652 averaged over a glacial-interglacial period. The regression between erosion and

653 sliding velocity amounts to a calibrated glacial erosion rule over the Quaternary.

654 Note that the sliding velocities used in A and B are different (see text).

655

656 Figure 7: Sensitivity of the regression analysis to model parameters. See Table 2 for

657 regression comparison. (A-C) are a higher sliding-shear stress exponent (value =3).

658 (B) is the same as A except for a 100ky cycle (versus 40 ky for A). (C) has a colder

659 glacial temperature. (D) has a lower sliding-shear stress exponent (value =1) and

660 (E) is the same as (D) but with an order of magnitude lower sliding-shear stress

661 coefficient. (F) is the uniform precipitation run.

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Table 1: Glacial Model Parameters for the southern Coast Mountains, BC

Variable name/description	Value (range)	Units
<i>Model set-up</i>		
Dimensions	420x320	km
Grid spacing	700 m	km
<i>Climate</i>		
* Base-level temperature (glacial)	[-2 8]	°C
* Periodicity of temperature variation	[40][100]	kyrs
a0	2.3	m/yr
a1	110	m/yr per (m/s)
alf	100	1/(m/s)
average wind speed	1	m/s
wind direction angle from x-axis	90	degrees
Atmospheric lapse rate	6.5	°C/km
<i>Glacial Parameters</i>		
Ice-flow constant	6.80E-24	1/(Pa ³ s)
	<i>See regression table</i>	
* Ice-sliding constant		(1/(Pa ^a s))
* Ice-flow exponent	3	non-dimensional
	<i>See regression table</i>	
Ice-sliding exponent		non-dimensional
Ice density ⁷	910	kg/m ³
Constriction constant	1.00E+03	m
Calving coefficient	2	1/yr
Basal heat flux	5.00E-02	W/m ²
Ice thermal conductivity	2.4	W/mK
Positive degree day melting coefficient ¹	8.00E-03	K m/yr
Annual temperature variations	15	°C
Now stability angle for avalanching	40	degrees

* Denotes free parameter explored in model presented in this manuscript.

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Table 2. Regression results for sliding velocity and erosion rate

Model	Coeffic- ient (k _g)	Expon- ent (a)	Period (kyrs)	Glacial T(°C)	Regression results			
					Slope (x10 ⁻⁵)	Intercept (x10 ⁻⁵)	R ²	p-value
mEXP3	1.0E-14	3	40	-2	7.2	5.3	0.61	0.003

mEXP3_100ky	1.0E-14	3	100	-2	5.7	5.4	0.63	0.002
mEXP3_cold	1.0E-14	3	100	-3	5.9	5.3	0.63	0.002
mUNIFORM_PPT	1.0E-14	3	40	-2	5.2	5.1	0.6	0.002
mEXP2	1.0E-09	2	40	-2	4.6	5.2	0.55	0.006
mEXP1	1.0E-03	1	40	-2	1.9	4.7	0.38	0.033
mEXP1_LOWSLID	1.0E-04	1	40	-2	7	4.9	0.49	0.012
* Uniform Precipitation								

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