Utilizing Remote and Numerical Methods to Provide Constraints for the Seasonal Development and Topographic Profiles of Rock Glaciers

The glacial valleys of Mount Sopris viewed from the north (October 2016)
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Time lapse camera overlooking the central valley of Mount Sopris (October 2016)
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Abstract

Rock glaciers represent the dynamic interaction between rock and ice in many alpine settings that lie below the Equilibrium Line Altitude (ELA). These periglacial systems are formed by avalanched snow and debris from an overlying headwall, and are adorned with distinct topographic lobes collectively known as rumples. The central rock glacier of Mount Sopris presents a clear expression of rumples, where the structures are well-defined throughout the 1.8 km long glacier. In addition to clearly-expressed rumples, the accumulation area is constrained to a narrow bowl at the base of the headwall that is easy to identify. To inform our understanding of rock glaciers, we surround existing remote sensing data with an array of techniques to quantify the spatial distribution of rumples, and qualitatively analyze the development of the avalanche cone with time-lapse photography.
Chapter 1: Introduction

1.1. Background

In many alpine settings that lie below the equilibrium line altitude (ELA), rock glaciers are the most prominent expression of the cryosphere. These rock-covered glaciers occupy alpine valleys with over steepened headwalls that were carved out of the landscape by glaciers during previous glacial maxima (Humlum, 2000). The ideal conditions for rock glaciers were created as Earth transitioned into the current interglacial period, and widespread annual ice cover disappeared from lower latitude alpine zones.

Rock glaciers develop when avalanched snow from the overlying headwall of an alpine valley augments local snowfall. A positive mass balance at the base of the headwall ensues, which creates year-round snow cover. Debris is also avalanched from the headwall into the accumulation area, scattering rock across the surface of the avalanche zone.
The rocky surface acts as a solar insulator, allowing snow to persist even further below the ELA (Shroder et al., 2000). As ice moves down valley from the avalanche cone, the headwall-sourced rocks are transported down the valley. This removal of rocky debris from the headwall is crucial to the morphology of alpine landscapes. In the absence of the removal of this material, the headwall would simply develop a gradually sloping talus field, rather than the dramatic walls of high relief that we see today, which in turn could shut down its backwearing. In this way, rock glaciers act as natural conveyer belts that carry rocks hundreds of meters from their source (Millar & Westfall, 2007).
In addition to their importance in shaping alpine landscapes, rock glaciers supply critical late summer water to alpine ecosystems. With global temperatures on the rise, pure ice glaciers are shrinking in size and disappearing altogether. This creates a number of problems, including loss of year-round water supply in certain regions. However, debris cover dampens the effect of climate change on rock glaciers, and will help to preserve their water storage potentials as water supply from pure glaciers dwindles (Scherler et al., 2011). Despite their growing importance, rock glaciers remain fairly enigmatic.

Numerical models for the behavior of rock glaciers have been developed by adapting existing models for pure glaciers (Anderson et al. 2016). These models provide insight into how rock glaciers respond to variable climate through time. However, many unknowns pose serious challenges to the existing models. One challenge is the lack of understanding of how the accumulation area evolves through winter. Constraining the
timing of snow and debris delivery to the avalanche cone will better inform future numerical models.

Rock glaciers also exhibit a unique appearance, characterized by rumbles that create crests and troughs across the glacier. The rumbles on active rock glaciers appear as waves that span the surface of the rock glacier from the accumulation zone to the terminus. The process that forms these unique surface waves is unknown.

One hypothesis for their formation is that the amount of material stacked in the avalanche cone depends on short-term climate. This would result in the formation of large lobes of material during high accumulation events. The subsequent lobe would then be transported away from the headwall by the rock glacier, and an ensuing low accumulation event would create a disparity in elevation between the lobe and the...
resulting trough. The spacing between rumples would then be dependent on velocity of the rock glacier and frequency of high accumulation events.

Thrust faulting from compressional stresses within the rock glacier could also create the rumples. Moore et al. (2010) attribute the formation of similar structures, known as arcuate bands of debris-rich ice, in pure glaciers to thrust faults generated by longitudinal compression. This too must be considered for rock glaciers, and will depend sensitively upon the velocity structure of the glacier.

![Figure 4: Schematic of thrust faults in a pure glacier creating arcuate bands of debris-rich ice. Arrow indicates flow direction (Moore et al., 2010).](image)

A valid hypothesis for rumple origin must honor empirical observations of the spatial distribution of rumples. In this study, we use high resolution Digital Elevation Models (DEMs) to constrain rumple wavelengths along several rock glaciers. We also attempt to develop a proxy that can be employed when high resolution elevation data is not available for rock glaciers, which is often the case. To this date, these interesting structural features are little-studied, and they present a great mystery.

The central rock glacier of Mount Sopris presents one of the best expressions of rumples, where the structures are well defined throughout the 1.8 km glacier. On a
cursory overview, order appears in this system. The mind is able to discern patterns and find repetition in the structures. This dynamic system is contained within an alpine valley at the base of a steep headwall that supplies snow and debris to form the glacier. In addition to clearly-expressed rumples, the accumulation area is also well defined as a narrow bowl at the base of the headwall. These rumples and the distinct accumulation area on the central Mt Sopris rock glacier present an opportunity to find order arising from dynamic surface processes, and are the motivation for this study.

Figure 5: Prominent rock glaciers of the Sopris Massif. The simplicity of the central rock glacier makes it an ideal target for this study. September 2011 Imagery obtained through Google Earth Pro.
1.2. Geologic Setting

The twin summits of Mount Sopris rise to an elevation of 3952 m in the West Elk Mountains of Colorado. The prominence of Mount Sopris is easily seen southwest of highway 82 between Glenwood Springs and Aspen Colorado. Along this stretch of highway, in the Roaring Fork Valley, two distinct lithologies appear to dominate the landscape. Lithologies that are defined by their respective colors: soft drapes of deep red and chalky grey that have been pierced by towering light-grey plutons.

The soft units are sedimentary, and are composed primarily of the Minturn, Maroon, and Entrada formations. The Pennsylvanian Minturn Formation contains evaporates and carbonates, which reflect a marine depositional environment. A shift in depositional environments to a low-lying fluvial system created the Maroon Formation, as the deep red pebbles and cobbles of an ancestral mountain range were deposited during the
Pennsylvanian/Permian. During the Upper Jurassic, the fluvial environment then transitioned to one dominated by wind, which covered the Maroon formation in the dune-scale aeolian deposits of the Entrada Formation (Welder, 1954). The Laramide Orogeny dramatically changed this landscape late in the Cretaceous, as flat slab subduction of the Farallon Plate beneath the North American Plate generated regional compression that brought these sediments to an elevation comparable to what we see today (Heller & Liu, 2016).

Following subduction, the Farallon Plate pulled away from the North American Plate at about 40 Ma. Hot asthenosphere was brought into contact with cold lithosphere beneath the West Elks, which generated large volumes of magma that lowered mantle density (Farmer et al., 2008). Buoyant forces caused the less dense mantle to rise, which promoted incision of rivers that led to the exhumation of the underlying plutonic rocks. These light-grey plutons occur in seemingly random blobs that comprise the other dominant lithology of the Roaring Fork Valley. The Mount Sopris Massif is graced by vast swathes of this broken Tertiary granite derived from this event, known as the Ignimbrite Flare up (36-25 Ma) (Donahue et al., 2014).
Once exposed to the surface, the prominent granitic blobs were subjected to weathering and erosion, and were eventually carved by glaciers of pure ice during glacial maxima in the Quaternary. The glacial valleys on the north and north-western flanks of Mount Sopris give the mountain an asymmetric appearance. A dominant wind direction maintains this asymmetry by removing rock from the steep downwind side of the asymmetric crest (the headwall). This causes increased avalanching and develops a pocket of high net ice accumulation, which bites at the base of the headwall to create an over-steepened slope that is more susceptible to erosion (Sanders et al., 2012; MacGregor et al., 2009). With greater rates of erosion, rock glaciers form as snow in the avalanche cone becomes buried in debris, and begins to move toward a lower elevation. As it moves, the debris is transported and is therefore prevented from burying the headwall in sediment.
1.3. Thesis Overview

In this study, we gather data to enhance our understanding of two distinct components of rock glaciers. The first approach draws upon quantitative methods to identify the spatial distribution of rumpled on the central rock glacier of Mount Sopris. We attempt to develop a viable proxy to detect each rumple in lateral space, but ultimately opt to generate high resolution elevation data to attack this problem directly. We then contextualize our findings by implementing the same numerical analysis on other rock glaciers for which high resolution elevation data has been developed. The second element of this study utilizes time-lapse photography to monitor the avalanche cone that serves as the source area for the central rock glacier of Mount Sopris. With this qualitative analysis, we better constrain the type of material that is delivered to the avalanche cone at different points in the year.
Chapter 2: Defining the Spatial Distribution of Rock Glacier Rumphles

2.1. Remote Methods

Before we can determine the spatial distribution of rumples across the central rock glacier on Mount Sopris, we must gather data. These rumples are topographic artifacts of an unknown processes. With elevation data, the structural prominence of rock glaciers can be quantified, and we can apply a variety of numerical methods to understand their spatial distribution. However, high resolution elevation data is not readily available for Mount Sopris, requiring an indirect first approach to this problem.

2.1.1. The Proxy

Low resolution digital elevation models (DEMs) for large swaths of land started to become readily available with the launch of Satellite Pour l’Observation de la Terre (SPOT) in 1986 by the French Space Agency (Chevrel et al., 1981). After 1986, the resolution and extent covered by DEMs increased with the launch of each new satellite. At present, a 10-m resolution DEM exists for the entire contiguous United States, including Mount Sopris (“The National Map: Elevation,” 2017).

This DEM works well for studying large features, but presents problems when trying to gather data for smaller features like rumples, which generally have wavelengths smaller than 10 m. Rumples are mostly absent from topographic profiles generated with the 10
m DEM, meaning this data is inadequate for our study. To quantify their spatial distribution, we must develop data that can resolve the rumbles.

![Image](image.png)

**Figure 8:** 10-m DEM used to generate Hillshade (left) and elevation profile (right) for the given transect across the central rock glacier of Mount Sopris. At this resolution, most rumbles are not visible.

Variation in brightness across the rock glacier reflects the presence of rumbles. As the sun illuminates each rumple, the high crests are illuminated, while the low troughs remain in the shadows. The crests are always brighter than their adjacent troughs, regardless of the position of the sun. When the values for the brightness of each pixel is paired with the distance of each pixel from the headwall, a profile that oscillates between high and low brightness values is generated. As expected, the local minima and maxima correspond to the troughs and crests of the rumbles, respectively.

Using an ArcGIS 3D analyst tool, I created a brightness profile along the rock glacier for images obtained at three times: September 1993, June 2006, and September 2011. The 3D analyst tool allows users to create a line across any raster dataset, and generates a two-column matrix of each raster data value paired with the location of that
raster cell along the length of the drawn line. Armed with matrices for all three years, I constructed profiles for brightness as a function of distance from headwall. This tool is also used to construct the elevation profiles throughout this study.

I obtained imagery for the brightness analysis through Google Earth, imported the files into ArcGIS, and used a third order polynomial to georeference the images to a basemap. For consistency, I exported each image from Google Earth with the same dimensions, meaning that I only needed to georeference one image, then apply that reference to the others. ArcGIS displays images as datasets composed of three rasters, corresponding to red, green, and blue wavelengths in the visible spectrum (bands 1, 2, and 3, respectively). The 3D analyst tool can only generate matrices with two parameters at a time: a single raster and a length scale. To account for this, I excluded the second and third visible spectrum bands from the raster, which allowed Arc to read only the first band of each pixel that the transect crossed.

To compare brightness between different images, the same transect should be used every time. However, Arc is unable to read different raster files with the same transect, and a new transect must be created for each raster (each satellite image). To limit variability between transects, I created a shapefile to use as a stencil when drawing the individual transects. The stencil does not create identical transects between images and only provides loose reproducibility. With these methods, I generated the brightness profiles shown in Figure 9 along the central rock glacier of Mount Sopris.
Figure 9: Transects (left) used to generate brightness profiles (right) with the 3D analyst tool of ArcGIS. Only the first spectral band is displayed for each year. Rock glacier images obtained through Google Earth Pro.
If these profiles accurately identify the ripples in space, we now have data that we can use to quantify them. However, we must acknowledge that measurements obtained using proxies have an associated error. Like any measurement, if the margin of error is manageable, the technique is viable. As the brightness analysis is a new technique and comes with a lot of uncertainty, its effectiveness must be verified.

To run a first-order test and determine whether these curves present reasonable data, I used a very simple comparison of two prominent signals that were well-defined in all three profiles. After picking the best signals, I found their associated distance from the headwall and compared between years by subtracting these distances from one another. Velocity of the signals can then be found by dividing differences in distance from the headwall of the same signal across two images by the time that passed between two images. The brightness method is supported as a viable method if the calculated velocities are reasonable in their magnitude, and in the downslope direction. If the velocities are anomalous in their magnitude, or go against the slope, then the brightness method is not viable. While this test is not quantitatively robust, it serves as a first check in determining whether the brightness method consistently picks up the same features in different images.
Figure 10: Prominent brightness signals and their associated locations on the rock glacier. Imagery obtained through Google Earth Pro.
<table>
<thead>
<tr>
<th>Year</th>
<th>Signal (See figure 10)</th>
<th>Distance from Headwall (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>September 1993</td>
<td>●</td>
<td>1627</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>■</td>
<td>1717</td>
</tr>
<tr>
<td>June 2006</td>
<td>○</td>
<td>1634</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>■</td>
<td>1731</td>
</tr>
<tr>
<td>September 2011</td>
<td>○</td>
<td>1641</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>■</td>
<td>1752</td>
</tr>
</tbody>
</table>

*Table 1: Distance from headwall of prominent signals identified in figure 10*

From these values, I calculated the velocity of the rock glacier as expressed by change in the positions prominent brightness signals:

\[
\begin{align*}
\text{○} - \text{○} & = \frac{0.55 \text{m}}{12.75 \text{years}} \\
\text{○} - \text{○} & = \frac{1.3 \text{m}}{5.25 \text{years}} \\
\text{■} - \text{■} & = \frac{1.1 \text{m}}{12.75 \text{years}} \\
\text{■} - \text{■} & = \frac{4.0 \text{m}}{5.25 \text{years}} \\
\end{align*}
\]

*Equation 1: Calculations for the velocity of the rock glacier at prominent brightness signals.*
These results are reasonable and do not condemn the brightness analysis. Rock glaciers move on the order of ~1 m/yr, which these values accommodate (Kaab et al., 1997). Differences between the values obtained for the same color pairs and different shapes

\[\text{e.g. } \bigcirc - \bigcirc \text{ vs } \square - \square\]

indicates differential velocity within the rock glacier. This is expected because rock glaciers move primarily by internal deformation that varies based on local topography and ice thickness (Arenson et al., 2002). Differences between values calculated for different color pairs and the same shapes

\[\text{e.g. } \bigcirc - \bigcirc \text{ vs } \text{\textcolor{green}{\bigcirc}} - \text{\textcolor{green}{\bigcirc}}\]

indicates variable flow velocity through time, which is commonly attributed to variation in climatic factors (Krainer & Mostler, 2006).

The values calculated for velocity are reasonable, and allow the conclusion that brightness conveys the true position of each rumple. However, the brightness method has an array of problems associated with its implementation.

2.1.2. Systemic Inaccuracies of the Proxy

The shortcomings of the brightness method for conveying the amplitudes of the rumples are obvious. The amplitude of brightness signal is a function of the time of year, time of
day, and albedo of the surface, as well as topography. Therefore, amplitude of variations in brightness do not convey the height of the rumples and is arbitrary with this analysis. For example, winter brightness values cannot be directly compared to summer values. The shaded troughs hold snow longer than the illuminated crests. Snow acts to increase the brightness of the troughs relative to the bare crests, which maintain the same brightness for most of the year. The brightness of the troughs is thus scaled up relative to that of the crests.

Even as a proxy for rumples, their horizontal position is shaky using brightness. While variations in prominent brightness signals through time could be interpreted as change in position of the rock glacier, the data also allows that the interpreted position of each rumple stems from variation in illumination. Because each image was taken at a different time of the day and year, the profile of brightness shifts laterally between images based not on the movement of the rock glacier, but on changes in lighting.

In addition, the use of ArcGIS effectively imposes limits on the reproducibility of profiles collected from one image to the next. The lateral distance of each brightness value from the headwall is entirely dependent on the length of each hand-drawn transect. ArcGIS requires using new a transect for each raster we analyze, which introduces variability to the length scale of each rumple. If any of the transects differed from one another, which they likely did because lines were hand drawn to a loose template, the locations of minimum and maximum brightness values would be variable between different years.
The velocities calculated in section 2.1.1. are therefore contaminated by the variability of lighting from one scene to the next. While the use of brightness is too inconsistent to locate these features in space, brightness may still be useful in identifying crests and troughs. However, for this study we require more precise data for our numerical analyses of the rock glacier rumples.

2.1.3. A Direct Solution

The goal of the brightness analysis was to define precisely the spatial distribution of the features of rock glaciers with a metric other than elevation. However, the locations of the crests and troughs and rumple height are all too contaminated in the brightness metric. To circumvent the problems associated with this proxy, we utilize the capabilities of the open-source NASA stereogrammetry software, Ames Stereo Pipeline (ASP), to create a direct solution.

ASP was developed by the Intelligent Robotics Group at the Ames Research Center to plan long range missions to planetary bodies. Before actually touching down on a planetary surface, satellites are sent to orbit the body and collect preliminary data as a precursor to the mission. The satellites generally retrieve high resolution imagery and low resolution DEMs to identify areas worthy of more thorough study. Once an area is chosen, NASA needs high resolution topographic data to plan and analyze the proposed mission. Because there is no low-altitude instrumentation on the planet, using conventional means like LiDAR to create detailed topographic data is not an option. To
address this problem and create high resolution elevation data, the Ames Research Center created Stereo Pipeline (Moratto et al., 2010).

The reconnaissance satellites record an array of information for each image, including the altitude and azimuth of the satellite and the time of day and year when the images were acquired. When two images of the same location are taken at different times, a stereopair is formed. Ames Stereo Pipeline uses contrasts in the pixels of a stereopair, with the information for conditions at the time of image capture, to create high resolution DEMs that are referenced to the low-resolution DEM (Shean et al., 2016). This routine of using photographs to derive quantitative distances is known as photogrammetry.

While the original intent of Ames Stereo Pipeline was reconnaissance on planets where LiDAR studies are not possible, this program efficiently creates high resolution DEMs here on Earth at a fraction of the cost of LiDAR. There is a wealth of high resolution satellite imagery for the surface of Earth, including Mount Sopris, that make the capabilities of ASP very valuable to this study.

We use ASP on two stereopairs that were created using Digital Globe imagery from June 2002, September 2011, and August 2012. Pairing these 1 m resolution images in any combination did not cover the entire mountain, meaning two separate DEMs had to be created and later mosaicked together. Based on overlap of the imagery, I created the first DEM from the June 2002 and August 2012 images and the other from the September 2011 and August 2012. Digital Globe provides these images in a suite of
separate individual strips. Mount Sopris is contained within two separate strips for June and August suites, which required we use the dg_mosaic tool to stitch the images together.

With the entirety of Mount Sopris contained in one image for each year, we use the mapproject function to roughly reference all of the images to ground coordinates obtained from the 10 m DEM. Next, we orthorectify the images with rational polynomial coefficients. This allows ASP to estimate the ground location of each pixel by giving the program rough constraints for pixel locations within each photo.

Having two pairs of images for each side of Mount Sopris mosaicked and projected, we use the stereo function to generate elevation data. After the long processing time of this step (~12 hours for each DEM), we have two sets of point cloud data that cover the entire Mt Sopris massif. We finally rasterized this point cloud data and projected it using a UTM zone 13N projection to make this data usable in GIS programs. I also chose the desired resolution for the output DEM in this step. A good rule of thumb is to make the resolution of the DEM about twice the resolution of your imagery (Armstrong, 2017). The Digital Globe imagery was taken at a 1 m resolution; therefore, we chose to generate a DEM of 2 m resolution. Now we have two DEMs in the right format (.tif) to import into ArcGIS and mosaic into one.
Figure 11: Visual comparison of the high resolution 2 m DEM generated using ASP to the 10 m DEM obtained from the national map. The pixelated boxes on the 2 m DEM are raster cells where ASP could not calculate elevation.

With elevation data that detects the rumbles, we use the 3D analyst tool in ArcGIS to extract a representative elevation profile along the rock glacier. Smaller features like slumps and boulders are detected by the high-resolution DEM, and create noise in the profile. A common method used to deal with noise in geomorphology is a swath profile, which computes an average of raster values perpendicular to a chosen transect being analyzed (Bishop et al., 2003). We chose a 10 m swath (10 m on either side of the transect) to account for noise without damping the signal of the rumbles by moving beyond the curvature of a given rumple.
This generated a profile that detects the rumples with greatly reduced noise. I then employed an array of numerical methods to understand the spatial distribution of the rumples.
2.2. Numerical Analysis: Quantifying the Rumpled to Estimate their Spectral Density

Using ASP, we generated a high enough resolution DEM to detect the characteristic rumplings on the central rock glacier of Mount Sopris. We now seek to quantify the dominant spacing of rumplings. Treating these structures as waves that propagate from the valley headwall to the terminus, we can analytically establish the dominant wave length of the system to determine patterns in the prevailing spacing between rumplings.

2.2.1. Detrending the Elevation Data

Before we analyze the rumplings, we must remove the underlying slope of the landscape and of the rock glacier surface to situate these features on a horizontal axis. Removing slope filters noise from the wave analysis by focusing the rumplings as the sole contributor to variation from a flat profile.
First, we smooth the elevation data to create a rumple-free profile of the valley. We then subtract the smooth profile from the high-resolution swath profile, which presents our data in a detrended profile. This effectively removes the underlying slope while preserving the elevation of the rumbles as variance from a horizontal line. Harnessing the computing power of Matlab, we used a moving average along 15 points on the profile to smooth out the rumbles. We use 15 points because this value removes the underlying slope without removing the smaller rumbles. This step requires discretion, because choosing too large a value will bias the data toward a larger wavelength by removing the smaller rumbles. If we choose too small a value, the underlying slope will not be completely removed. The smoothing function pairs averaged elevation values with each raster transected by the profile. Because the transect is 1898 m long at a 2m resolution, there are 948 raster values plotted on the x-axis of the smoothed elevation plot.
2.2.2. Defining a Dominant Topographic Wavelength

With the underlying slope of the valley removed from our elevation profile, topographic crests and troughs are clearly expressed as the only topographic features (Figure 16). The isolated rumple elevation profile creates a signal across the rock glacier. To analyze the dominant wavelengths embedded in this signal (i.e. the dominant rumple spacing), we use Matlab to generate a periodogram (from the detrended profile) and estimate the power spectral density for this system. Most spectral density estimates are performed on a signal with an amplitude that varies in time (i.e. a frequency), but the method will also quantify our signal, whose amplitude varies in space (Stoica et al., 2005).

![Periodogram Power Spectral Density Estimate](image)

*Figure 15: Power spectral density estimate for the detrended rumple profile of the central rock glacier. The Power/frequency is given on a log scale.*
We convert between the normalized frequency output by the periodogram and a length scale in standard units using the equation:

\[ \lambda = \frac{2\mu}{\phi} \]

*Equation 2: Conversion between normalized frequency (\(\phi\)) and wavelength (\(\lambda\)), where \(\mu\) is resolution of the dataset.*

where \(\lambda\) is the dominant wavelength, \(\mu\) is resolution of the data, and \(\phi\) is the normalized frequency value corresponding to the peak power/frequency. Note that the units of wavelength are completely dependent on the data, as the normalized frequency is unitless.

*Figure 16: Close up of the dominant normalized frequencies. The dominant frequency, 0.119, is almost an order of magnitude more prominent than the next most important frequencies.*
Based on this periodogram, we find the dominant normalized frequency to be 0.119, which we use for $\phi$ in equation 2 with a 2-m resolution to yield the estimate of 34 m as dominant wavelength.

**2.3. Discussion**

We use elevation to escape inconsistencies of the brightness method by directly quantifying the topography of the rock glacier at a 2 m resolution. Shean et al. (2016) employed a series of tests to analyze the accuracy of Ames Stereo Pipeline (ASP) in polar regions, finding that the program creates very reliable high resolution DEMs from satellite imagery. Shadows created by very steep slopes, clouds obstructing the imagery, and vegetation were all found to pose challenges to ASP. Fortunately, ASP did not face such challenges in this study, as the rock glacier lies on a gradual slope, free of vegetation, and the imagery was collected on clear days. In addition, if ASP cannot calculate the elevation for a cell, the cell is filled with low resolution data from the existing DEM, and not with faulty guesses. The cells that cannot be calculated are obvious when viewed in a GIS program. This gives us confidence that the data-filled cells can be trusted as accurate depictions of elevation in the analysis.
The calculated wavelength of 34 m for the central rock glacier does not allow discussion of rock glacier rumple spacing at large. To place this wavelength in the context of other systems, we use the same analysis on other rock glaciers contained within a 0.5 m DEM of rock glaciers outside of Laurichard France.
This DEM was originally collected with LiDAR to study rock glacier flow morphology associated with internal deformation (Bodin et al., 2008). The DEM was graciously provided by Station Alpine Joseph Fourier, CNRS Université Grenoble-Alpes. We analyzed two of the rock glaciers contained in this DEM using a 7 m swath orthogonal to the rumples. We again smooth both profiles along 15 points to remove the underlying topography without biasing the data toward a larger wavelength.
Figure 19: Top depicts the detrended elevation profile for the Western Laurichard rock glacier. Periodogram generated with the detrended profile shown next. Bottom figure emphasizes the periodogram’s dominant power frequency (0.0910).

After detrending the profile, I generated the periodogram illustrated in Figure 21. The highest power/frequency lies around 0.0910, which I used in equation 2 with a 0.5 m resolution to find a dominant wavelength of 11 m.
Using the same approach, I also calculated 11 m for the dominant structural wavelength of the eastern rock glacier. We cannot conclude this wavelength is identical between the two rock glaciers, because the resolution of the data gives a 0.5 m error on either side of this value. Nonetheless, the wavelengths are remarkably close to one another. The proximity of these values is very interesting, even more so when compared with the much larger wavelength of found on Mount Sopris.
We need data for the wavelengths of rumples across many more rock glaciers to draw conclusions for the cause of this difference. The local lithology, local climate, variations from local climate, slope and height of the overlying headwall, slope and width of the valley, velocity of the rock glacier, thickness of the ice, and orientation of the rock glacier must all be considered as factors that influence the structural wavelength.

Unfortunately, the DEM generated by ASP for Mount Sopris is missing data in many of the raster cells located on the other rock glaciers. We cannot use this data to analyze the rumples of these rock glaciers because the missing pixels are filled with low-elevation data from the 10 m DEM, and do not accurately reflect elevation at a 2 m resolution. This creates large variation from the general topographic profile, which these methods cannot distinguish from rumples.

![Figure 21: Problems associated with raster cells ASP does not define elevation for. Undefined cells, depicted as cloudy grey splotches on the Sopris hillshade, create erratic spikes in the elevation profile, which are interpreted as very large rumples in the elevation profile.](image)

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Chapter 3: Developing a Qualitative Understanding of the Avalanche Cone through Time-Lapse Photography

3.1. Field Methods

In addition to gathering data for the spatial distribution of rumbles, we seek a qualitative understanding of how the avalanche cone that serves as the source area for the rock glacier develops through the winter. Existing models demand a constraint on the timing of material being delivered to the avalanche cone, which we define using time-lapse photography. With this tool, we monitor the composition of material being delivered to the avalanche cone on a three-hour interval.

On October 8 2016, Katie Pellicore and I installed a rugged 12-megapixel Moultrie time-lapse camera on the east ridge of the central valley of Mount Sopris. I chose a snow-free location for the camera using Google Earth imagery taken in early April of 2015. Peak snow accumulation in the Colorado alpine is generally around April 1, which makes April imagery representative of places that are typically bare (Barlage et al., 2010).
On this clear Autumn day, we hiked to Thomas Lake (via the Thomas Lake Trail), and ascended the central valley of Mount Sopris along the eastern crease of the rock glacier. With no trail after Thomas Lake, we moved slowly over loose rocks that cluttered the remaining 2 km. Because the coordinates I had chosen for placement of the camera had an obstructed view of the rock glacier, I installed the camera further up the ridge. The new location was free of snow in the April 2015 imagery, and has a clear line of sight to the avalanche cone. We stabilized the 1m tripod with rocks and attached the camera. I set the time-lapse for every three hours starting at 3 pm that day.

On New Year's Day 2017, I returned to the camera using a splitboard. I originally planned to swap out the memory card and change the camera batteries, but I opted to leave everything with the intention of preserving the time-lapse settings. Unfortunately,
while I was checking the battery and memory capacity of the camera, I changed the time-lapse to start an hour and forty-nine minutes later, at 1:49 instead of noon.

My last journey to the camera was on February 25, 2017 with Garret Hammack. On this mission, I changed the memory card to gather the initial results, while leaving the instrumentation to finish collecting images through the winter. I also hoped to change the batteries, which are only accessed by removing the camera from the tripod, but the camera was frozen in place. Because I did not want to risk breaking the mount, I left the batteries, which were still at a 60% charge.

![Figure 23: Time-lapse camera installed on Mount Sopris. Left to right: October 8 2016, January 1 2017, and February 25 2017.](image)

### 3.2. Discussion: Qualitative Observations Derived from the Time-Lapse

Between October 8, 2016 and February 25, 2017, our camera collected 1,142 pictures, 569 of which were taken during the day. At the time of this study, the camera remains on Mount Sopris to record the avalanche cone through the remainder of the winter.

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While we do not currently have access to data for an entire winter, we can begin to draw conclusions about the development of the avalanche cone.

![October 7, 2016](image)

Figure 24: First image collected by the time-lapse camera.

From the four and a half months of data available at this time, we find that large rockfall onto the surface of the avalanche cone does not occur during the winter. A potential explanation for this is that the rocks are frozen in place by the cold grip of winter, and are only released from the bedrock when temperatures warm the headwall. Hopefully, successful image collection continues through the spring, and will provide insight into when large rocks are delivered to the avalanche cone.
Humlum et al. (2007) found that rock is delivered to avalanche cones throughout the year in Svalbard, with the concentration and size of rocks increasing as temperatures rise. These observations do not necessarily contradict our interpretations of the Sopris time-lapse. Rocks delivered during the winter could be small fragments that are intermixed with snow. From our time-lapse, which was about 400 m from the avalanche cone, we may not be able to identify smaller rocks that are mixed in with the cone. It therefore appears that deposition of the large rocks that cover the Mt Sopris rock glaciers is not occurring in the winter, but we cannot make claims on smaller rocks mixed in with avalanched snow. In future studies, I advocate digging a trench through the avalanche cone to create a stratigraphic column that would better constrain the concentration of rocks in the avalanche cone in this setting.
Chapter 4: Conclusion

While our data does not allow us to constrain the origin of rock glacier rumbles, we provide high resolution elevation data for Mount Sopris and numerical data that future studies must honor when drawing hypotheses of rumple origin. With data obtained in this study, we conclude that the wavelength of rock glacier rumbles is not universal. It varies, and is likely dependent on factors inherent to the local environment of each rock glacier. Although we also discredit brightness as a proxy for rumple position, meaning that it cannot be used to constrain glacier speed, in snow-free imagery we can effectively identify crests and troughs as local maxima and minima in brightness, allowing measurement of wavelengths.

Our time-lapse photographic record of the avalanche cone at the head of the central valley of Mount Sopris constrains the delivery of large rocks to the source area of the glacier. At the time of image collection, there is no sign of the large rocky debris being delivered to the avalanche cone. We suspect that delivery occurs later in the winter and spring seasons.

Together, these numerical and qualitative constraints will inform future numerical models of rock glaciers by better illuminating natural features that arise from harmony between rock and ice in alpine landscapes.
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