Glacier sliding from space: multi-scale remote sensing, geodesy, and numerical modeling to understand glacier mechanics

by

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Abstract

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Glacier sliding from space: multi-scale remote sensing, geodesy, and numerical modeling to understand glacier mechanics

Thesis directed by Professor Robert S. Anderson

Glacier basal sliding is a poorly understood aspect of glacier mechanics, and its spatial and temporal distribution affects glacier change and the evolution of alpine landscapes. In these studies, we use on-glacier GPS, moderate- and high-resolution optical satellite imagery, and numerical ice flow modeling to investigate the mechanics of glacier sliding across a variety of scales.

First, we employ on-glacier GPS to investigate the intimate link between subglacial water pressure and the rate of glacier sliding in response to the onset of spring melting on Kennicott Glacier, Alaska. We find large diurnal glacier velocity fluctuations during times of high and rising water level on a well-connected ice-marginal lake. The ice surface speedup at an up-glacier station is first driven by longitudinal coupling to down-glacier ice, but then evolves to respond to local basal conditions.

We then utilize high-resolution WorldView satellite imagery to document the spatial pattern of the seasonal evolution of ice surface velocity over the 45 km² terminal reach of Kennicott Glacier. We develop a numerical ice flow model to explore the distribution of basal sliding required to explain the observed surface speedup. We find the ice surface speedup is insensitive to the exact distribution of basal sliding, which may allow for simpler sliding parameterizations in glacier models.

Finally, we employ Landsat 8 satellite imagery to characterize the spatial patterns of glacier sliding over a 45,000 km² area covering 64 glacier longitudinal profiles from ice divide to terminus. We find the entire ablation area of glaciers speeds up in a uniform manner, with the speedup magnitude insensitive to winter surface speeds. These patterns of sliding may drive patterns of glacier erosion that leads to the formation of icefalls.
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Chapter 1

Introduction

1.1 Motivation for studying glacier basal motion

Better understanding of the mechanics of glacier basal motion is required for a complete description of the processes governing glacier behavior. In the absence of an understanding of glacier basal motion, we cannot accurately predict how glaciers will change in the future or how they have shaped landscapes in the past.

1.2 Mechanics of glacier basal motion

1.2.1 Modes of glacier motion

Spatial variations in glacier ice surface elevation give rise to spatial gradients in potential energy that drive ice flow. This surface-slope-induced “driving stress” is balanced by viscous stresses due to deformation both vertically and transversely throughout the ice body, and by friction at the basal ice-rock interface (Figure 1.1). It is this last process, basal motion, that is the focus of this thesis.

The one-dimensional shallow ice approximation for glacier velocity is given by,

\[
u(z) = \frac{2A}{n + 1} \left[ \rho g \sin \alpha \right]^n \left[ H - z \right]^{n+1} + \frac{u_b}{n + 1} \quad \text{(1.1)}
\]

where \( u \) is longitudinal glacier velocity, \( z \) is the vertical coordinate (taken to be equal to 0 at the ice surface, and positive downwards), \( A \) is the Glen’s flow law rate factor (which scales the strain rate induced by a given stress, and can be thought of as “ice softness” or inverse viscosity; for temperate ice \( A \approx 2.4 \times 10^{-24} \text{ Pa}^{-3} \text{ s}^{-1} \) \cite{Cuffey2010}), \( n \) is the Glen’s flow law exponent (which governs the non-linearity of the stress-strain rate constitutive equation; depending on the mode of ice formation \( 1 \leq n \leq 4 \) \cite{Cuffey2010}, although \( n = 3 \) is generally assumed for models of bulk ice flow), \( \rho \) is the
Figure 1.1: Schematic of the modes of glacier motion. a) The straining of originally straight vertical and transverse lines due to viscous deformation throughout the column of glacier ice. The deformation of the red line reflects the ice velocity field, $u(z) \propto (H - z)^4$. b) Deformation of originally straight vertical and transverse lines due to viscous ice deformation (causing shape change) and basal motion (causing translation). Two tents and a GPS monument rest on the ice surface.
column-averaged density of glacier ice and overlying snow and firn \((\rho = 917 \text{ kg m}^{-3} \text{ for bubble-free glacier ice})\), \(\alpha\) is the ice surface slope, \(H\) is the ice thickness, and \(u_b\) is the basal velocity.

Equation 1.1 is derived by first stating that the deviatoric stress (i.e., stresses that cause shape change, here denoted \(\tau_{xz}\)) scales with the weight of the ice overburden, given by,

\[
\tau_{xz}(z) = (\rho g \sin \alpha) z
\]

where, again, \(z = 0\) at the ice surface and increases downwards. This deviatoric stress induces deformation throughout the ice column. The strain rate \((\dot{\epsilon}_{xz} = \frac{\partial u}{\partial z})\) is related to the imposed stress by the constitutive relation, given by,

\[
\frac{\partial u}{\partial z} = A\tau_{e}^{n-1}\tau_{xz} = A\tau_{xz}^{n-1}
\]

where \(\tau_{e}\) is the “effective stress” \((\tau_{e}^2 = 1/2 [\tau_{xx}^2 + \tau_{yy}^2 + \tau_{zz}^2 + \tau_{xz}^2 + \tau_{xy}^2 + \tau_{yz}^2])\), which represents the summed influence of stresses in all directions and accounts for the strain-rate weakening rheology of glacier ice. The simplification after the second equality in Equation 1.3 is only strictly true if \(\tau_{e} = \tau_{xz}\), and is approximately true for laterally confined \((\tau_{yy} = \tau_{yz} \approx 0)\), longitudinal uniform \((\tau_{xx} = 0)\) flows, with large width/depth ratios \((\tau_{xy} \approx 0)\).

It is apparent from Equation 1.1 that the surface velocity \((u_s = u(z = 0))\) is the linear superposition (i.e., sum) of the velocities due to internal deformation and basal motion. This is not strictly true for two main reasons, aside from the assumptions listed above. These reasons are: 1) Internal deformation and basal motion are intimately linked. Deformation throughout the ice column only occurs because the basal interface is coupled to underlying material by frictional stress \((\tau_b)\). If the basal plane were entirely frictionless, the overlying glacier ice would slide downhill as a coherent block, with no deformation occurring throughout the ice column. In more realistic scenarios, smaller \(\tau_b\) may cause higher rates of basal slip, although this would be partially offset by a reduction in internal deformation; and 2)
Stress-gradient coupling to adjacent ice in the face of spatially non-uniform basal motion causes the ice surface speed to depart from the simple sum of deformation and basal motion (Section 1.2.4). In effect, the signal of basal motion is “smeared” or “diffused” by this stress coupling.

1.2.2 Link between glacier hydrology and basal motion

The rate of basal motion depends sensitively on the state of the subglacial hydrologic system, which evolves on an annual basis. There are two fundamental modes of subglacial drainage: 1) an inefficient, poorly connected, “distributed” drainage system in which water pressure increases with the rate of subglacial discharge (Figure 1.2a; [e.g., Walder and Hallet, 1979]), and; 2) an efficient, well-connected, “channelized” drainage system in which water pressure decreases with the rate of subglacial discharge (Figure 1.2b; [Rothlisberger, 1972]).

The state of the subglacial drainage system is not static, and is thought to evolve from...
Figure 1.3: Subglacial drainage evolution schematic. a) Dissipation of kinetic energy due to water’s viscous resistance to flow generates heat which is applied to enlarge the containing channel by wall melt. b) Ice overburden pressure causes viscous collapse of channel walls if the rate of creep closure exceeds that of dissipative wall melt.

This evolution is caused by dissipative heat generation by water’s viscous resistance to flow. This heat is applied to the ice walls confining the subglacial channel, which causes channel growth and increased ability to transmit meltwater (Figure 1.3a). This process for subglacial channel growth is opposed by the viscous collapse of channel walls due to the ice overburden pressure (Figure 1.3b). This summary necessarily simplifies known complexities, such as the apparent non-arborescence of channelized subglacial drainage [e.g., Harper et al., 2005] and the existence of “unconnected” or “isolated” basal water [Hoffman et al., 2016].

The presence of subglacial water reduces the frictional stress exerted by underlying rock on basal ice. In the absence of widespread subglacial water, glacier ice “slides” via regelation [Weertman, 1957], in which ice melts on the stoss side and refreezes on the lee side of bedrock obstacles. This process is allowed by the unusual inverse relationship between pressure and the melting point of ice. Where pressurized subglacial water exists, it acts to resist the weight of the ice overburden and reduces the effective pressure exerted by the glacier on its bed. Frictional resistance scales with an object’s normal stress, so high water pressure...
(low effective pressure) reduces basal friction, causing the rate of basal motion to increase [Lliboutry, 1968; Kamb, 1970]. We refer to this process as “hydro-sliding”. Early field [Iken, 1981; Iken and Bindschadler, 1986] and laboratory studies [Budd et al., 1979] confirmed theoretical expectations, finding an inverse relationship between the rate of basal motion and the pressure of subglacial water. These observations gave rise to a simple sliding relation

\[ u_b \propto \tau_b^p (P_i - P_w)^{-q}, \tag{1.4} \]

where \( u_b \) is the rate of basal motion, \( \tau_b \) is the basal shear stress, \( P_i \) and \( P_w \) are ice overburden and subglacial water pressures, respectively, and \( p \) and \( q \) are empirical positive constants, usually taken to be \( p = 3 \) and \( q = 1 \) [Budd et al., 1979]. For this summary, we omit discussion of problems with Equation 1.4, namely that \( \tau_b \) increase without limit and that \( \tau_b \) is likely multivalued with respect to \( u_b \) [Schoof, 2005; Gagliardini et al., 2007; Pimentel et al., 2010]. Other relations have been proposed, where \( u_b \) scales with subglacial water volume, rate of change of subglacial water storage [e.g., Bartholomaus et al., 2008], or the proportion of bed covered by water [e.g., Howat et al., 2008]. Despite these works, other studies have found complicated or no links between subglacial water pressure and the rate of basal motion [Harper et al., 2007], which may reflect the high spatial variability of the basal water pressure field [Andrews et al., 2014].

1.2.3 Faster glaciers in a warmer world?

The discovery that the ice sheets could respond just as sensitively to water inputs as alpine glaciers [Zwally et al., 2002] spurred the notion that ice sheets did not respond to climate warming slowly and only by ice loss due to enhanced negative surface mass balance. van de Wal et al. [2008] echoed these findings, showing factor-of-four speedups across wide areas of the Greenland Ice Sheet in response to surface meltwater inputs. Parizek and Alley [2004] incorporated a hydro-sliding feedback into a numerical glacier flow model to show that the link between hydrology and glacier basal motion could result in \( \sim 5 - 20 \text{ cm more} \) sea level rise from the Greenland Ice Sheet than a model not incorporating this feedback.
However, more recent work has shown that summer velocity increases due to enhanced melt may be more than compensated for by especially slow winter speeds following hot summers [Tedstone et al., 2013; Van De Wal et al., 2015]. Indeed Burgess et al. [2013a] showed that winter speeds are inversely correlated with the preceding summer’s melt magnitude across a large range of Alaska glaciers.

The proposed physical explanation for these observations is that enhanced summer melt allows for widespread and complete development of efficient, channelized subglacial drainage. The very low pressure of these large conduits would draw down water from the distributed system, dewatering large portions of the glacier bed. The following fall and winter would thus have small volumes of water stored in the distributed system and thus high average ice-bed friction and lower rates of winter basal motion. This negative feedback may stabilize glaciers from a runaway hydro-sliding driven collapse in the face of climate warming.

There are, however, other processes that may cause faster rates of dynamic ice mass loss in a warming world. One of these mechanics is dubbed “cryohydrologic warming” [Phillips et al., 2010], in which the latent heat released by meltwater refreezing within and beneath the glacier (e.g., in moulins or crevasses) raises the surrounding ice temperature and subsequently reduces its effective viscosity (Equation 1.3). Cryo-hydrologic warming of deep ice is required to explain modern Greenland ice surface velocities [Phillips et al., 2013] and englacial temperature profiles [Lüthi et al., 2015], confirming the active presence of this process. If continued cryo-hydrolgic warming of Greenland basal ice continues, it could cause a 5% loss of ice volume and yield an additional ~ 30 cm of sea level rise over the next 500 years [Colgan et al., 2015].

In addition, supraglacial melt ponds can hydrofracture through km of cold glacier ice [Tsai and Rice, 2010] and cause large, but short-lived ice surface speedups [Das et al., 2008; Hoffman et al., 2011; Joughin et al., 2013]. The inland expansion of supraglacial lakes in a warming climate [Leeson et al., 2014] may provide another mechanism by which ice velocities within the Greenland interior may accelerate in the future, although the compressive stress regime in the interior currently may prevent lake hydrofracture to establish surface-bed
connections [Poinar et al., 2015]. However, less compressive (or extensional) strain rate transients associated with other processes [Stevens et al., 2015] or evolution of the mean surface velocity field may allow for inland propagation of supraglacial lake drainage.

Aside from the above-discussed mechanisms that alter the local glacier force balance, terminus-forced perturbations can also cause inland glacier acceleration and enhanced mass loss. For example, inland glaciers accelerate following the loss of terminal ice shelves [Scambos et al., 2004; Rignot et al., 2004] that provided resistance to seaward ice flow. Ice shelf collapse is often induced by surface-melt-pond-driven hydrofracture [MacAyeal et al., 2003; Scambos et al., 2009], which may cause rapid mass loss from Antarctica over the coming centuries [Pollard et al., 2015]. These terminus-forced perturbations can readily propagate inland if the associated glacier thinning field acts to reduce basal friction at a faster rate than it decreases the gravitational driving stress [Pfeffer, 2007].

1.2.4 Expression of basal motion at the glacier ice surface

Basal motion is hidden from the surface-based observer by \( \sim 10^1 - 10^3 \) m of viscously deforming ice. Balise and Raymond [1985] analyzed the transmission of basal velocity perturbations to the ice surface using a linear viscous glacier model. The authors showed that, at intermediate length scales (5 – 10 ice thicknesses), the ice surface speedup should be a diffused, lower amplitude representation of the basal velocity perturbation (Figure 1.4). Gudmundsson [2003] extended this work to include variations in basal topography as well as two-dimensional effects in a linear viscous glacier model. He found that short wavelength (\( \lambda \)) basal perturbations were completely absorbed by enhanced deformation in adjacent ice due to the associated high strain rates and hence low effective basal viscosity (Figure 1.5). As the perturbation wavelength increases, the magnitude of the surface speedup approaches that of the basal sliding perturbation (Figure 1.5). Only at this point (\( \lambda \rightarrow \infty \)) is the surface velocity the linear superposition of the locally-calculated deformation velocity (i.e., a function only of local ice thickness and surface slope) and rate of basal slip. At smaller \( \lambda \), the ice surface speedup is a spatially diffused, lower magnitude cousin of the associated increase in the rate of basal motion.
Figure 1.4: Expression of a step change in basal velocity when observed at the ice surface. The heavy dashed blue line ($u_{bed}$) represents the basal velocity, the heavy solid red line ($u_{surface}$) shows the associated, “diffused” ice surface velocity. The thin dashed gray line showed the ice surface vertical velocity field. Modified from Balise and Raymond [1985].

Figure 1.5: Ratio of surface speedup ($\Delta u_s$) to basal velocity perturbation ($\Delta u_b$) as a function of basal perturbation wavelength ($\lambda$) normalized by ice thickness ($H$). Modified from Gudmundsson [2003].
1.3 Methods of monitoring glacier velocity fluctuations

1.3.1 On-glacier surveying utilizing the Global Positioning System

We employ on-glacier geodetic-grade global positioning system (GPS) units to measure glacier displacement with high spatial and temporal resolution. To triangulate a user’s position, a typical handheld GPS uses only pseudorange data (i.e., one-way travel time from the transmitting satellite to GPS receiver) of only one frequency at which GPS satellites transmit. The long wavelength of the psuedorange code and influence of atmospheric time delays limit the accuracy of these units. Our geodetic-grade GPS units use the carrier phase of both frequencies at which the GPS satellites transmit code. This method produces higher accuracy position solutions than pseudorange because: 1) the carrier phase is much shorter wavelength (\(\lambda \approx 20 \text{ cm}\)) than pseudorange, and 2) employing both frequencies of code transmission allow for the removal of ionospheric time delays. In addition, we differentially process GPS data relative to an unmoving base station, which allows us to remove the influence of near-surface stratospheric time delays. With these processing techniques, our on-glacier antenna position estimates are accurate to \(\pm 2 \text{ cm}\) in the horizontal plane and \(\pm 4 \text{ cm}\) vertically. This error is usually normally distributed about the true position and we can increase our accuracy by averaging our 30 second observations over 15 minute intervals (Section 2.2.3).

1.3.2 Cross-correlation of optical satellite imagery

Cross-correlation (a.k.a. pixel tracking, speckle tracking, or feature tracking) of satellite imagery provides spatially extensive displacement fields. Image cross-correlation estimates the pixel offsets required to align distinctive pixel chunks (“chips”) between two images (Figure 1.6; [Scambos et al., 1992]). If the two images are well geolocated (i.e., pixels are accurately mapped to their ground coordinates), the offsets reflect displacement of the objects contained within the image. We apply this technique to glaciers, so the displacements we estimate are those due to glacier motion over the time separating the two images. Although the fundamental unit of measurement in satellite imagery is the pixel, image cross-correlation
can estimate displacements with higher spatial resolution than the pixel resolution [Scambos et al., 1992]. The cross-correlation routines we employ fit the correlation surface using a two-dimensional Gaussian function. The resolution of this Gaussian is not tied to the image resolution, and thus can locate displacement peaks between integer pixel offsets, providing much higher accuracy measurements [Fahnestock et al., 2016].

1.3.3 Tradeoffs in resolution and extent

We exploit different observation technologies and utilize the varied spatial extents and temporal resolutions of these datasets to address several questions of interest (Table 1.1). On-glacier GPS provide very high spatial and temporal resolution horizontal velocity data and constrain vertical ice uplift, but are point observations and therefore provide poor spatial context. WorldView imagery provides high resolution spatial context and allows the robust correlation of slowly-moving glaciers over short timespans, but requires extensive pre-processing and covers a relatively small spatial area. The moderate resolution of Landsat 8 imagery allows for glacier velocity estimation over large areas and requires minimal pre-processing, but has a higher noise level than velocity estimates derived from WorldView imagery, which limits the method’s effective temporal resolution.
Table 1.1: Spatial and temporal resolution, as well as spatial extent and measurement density, for the datasets used in these studies. $\Delta x =$ spatial resolution. $x$, $y$, and $z$ reflect spatial resolution in the horizontal ($x-y$ plane) and vertical directions, respectively. $\Delta t_{\text{min}}$ is the theoretical minimum time between observations. Measurement density shows the number of observations per square kilometer (Landsat8 and WorldView) or per kilometer (GPS). Spatial extent shows the area (Landsat8 and WorldView) or linear distance (GPS) covered by our study transect (GPS) or a single image footprint (Landsat8 and WorldView).

<table>
<thead>
<tr>
<th>Dataset</th>
<th>$\Delta x$</th>
<th>$\Delta t_{\text{min}}$</th>
<th>Measurement density</th>
<th>Spatial extent</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>[m]</td>
<td>[d]</td>
<td>[km$^{-1}$] or [km$^{-2}$]</td>
<td>[km] or [km$^2$]</td>
</tr>
<tr>
<td>GPS</td>
<td>0.02 ($x/y$) / 0.04 ($z$)</td>
<td>0.01</td>
<td>$3.9 \times 10^{-1}$</td>
<td>$1.3 \times 10^1$</td>
</tr>
<tr>
<td>WorldView</td>
<td>0.5 ($x/y$) / NA ($z$)</td>
<td>3.7</td>
<td>$4.0 \times 10^6$</td>
<td>$3.3 \times 10^2$</td>
</tr>
<tr>
<td>Landsat8</td>
<td>15 ($x/y$) / NA ($z$)</td>
<td>16</td>
<td>$4.4 \times 10^3$</td>
<td>$3.1 \times 10^4$</td>
</tr>
</tbody>
</table>

1.3.4 Thesis structure

This thesis is not ordered in the chronological progression in which I undertook studies, but rather is structured to expand from high temporal and spatial resolution to coarse spatial and temporal resolution but large spatial extent. In Chapter 2, I explore on-glacier GPS time series and hydrology data to document the link between subglacial water pressure and glacier motion, as well as the evolution of links between glacier speed across a 10 km longitudinal profile during the onset of the 2012 and 2014 melt seasons. In Chapter 3, I observe the spatiotemporal distribution of ice surface velocity over the terminal reach of Kennicott Glacier over the 2013 melt season. I then employ numerical models to constrain the possible distributions of basal motion required to explain the observed surface speedup. In Chapter 4, I utilize Landsat 8 imagery to characterize the spatial patterns of glacier summer speedup on a mountain-range-wide scale and seek physical explanations for their spatial distributions.
Chapter 2

Hydro-sliding and the springtime dynamical evolution of
Kennicott Glacier, Alaska

Abstract

Glacier basal motion is a poorly understood aspect of glacier mechanics that is responsible for the majority of ice flux on fast-flowing glaciers, enables rapid changes in glacier motion, and provides the means by which glaciers shape alpine landscapes. We collect hydrometeorologic data and GPS-derived ice surface motion to probe the link between subglacial water pressure and the evolution of glacier velocity. We find a chaotic timeseries of > 50 m fill-and-drain sequences on the well-connected ice-marginal Donoho Falls Lake. Glacier velocity in the down-glacier reach responds sensitively to lake stage, with high amplitude diurnal velocity fluctuations during high or rising stage. The timing of velocity peaks precedes peak stage by 2-3 hours, and synchronously shifts earlier in the day throughout our observation period. We find the up-glacier station appears to first speed up in response to longitudinal coupling with accelerating down-glacier ice before responding to local variations in basal traction. We find the transition to responding to local basal conditions results in the glacier behaving more uniformly, with similar magnitude diurnal velocity fluctuations, synchronous timing of velocity extrema across the 10 km study reach, and steadier longitudinal strain rates.

This chapter is in preparation for submission:
2.1 Introduction

Sub-annual glacier surface velocity changes are driven by variable rates of basal motion [cf. Willis, 1995, and references within]. Basal motion, in turn, is driven by subglacial water pressure, with high subglacial water pressure [e.g., Iken and Bindschadler, 1986] or rate of change of water storage [e.g., Bartholomaus et al., 2008] corresponding to times of rapid basal motion, which may be accomplished by slip at the ice-rock [Weertman, 1957; Lliboutry, 1968], slip at the ice-till interface, or by deformation within underlying till [Truffer et al., 2000]. Irrespective of the exact physical mechanism, as water pressure approaches the ice overburden pressure, basal resistance approaches zero and the driving stress must be balanced by longitudinal and transverse stresses transmitted to adjacent ice that is more tightly coupled with underlying bedrock [e.g., O’Neel et al., 2005; Price et al., 2008; Armstrong et al., 2016].

Several recent field studies have documented the important, and often overlooked role of longitudinal stress transfer in controlling glacier ice flux. Ryser et al. [2014] showed that horizontal stress transfer between time-variable “sticky patches” (i.e., regions of high basal friction) and surrounding “slippery patches” controlled the spatial pattern of ice motion over a large portion of the Greenland ice sheet ablation zone. On a smaller scale, Flowers et al. [2016] showed high sensitivity of the rate of basal motion to water inputs at one station drove surface speed variations at other stations which were less sensitive to local meltwater inputs.

The importance of longitudinal stress gradient coupling can vary throughout the year due to the spatiotemporal evolution of subglacial drainage and the sensitivity of basal motion to meltwater inputs. Borehole water pressure often record large variations in the magnitude and temporal behavior of subglacial water pressure over short horizontal spatial scales (< 20 m; [Harper et al., 2002, 2005]) and often exhibits a complicated, if any, link to surface velocity changes [Harper et al., 2007]. Andrews et al. [2014] found that water pressure in moulins correspond well over large horizontal distances and are well correlated with short-term velocity changes. That moulin water pressures better explain surface velocity variations than
borehole water pressures may reflect that moulins serve as water pressure “integrators” and
represent water pressure averaged over a length scale large enough to be relevant for glacier
force balance [Andrews et al., 2014]. The authors argue that diurnal velocity fluctuations
are controlled by pressure variations in an overwhelmed efficient subglacial drainage system,
although long-term (~ weeks) velocity decrease is associated with decreasing areal extent of
“unconnected” subglacial water [Hoffman et al., 2016].

In this study, we use a well-connected ice-marginal lake to document the link between
subglacial water pressure and glacier basal motion, as well as the evolution of longitudinal
ice coupling, during the establishment of efficient subglacial drainage through the onset of
the melt season on Kennicott Glacier, Alaska (Figure 2.1). We first document the history
of subglacial water pressure, as reflected by stage (i.e., water depth) on Donoho Falls Lake
(Section 2.3.1). We then document the evolution of ice surface velocity, ice uplift, and
longitudinal strain rates over a 10.3 km centerline profile (Section 2.3.2). We then document
links between ice surface velocity and subglacial water pressure (Section 2.3.4) and evolving
relationships between velocity behavior along the centerline profile (Section 2.3.3).

2.2 Methods

2.2.1 Study area

Our investigation takes place on Kennicott Glacier in the Wrangell Mountains of Alaska
(61.50° N, −142.95° E; Figure 2.1). Kennicott is a 43 km long temperate, land-terminating
glacier, with ice thickness exceeding 500 m (M. Truffer, unpublished data; Armstrong et al.
[2016]). Kennicott spans 4500 m of elevation from 4996 m a.s.l. Mt Blackburn to the glacier’s
terminus at 490 m a.s.l. Kennicott Glacier has retreated by approximately 500 m from its
Little Ice Age (LIA) maximum extent [Rickman and Rosenkrans, 1997] and has lost 0.43
m w.e. a −1 on average over 2000-2013 [Larsen et al., 2015]. A growing proglacial lake lies
between the terminus and LIA moraine, but the glacier does not have an active calving front.
Approximately 19% (~ 70 km²) of Kennicott Glacier is debris-covered, with debris thickness
upwards of 1 m (L. Anderson, unpublished data).
The nearby town of McCarthy provides access to the glacier. Kennicott River, the glacier’s outlet stream, is spanned by a steel footbridge, which allows for continuous stage measurements. Kennicott Glacier acts as a dam to Hidden Creek (Figure 2.1), which forms an ice marginal lake that drains catastrophically between late June and early August and releases $\sim 25 \times 10^6$ m$^3$ of water beneath the glacier [Anderson et al., 2003]. This annual outburst flood has occurred since at least $\sim 1910$ when copper mining in the area began [Rickman and Rosenkrans, 1997]. The outburst flood provides a natural experiment to probe the effect of subglacial hydrology on basal sliding. The ice marginal Donoho Falls Lake (Figure 2.1), $\sim 8$ km downstream of Hidden Creek, acts as a manometer, proving a proxy for subglacial water pressure during the flood and in spring.

Previous research on the annual outburst flood of Hidden Creek Lake [Anderson et al., 2003; Walder et al., 2006] and its link to glacier motion [Anderson et al., 2005; Bartholomaus et al., 2008, 2011] have revealed that the Kennicott can serve as a natural experiment to probe sliding-hydrology connections. Bartholomaus et al. [2008, 2011] used GPS and on- and near-glacier hydrometeorologic equipment to show glacier motion increased during periods of increasing en- and subglacial water storage on daily, season, and outburst flood timescales. Our field campaigns expand on the work of Bartholomaus et al. [2011] by extending our monitoring timeseries to include multiple years, both earlier and later season observations, by maintaining a GPS monument over winter, and by deploying time lapse cameras on ice marginal lakes.

2.2.2 GPS installation and maintenance

We maintained on-glacier GPS monuments in 2006, 2012, 2013, and 2014 (Table 2.1). The 2006 data are published in Bartholomaus et al. [2008, 2011]. We anchored the GPS antennae to the glacier following the methods of Anderson et al. [2004] (Figure 2.2). We drilled 6 m holes into the glacier ice using a Heucke steam drill or Kovacs mechanical ice auger. Into these holes we insert two 3 m long, 2.5 cm diameter steel conduit which are joined using a slightly thinner 10 cm long coupler. On these poles we used hose clamps to secure a planar wooden triangle with holes drilled to accommodate the poles. This triangle has a bolt for
Figure 2.1: Map of field site and monitoring equipment. Symbols show equipment locations. Stars indicate on-glacier global position system (GPS) monuments. Circles and squares indicate timelapse cameras and pressure transducers, respectively. The triangle shows a National Park Service maintained automated weather station (AWS). Inset shows study location within the state of Alaska, indicated with an arrow. Glacierized area is shown in teal. Glacier outline is from the Randolph Glacier Inventory [Arendt et al., 2012]. Elevation data are extracted from the National Elevation Dataset and are shown with a 100 m contour interval. Stippled pattern represents surficial debris cover, mapped from a 2015 Landsat 8 image.
mounting the Trimble Zephyr Geodetic antenna. We subsequently drilled two 2 m holes to anchor 3 m lengths of 2.5 cm diameter PVC for supporting the solar panel and Pelican case containing the Trimble NetR9 GPS receiver and 18 Ah sealed lead acid battery. In summer 2014 we used Topcon GPS receivers instead of Trimble. The equipment is attached using a sliding ring and “floats”, moving down with the ice surface. The steel conduits supporting the antenna are frozen in or otherwise supported at depth and do not lower with ice surface ablation.

We installed 4 GPS monuments along the approximate glacier centerline every 2-3 km along a 10 km reach ranging from 750-1020 m a.s.l. At our most up-glacier and down-glacier GPS monuments, we measured air temperature inside a radiation shield and the rate of surface melt using an open-source data logger mounted on the steel antenna support poles [Wickert, 2014]. We periodically re-drilled the stations throughout the ablation season to ensure the antennae remain near level. GPS equipment was provided by UNAVCO.

We maintained 1 GPS monument over winter 2012-2013. Due to a lack of sun in winter, we powered the station using Cegasa air-alkaline batteries (model AS10-2). We connected ten 1.5 V 1200 Ah in series to create a long-lived 15 V power supply. These batteries provide a large amount of power for their size due to a manganese dioxide cathode that is continuously regenerated by reacting with atmospheric oxygen. This setup thus requires a “snorkel” built into the container housing the batteries to ensure adequate circulation of ambient air when the case is sealed and buried in snow.

<table>
<thead>
<tr>
<th>Year</th>
<th>Start DOY</th>
<th>End DOY</th>
</tr>
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<tbody>
<tr>
<td>2006</td>
<td>132</td>
<td>230</td>
</tr>
<tr>
<td>2012</td>
<td>131</td>
<td>207</td>
</tr>
<tr>
<td>2013</td>
<td>184</td>
<td>260</td>
</tr>
<tr>
<td>2014</td>
<td>130</td>
<td>227</td>
</tr>
</tbody>
</table>

Table 2.1: Observation periods for on-glacier GPS. Day of year is indicated by DOY where DOY 1 is January 1.
Figure 2.2: A typical on-glacier GPS monument, with instruments indicated. The look-down ultrasonic sensor is an ALogger [Wickert, 2014]. The receiver and battery, as well as a load controller, sit inside the Pelican case.
2.2.3 GPS processing for velocity timeseries

We use UNAVCO’s runpkr00 program to convert proprietary Trimble (.T02 files) and Topcon data formats into more easily processed formats. We concatenate .T02 files where multiple files exist for one day due to power loss or data download. We use UNAVCO’s teqc software to convert the runpkr00 into RINEX (Receiver INdependent EXchange format) format data. We then employ the TRACK module within GAMIT-GLOBK [Herring et al., 2006] for to solve for the location of the on-glacier GPS antennae using differential phased-based kinematic positioning. Our processing routine uses the ionosphere-free L1/L2 frequency combination and minimizes error due to tropospheric delay by processing relative to a stationary, off-glacier base station ∼ 10 – 20 km away. Our processing routine uses the GPT temperature-pressure atmosphere model and applies a Kalman filter. We exclude observations from satellites ≤ 15° above the horizon.

We identify and manually correct position change due to GPS station maintenance. Data with > 0.02 m reported horizontal position uncertainty are removed from the remainder of analysis. We manually remove additional low-quality data, assessed largely by abrupt, non-physical variation from “background” behavior. Position change during data gaps due to power loss is estimated by linear interpolation, which allows us to determine the average rate of position change over the gap. We subtract the first position from each station’s time series to define relative position change. Positions are converted from a geographic coordinate system to a “glaciologic” reference frame (i.e., relative to mean flow direction) as follows. We first define the mean flow azimuth (γ) by calculating

$$\gamma = \tan^{-1}\left(\frac{\Delta N}{\Delta E}\right)$$

(2.1)

where ΔN and ΔE are the changes in northing and easting, respectively, over the period of record. γ is the angle formed by the GPS mean flow direction and a line of longitude (i.e.,
it is relative to east, not north). We then apply the transformation matrix \( R \),

\[
R = \begin{bmatrix}
\cos \gamma & \sin \gamma \\
-\sin \gamma & \cos \gamma
\end{bmatrix}
\] (2.2)

where \( \vec{x} \) is the “glaciologic” coordinate pair, calculated as

\[
\vec{x}_i = \begin{bmatrix}
x_i \\
y_i
\end{bmatrix} = R \begin{bmatrix}
N_i \\
E_i
\end{bmatrix}
\] (2.3)

Here subscript \( i = [1, 2, \ldots, n - 1, n] \), where \( n \) is the number of observations, \( x \) and \( y \) are the down-glacier and cross-glacier coordinates, respectively, and \( N \) and \( E \) are the relative northing and easting, respectively. We convert the vertical coordinate to relative change but do not perform any further manipulation.

The resulting position data have high-frequency noise that render accurate velocity estimation impossible without further processing. We “up-sample” data (i.e., coarsen its temporal resolution) by interpolating the 30 second positions to 15 minute positions and smooth the data using a bisquare (quartic) kernel weighted mean. We use \( \pm 3.5 \) hours for the width of the bisquare kernel for smoothing horizontal positions, which we found minimizes noise while retaining real diurnal velocity fluctuations. We smooth the vertical coordinate using a \( \pm 12 \) hour kernel width, as the vertical coordinate has higher position uncertainty. Finally, we calculate the velocity by differencing the smoothed 15 minute positions.

2.2.4 Hydrometerology

We deploy Solinst LevelLogger (Model 3001) pressure transducers on ice-marginal lakes as well as the proglacial lake and Kennicott River. We convert pressure to water height using an assumed water density and neglecting barometric pressure variations. These assumptions introduce very minor error into our analysis. The transducers are accurate to 5% of the full range of the instrument’s measurement capability. Some pressure transducers measure up to
100 m water depth, while others are only rated to measure up to 30 m. The corresponding water height uncertainty of these sensors is therefore ±5 cm and ±1.5 cm, respectively. The sensors also record temperature, and the proglacial sensor measures electrical conductivity as well. We record water pressure, temperature, and electrical conductivity, where applicable, every 15 minutes.

In addition to pressure transducers, we deploy Moultrie timelapse cameras to monitor lake levels. These cameras capture an image every 1 hour and allow lake level reconstruction when the water level dropped below the pressure transducers. We calibrate time lapse imagery using the mid-summer outburst flood record, at which time we have concurrent pressure transducer data and images as the flood wave completely fills to the once-empty basin. We digitize the lake levels associated with ~ 10 m changes in the pressure transducer record to establish a calibrated reference image (Figure 2.3). We have pressure transducer data at this time because we moved the instrument to the lowest portion of the basin after the lake drained in early summer. We then manually identify daily lake stage maxima and minima from the timelapse imagery and estimate the associated lake stage using the reference photo (Figure 2.3). While digitizing, we withhold the spring pressure transducer record where there is overlap. Therefore, comparing these two datasets provides a measure of uncertainty in the digitizing process.

We construct a stage-volume relation for Donoho Falls lake by approximating the lake’s geometry with a right triangular pyramid (Figure 2.4 inset). Lake volume \( V \) can therefore be related to height \( h \) by,

\[
V = \frac{h^3}{3 \tan \theta \tan \Omega}
\]

where \( \theta \) is the average slope of the basin along the lake’s short planview axis, and \( \Omega \) is the average slope along the half-width in the lake’s long planview axis (Figure 2.4 inset). From field survey and analysis of satellite imagery, we find \( \theta = 0.32 \ rad = 18^\circ \) and \( \Omega = 0.27 \ rad = \)
Figure 2.3: Reference image used for digitization of timelapse photography. Red lines show apparent lake extent associated with the pressure transducer stages indicated in orange. Heavy orange line shows the near vertical ice wall used as a reference to establish lake stage. Thin orange lines show unchanging features used for image coregistration.
15°. The rate of change of volume as a function of lake height \( \frac{\partial V}{\partial h} \) is then calculated as,

\[
\frac{\partial V}{\partial h} = \frac{h^2}{\tan \theta \tan \Omega}
\]

We estimate snow and ice melt using a simple positive degree day (PDD) melt model [Hock, 2005] that is constrained by continuous on-glacier look-down ultrasonic sensors.

2.3 Results

2.3.1 Spring stage variations on Donoho Falls Lake

In each spring, we observe a chaotic sequence of fill and drain on the ice-marginal Donoho Falls Lake (DFL; Figures 2.1 and 2.5). In 2013, we record the initial filling of DFL on DOY 126, which closely coincides with the onset of > 0 °C air temperatures (hence PDD > 0; Figure 2.5). Spring 2013 was relatively late and cool and we observe relatively steady 31 m DFL stage increase over DOYs 126-134, corresponding to an average filling rate of 3.7 m d\(^{-1}\). After this point, the filling rate slows to 1.1 m d\(^{-1}\) over DOYs 134-141. Diurnal stage fluctuations then appear and the average filling rate further slows to 0.50 m d\(^{-1}\). Coincident with a sudden increase in air temperature (DOY \sim 145; Figure 2.5), DFL stage drops by 14.8 m over 2.3 d. This drop is followed by a standstill in mean stage with 4.5 m diurnal stage
fluctuations (peak-trough amplitude). This time is followed by a 25 m stage drop with 10 m amplitude diurnal fluctuations. We then observe a 31.5 m stage rise over 2.5 d, coincident with another period of high air temperature (DOY $\sim 157$; Figure 2.5). After rapid lake drainage initiating late on DOY 161, we observe two short-lived refilling events before the basin drains for the remainder of the spring.

The time of peak stage evolves throughout the course of spring (Figure 2.6). The stage minima are roughly steady at 09:00-10:00 local time. The maxima, however, occur around 02:00-03:00 with the onset of diurnals. The maxima then occur progressively earlier ($\sim 20:00-21:00$) as stage decreases. With the large stage increase on DOY 157 the maxima move later, to $\sim 24:00$ (Figure 2.6).

The 2012 and 2014 DFL stage time series exhibit similar dynamics to those described...
above (Figure 2.5). In 2014, the final lake drainage occurred earlier, on DOY 152, which may be related to that year’s warm spring temperatures.

2.3.2 GPS-derived velocity time series

We develop high quality glacier surface velocity time series for the 2012-2014 ablation seasons. Due to power loss, gaps exist within each time series and truncate the record in places (Figure 2.7). Thus, we do not possess a continuous melt-season record. However, from piecing together records across years, we are able to characterize three timescales over which glacier surface velocity varies: 1) high amplitude diurnal fluctuations of varying amplitude but consistent presence across all time periods; 2) short-lived many-fold speedups related to the annual Hidden Creek Lake outburst flood as well as a fall rain event, and ;3) spring
speedup events initiating around DOY 130-140 and lasting until DOY 150-170 (Figure 2.7). We have substantial temporal overlap in early-mid season (∼ DOY 130-190) in 2012 and 2014 (Figure 2.7). The first portion (DOY 130-165) of this period coincides with the stage variations on DFL, and allow us to investigate the link between the glacier hydrologic system and velocity change during the spring speedup events of these two years.

In 2012, the spring speedup at the down-glacier GPS2 and GPS3 stations definitively begins late on DOY 141, evidenced by the onset of large amplitude diurnal velocity fluctuations and an increase in mean velocity, though there is some suggestion of a slow speedup beginning DOY 137 (Figure 2.8b). The possible onset of speedup on DOY 137 closely coincides with the initiation of > 0 °C air temperatures (Figure 2.5). This down-glacier speedup is
accompanied by acceleration at the up-glacier GPS4 and GPS5 stations, with much smaller or negligible diurnal velocity fluctuations (Figure 2.8b). The onset of high amplitude diurnal velocity fluctuations does not begin at GPS5 until DOY 158, lagging the down-glacier stations by 11 days. In 2014, we appear to miss the initiation of the spring speedup at the down-glacier GPS3, potentially due to a warm spring (Figure 2.5), though we see a multi-day speedup at the up-glacier GPS5 (Figure 2.9b). We find large amplitude diurnal velocity fluctuations at GPS3 throughout the period of record and more muted, but present, diurnal cycles at GPS5 (Figure 2.8b).

The speedup is accompanied by a two-phased ice uplift at all stations, with uplift amplitude decreasing moving up-glacier (Figure 2.8d). The onset of uplift appears to propagate as a wave, with GPS5 lagging GPS2 by 5.5 days, which corresponds to a wave speed of 1.87 km d$^{-1}$. GPS2 is uplifted 0.32 m over 6.9 days, an uplift rate of 0.046 m d$^{-1}$. The uplift rate then slows to 0.032 m d$^{-1}$ over DOYs 148.3-162.7.

The most significant difference between the spring GPS time series in 2012 and 2014 is the vertical uplift history (Figures 2.8d and 2.8d). The 2012 record is characterized by monotonic uplift at all stations. In 2014, however, we observe lower-amplitude (0.4 m) uplift at GPS4 and GPS5 that occurs quickly (8 cm d$^{-1}$ sustained for 2.3 d, and then relatively stable for 10 d, followed by decay to zero in a quasi-exponential fashion (Figure 2.8d). GPS3 records similar behavior, although lower in amplitude and lagged by several days.

Longitudinal strain rates ($\dot{\epsilon}_{xx}$) are generally compressive through the reach, with median $\dot{\epsilon}_{xx}$ ranging from $-2.41 \times 10^{-5}$ d$^{-1}$ to $-2.68 \times 10^{-5}$ d$^{-1}$ depending on chosen stations (Figures 2.8c and 2.9c). We observe significant variation about the median value over diurnal and multiday timescales due to differing phasing of velocity between stations. The onset of the spring speedup at the down-glacier stations causes a reduction in the mean magnitude of compressive strain rate to $\dot{\epsilon}_{xx} \approx -1.00 \times 10^{-5}$ d$^{-1}$. This period is accompanied by short-lived excursions to extensional strain rates ($\dot{\epsilon}_{xx} = 2.2 \times 10^{-5}$ d$^{-1}$). The onset of the spring speedup at GPS5 brings $\dot{\epsilon}_{xx} \approx -1.00 \times 10^{-5}$ d$^{-1}$ back to a value similar to its pre-speedup state. The increasing synchronicity between station velocity change later in the melt season
reduces the magnitude of diurnal variations in $\dot{\epsilon}_{xx}$ (Figure 2.8c).

2.3.3 Evolving relationship between station velocities

We analyze time lags between GPS stations to determine the spatial and temporal coherence of glacier velocity fluctuations throughout our study reach. We iteratively calculate inter-station correlations using varied time lags (Figure 2.10) to determine the best-fitting inter-station lag time between peak velocities. We calculate correlation using 8 days of record and lagging velocity data in increments of 15 minutes with maximum/minimum lags of $\pm 12$ hours. The time lag with the highest correlation over this 8 day span is chosen as the best fitting lag time (Figure 2.10). We repeat this analysis over our observation period
Figure 2.9: Time series of 2014 hydrology and glacier motion. a) Donoho Lake stage (dark blue for pressure transducer, light blue with circles for digitized photos) and air temperature (red), b) on-glacier GPS longitudinal velocity, c) longitudinal strain rate between GPS3 and GPS5, and d) relative elevation of the GPS antenna. The heavy black line shows the uplift expected from vertical strain (Section 2.4.1). Vertical dashed line indicates timing of final drain of Donoho Falls Lake.
to produce a record of inter-station lag time of peak velocities over the course of the melt season (Figure 2.11). We find consistent relationships between stations, with closer regions of ice exhibiting shorter lags throughout the melt season (Figures 2.10 and 2.11). In the early season (DOY 146), GPS5 lags the down-glacier stations by 0.75, 5, and 7 hours (Figures 2.10b). The respective stations are 3.94, 7.78, and 10.27 km down-glacier from GPS5. Later in the summer, we find the timing of peak velocity becomes more similar (Figure 2.11); in mid-summer (DOY 175), GPS5 lags by only 0.15, 1.00, and 1.25 hours, respectively. In addition to timing, we find the maximum correlation value increases through the course of the melt season (Figures 2.10b,d), indicating the stations are behaving more similarly.

Evolving links between GPS stations is apparent when the velocity at one station is plotted against another (Figure 2.12). At the down-glacier stations GPS2 and GPS3, which are separated by 2.49 km, we find high velocity covariance throughout the spring period of record (DOYs 130-165; Figure 2.12a). Plotting along the 1:1 line on Figure 2.12 indicates stations recording similar velocity time series. Figure 2.12a exhibits counter-clockwise hysteresis, which indicates GPS2 velocity increases precede those at the further up-glacier GPS3. The width of the hysteresis loops indicates the magnitude of temporal lag between stations; thus, GPS2 and GPS3 experience peak diurnal velocity timing throughout the record, indicated by the narrow hysteresis loops. Comparing GPS3 and GPS5, however, (Figure 2.12b) we find an evolving relationship. Both stations experience similar multi-day speedup over DOYs 130-145, shown by early-season drift along the 1:1 line. Around DOY 145 high amplitude diurnal velocity fluctuations initiate at down-glacier GPS3 with no response at GPS5, indicated by horizontal lines on Figure 2.12b. After this point the spring speedup passes at GPS3, indicated by leftward drift, and around DOY 155 high amplitude diurnals occur at both stations. The relatively wide counter-clockwise hysteresis loops at this time indicate that the timing of peak velocity at GPS5 lags between the down-glacier GPS3 (≈ 2 hours as shown on Figure 2.11).
Figure 2.10: Illustration of inter-station velocity lag timing relative to the up-glacier GPS5. a) and c) show velocity timeseries centered on DOY 146 and DOY 175, respectively. b) and d) show correlation coefficients when data are iteratively lagged by varied amounts. Correlation maxima indicate the time GPS5 lags behind these stations. Positive lags indicate that GPS5 peaks after the down-glacier station. Locations shown on Figure 2.1.
Figure 2.11: Time series of best lag time at GPS5. Positive values indicate the down-glacier station velocity increase preceding that at GPS5. Locations shown on Figure 2.1.

Figure 2.12: Evolving link between glacier velocity at different points. a) Down-glacier stations GPS2 and GPS3. b) Down-glacier GPS3 and up-glacier GPS5. Points are colored by time of year. Black dashed line shows 1:1. Locations shown on Figure 2.1.
2.3.4 Link between Donoho Falls Lake stage and glacier velocity

Having characterized the DFL stage history (Section 2.3.1), the glacier horizontal displacement and vertical uplift time series (Section 2.3.2), and evolving links between glacier velocity at different points (Section 2.3.3), we now probe the relationship between glacier velocity and DFL stage during spring.

In both 2012 and 2014, we observe tight links between DFL stage and the speeds of the down-glacier stations GPS2 and GPS3 (Figures 2.8 and 2.9), which are located 2.0 and 3.4 km on a straight line from DFL, respectively (Figure 2.1). GPS2 and GPS3 are 0.91 and 3.50 km up-glacier from a transverse line drawn across the glacier at the location of DFL. We find the 2012 spring speedup event began around DOY 138 at GPS2 and GPS3, coincident with a 10 m amplitude DFL stage increase over the course of 4 days. The up-glacier GPS4 and GPS5 speedup began late on DOY 141, coincident with peak DFL stage (Figure 2.8a-b). In 2014 we missed the initiation of the spring speedup, which began early due to warm air temperatures (Figures 2.6 and 2.9), although the general link between DFL stage and glacier velocity appear similar.

We find high amplitude diurnal velocity fluctuations on the down-glacier GPS3 when DFL stage is high (> 35 m) and steady or rising (Figure 2.13a). During these times, we
find clockwise hysteresis: timing of peak velocity precedes the timing of peak stage. The mean velocity and amplitude of diurnal fluctuations decreased during times of low DFL stage and especially falling stage (Figures 2.8a-b and 2.13). At the up-glacier GPS5 (11.18 km up-glacier), the correlation is low between glacier velocity and DFL stage until DOY $\sim$160 when GPS5 diurnal velocity fluctuations begin, coinciding with high DFL stage (Figure 2.13b).

Velocity extrema (i.e., maxima and minima) generally precede stage extrema by $\sim$ 3 hours (Figure 2.14). Further, velocity extrema shift earlier in the day over the study period, retaining the same lag time with DFL stage as its timing evolves. During the large-scale DFL re-filling of DOYs 158-163, correspondence of the timing of extrema in stage and velocity is lost as stage extrema occur later in the day and the timing of velocity extrema remain unchanged (shown by the deviation from the stage fit lines around DOY 160 on Figure 2.14).

2.4 Discussion

2.4.1 Interpreting spring uplift

The vertical motion of a point that is attached to the subsurface and hence does not lower with ice surface ablation, as is the case with our GPS antenna, reflects the sum of elevation change due to vertical strain, horizontal motion down a sloped surface, and changing volume of subglacial cavities that cause changing ice-bed separation [e.g., Anderson et al., 2004]. Mathematically, this may be stated,

$$w(t) = w_{\text{strain}} + w_{\text{slope}} + w_{\text{sep}} + w_{\text{tilt}}$$

(2.6)

where $w(t)$ is the time-variable GPS antenna vertical velocity (taken to be positive upwards), $w_{\text{strain}}$ is the vertical velocity due to vertical strain, $w_{\text{slope}}$ is the vertical velocity due to ice surface parallel motion, $w_{\text{sep}}$ is the vertical velocity due to changing bed separation, and $w_{\text{tilt}}$ is the velocity associated with GPS antenna tilt during melt out.

The vertical velocity arising from longitudinal compression can be estimated from

$$w_{\text{strain}} = -H \frac{\partial u}{\partial x},$$

(2.7)
Figure 2.14: Evolution of maximum and minimum stage and velocity in spring 2012.
where $H$ is the ice thickness, and $\frac{\partial u}{\partial x}$ is the longitudinal strain rate. This equation is approximately true when cross-glacier velocity is near zero and the longitudinal strain rate at the surface closely reflects the column-averaged longitudinal strain rate. The ice thickness in the location of GPS2 and GPS3 is likely $\sim 400 - 600$ m [Armstrong et al., 2016]. Using $H = 400$ m and $\frac{\partial u}{\partial x} = -0.75 \times 10^{-5}$ d$^{-1}$ to provide a lower bound (Figure 2.8c) for vertical strain associated with longitudinal compression, we find $w_{\text{strain}} = 0.3$ cm d$^{-1}$. Using $H = 600$ m and $\frac{\partial u}{\partial x} = -3.1 \times 10^{-5}$ d$^{-1}$ to provide an upper bound (Figure 2.8c), we find $w_{\text{strain}} = 1.86$ cm d$^{-1}$. Importantly, this uplift should not be reversible if the strain regime remains relatively steady. As the ice surface slope through this reach is $\sim 3\%$, surface-parallel motion results in surface lowering of $7.5 - 15$ cm over a 10 day period ($w_{\text{slope}} = -0.75$ to $-1.5$ cm d$^{-1}$) for ice moving $25 - 50$ cm d$^{-1}$. Elevation change due to antenna tilt is a small, but non-zero contributor to the vertical time series. Due to frequent re-drilling, the GPS antenna never tilted more than $20^\circ$ from vertical. If an antenna were 4 m above the ice surface, as was sometimes the case at the end of the season, and tilted at $20^\circ$ off vertical, the station would lower $\sim 14$ cm over the period of tilt. Using 20 d as a conservative estimate of how quickly this could occur, this would result in $w_{\text{tilt}} = -0.68$ cm d$^{-1}$. Taking values to minimize our estimated bed separation ($w(t) = 4.6$ cm d$^{-1}$, $w_{\text{strain}} = 1.9$ cm d$^{-1}$, $w_{\text{slope}} = -0.75$ cm d$^{-1}$, and $w_{\text{tilt}} = 0.0$ cm d$^{-1}$), we find $w_{\text{sep}} = 3.5$ cm d$^{-1}$, 76% of $w(t)$. Using more typical and defensible values ($w(t) = 6.0$ cm d$^{-1}$, $w_{\text{strain}} = 1.3$ cm d$^{-1}$, $w_{\text{slope}} = -1.0$ cm d$^{-1}$, and $w_{\text{tilt}} = 0.0$ cm d$^{-1}$), we find $w_{\text{sep}} = 5.9$ cm d$^{-1}$, 98% of $w(t)$. Therefore, we conclude the early season uplift (DOYs 130-165 in 2012; Figure 2.8d) is largely accomplished by increasing bed separation due to high rates of basal motion associated with high pressure in subglacial cavities. Even the slower, quasi-steady uplift rates following DOY 152 are too large to be produced by vertical strain at the highest observed $\dot{\epsilon}_{xx}$ we observe unless $H > 1000$ m, which is unreasonably large for this reach.

Vertical uplift during the spring event differs significantly between 2012 and 2014 (Figures 2.8d and 2.9d). The rapid uplift rate of GPS4 and GPS5 over DOYs 135-140 in 2014 can only be explained by increased bed separation. This explanation is supported by the
subsequent elevation loss (Figure 2.9d; DOYs 148-160) which likely reflects cavity collapse.

We now ask what causes the observed difference in uplift behavior between 2012 and 2014? Recall that spring 2012 was relatively cold, while in 2014 spring was warm and early. Consistently > 0° temperatures began around DOY 121 in 2014, while they began DOY 142 in 2012 (Figure 2.5). In addition, the final draining of DFL occurs on DOY 152 in 2014, while it occurred 11 days later in 2012 (Figure 2.5). We posit the slower input of meltwater to the subglacial drainage system in 2012 resulted in slower development of efficient subglacial drainage that resulted in long-lived bed separation (Figure 2.8d), higher mean velocity, and larger amplitude diurnal velocity fluctuations (Figure 2.8b). In contrast, high early season melt inputs in 2014 resulted in early maturation of subglacial drainage, which caused the collapse of distributed subglacial cavities (Figure 2.9d).

2.4.2 Early season link between water pressure and velocity

Generally, we find high amplitude diurnal velocity fluctuations at high DFL stage (Figure 2.13). This agrees well with theory and previous observations showing high sensitivity of the rate of basal motion to water pressure at pressures approaching that of the ice overburden [Iken and Bindschadler, 1986]. We find velocity maxima generally precede stage maxima (Figure 2.14) and better coincide with the maximum rate of change of subglacial water pressure, in accord with previous studies [Bartholomaus et al., 2008]. The tight link between subglacial water pressure and glacier velocity we observe here is more straight-forward than that observed in some borehole studies [e.g., Harper et al., 2007] and likely reflects that DFL records water pressure variations integrated over an area large enough to be “relevant” for glacier force balance.

We posit Donoho Falls lake acts as a manometer, filling with “backed up” subglacial water during times of high subglacial water pressure. There are several reasons for believing this is how DFL functions: 1) the water is very turbid relative to the lake-supplied stream feeding into the basin, suggesting a subglacial origin; 2) evolving timing of stage extrema are not readily explained by non-subglacial inputs, and; 3) the > 30 m fill-and-drain sequences are likely too large to be explained by variable inputs from the small Donoho Falls Creek and
melt inputs from the small supraglacial catchment but instead suggest intimate connection with a dynamic subglacial hydrologic system. To estimate the role of non-subglacial water on the DFL stage history, we approximate the volume of water inputs from other sources. We conservatively estimate the Donoho Falls Creek discharge using a step function that inputs 2 m$^3$ s$^{-1}$ for 12 h and 0 m$^3$ s$^{-1}$ for the remainder of the day. This inputs $86 \times 10^3$ m$^3$ over the course of the day. Using a high resolution digital elevation model and satellite imagery, we find the supraglacial catchment contributing to DFL is $960 \times 10^3$ m$^2$. On a warm mid-summer day, the ice surface melts 4 cm d$^{-1}$. Applying this melt rate over the supraglacial catchment area, we find supraglacial melt may contribute $38 \times 10^3$ m$^3$ via direct runoff. Using an idealized DFL geometry (Figure 2.4b), these conservative estimates of non-subglacial water inputs could explain several m stage swings at peak stage if there is no efflux of water into the subglacial drainage system. Starting from an empty basin, the same volume could fill the basin with 30 m of water (Figure 2.4a) if there is no outflow. That being said, we believe these stage variations are largely driven by subglacial drainage dynamics because: 1) the input from Donoho Falls Creek is likely poorly approximated by a half-day-long step function because it is lake-fed and not snow-melt fed, likely resulting in an overestimate of stream inputs; 2) 4 cm d$^{-1}$ is a high melt rate for this time of year, resulting in an overestimate of local supraglacial meltwater inputs; 3) > 10 m stage swings at water levels higher than an empty basin cannot be explained by conservative estimates of non-subglacial water inputs, and; 4) turbid water and evolving timing of stage extrema are most clearly explained by connection with subglacial drainage. For the remainder of the discussion, we assume DFL is continuously well-connected to the subglacial drainage system and reflects regionally averaged subglacial water pressure within the glacier, as has been found to be true for water pressure measured in moulins [Andrews et al., 2014].

We interpret the seasonal evolution towards earlier DFL peak stage (Figures 2.6b and 2.14) to reflect both: 1) shorter lag time between surface melt production and transmission to the subglacial hydrologic system due to declining snow cover, and; 2) greater channelized extent of subglacial drainage that is capable of quickly evacuating meltwater inputs. During late
spring high stands (beginning DOY ∼ 158 in both 2012 (Figure 2.14) and 2014 (Figure 2.6)) we find a reversal in the long-term trend towards earlier peak stage. We interpret this change in timing to reflect temporarily overwhelmed channelized subglacial drainage due to large meltwater inputs following several cool days (Figure 2.6). This ∼ 50 m late-spring stage increase is accompanied by increases in mean velocity, amplitude of diurnal velocity fluctuations, and ice uplift (Figure 2.9).

2.4.3 Evolving velocity links along a centerline profile

Longitudinal strain rates are highly variable prior to the onset of vertical uplift at GPS5 (DOY 148; Figure 2.8d), although consistently more extensional than before the down-glacier spring speedup or those following GPS5 uplift. We suggest GPS5 speedup over DOYs 142-148 (Figure 2.8b) may be due to loss of longitudinal resistive stress from the down-glacier ice that accelerates in response to local loss of basal traction due to widespread high pressure subglacial water storage. This is the opposite of the “ice dam” effect observed by Howat et al. [2008] but similar to that modeled by Price et al. [2008] who found variations in basal conditions in low elevation ice can cause non-locally forced speedup of up-glacier ice.

We interpret GPS5 uplift to reflect widespread high pressure subglacial water storage (Section 2.4.1). Decreased lag of the timing of peak velocity (from ∼ 6 h to ∼ 2 – 3 h; Figure 2.11) and onset of GPS5 diurnal velocity fluctuations (Figure 2.8b) that are of similar magnitude to those at down-glacier stations (Figure 2.12b) follow uplift at GPS5. We interpret the evolution of GPS5 behavior to reflect its transition from response to down-glacier changes in the rate of basal motion that are transmitted up-glacier through stress gradient coupling to response to locally-forced variations in the rate of basal motion. Importantly, this time evolution of inter-station velocity links to discern between locally- versus “globally”-forced velocity change. At the down-glacier stations (Figure 2.12a) we see high correlation of ice surface velocity throughout the period of record; this time invariance obscures physical controls on the observed behavior.

Interestingly, the evolution towards locally-forced variations in ice surface speed result in more homogeneous glacier behavior over our 10.3 km study reach (Figure 2.10c-d), consistent
with remote sensing observations documenting the uniformity of summer ice surface speedup [Armstrong et al., 2016; Armstrong et al.]. Although the surface speedup is not a faithful representation of basal motion [Balise and Raymond, 1985; Gudmundsson, 2003; Raymond and Gudmundsson, 2005; Armstrong et al., 2016], these observations of changing velocity links suggest spatially uniform locally-forced basal motion at least partly drive the uniformity of the surface speedup.

This work contrasts with that of Ryser et al. [2014] and Flowers et al. [2016] who found a time-invariant spatial distribution of “slippery” and “sticky” spots, with small regions driving variations in ice motion over a large area. We find evidence for growth of “slippery” spots, with a transition from globally-forced to locally-forced variation in ice surface speed.

2.5 Conclusions

We present and analyze velocity data from on-glacier GPS monuments as well as subglacial water pressure variations, as recorded by a well-connected ice marginal lake. On the ice-marginal Donoho Falls Lake, in spring (DOYs 130-165) 2012-2014, we find a chaotic fill-and-drain sequence with > 50 m stage swings over 2-3 day periods, with superimposed 2-25 m diurnal stage fluctuations. We find the timing of stage extrema (i.e., maxima and minima) shift earlier in the day over our period of record and consistently lag velocity extrema by several hours. We find times of high stage correspond to vertical uplift, higher mean velocity, and increased amplitude of diurnal velocity fluctuations on nearby on-glacier GPS monuments. We therefore interpret the chaotic evolution of subglacial water pressure recorded by Donoho Falls Lake to document the halting establishment of efficient subglacial drainage within the glacier.

We find evolving links of glacier velocity along our 10.3 km centerline transect during the evolution of subglacial drainage. In the early portion of our study record, we find down-glacier stations first speed up, initiate diurnal velocity fluctuations, and experience ice uplift. We interpret these observations to reflect overwhelming distributed subglacial with the onset of spring meltwater inputs. In the early season, the up-glacier station slowly speeds up, lacks pronounced diurnal velocity fluctuations, the speedup is not accompanied
by surface uplift, and the timing of velocity extrema lag those at the down-glacier by 5-7 hours. We interpret this to reflect “global” control on the early season speedup at this up-glacier site, which accelerates due to longitudinal stress-gradient coupling to the “slippery” bedded ice downstream.

Following the initiation of ice uplift at the up-glacier station, we find the onset of diurnal velocity fluctuations, presumably reflecting the local development of high pressure distributed subglacial drainage and the ice surface velocity responding sensitively to “local” basal conditions. Across our 10 km study reach, the onset of diurnal velocity peaks is accompanied by a reduced temporal lag between up- and down-glacier velocity extrema to 2-3 hours, by similar amplitude diurnal velocity fluctuations, and by a relatively steady strain rates.

Thus, we find the glacier ice surface velocity sensitively responds to the time evolution of the water pressure field. We find evidence for a seasonal shift from “global” to “local” control on velocity variability, and that the ice acts very uniformly during times of widespread basal motion.

2.6 Acknowledgements

We gratefully acknowledge Katy Barnhart, who spent many weeks on Kennicott Glacier deploying and maintaining GPS instrumentation. We thank Henry Berglund, who assisted in GAMIT installation and GPS processing. In addition, we thank Kristine Larson for providing GPS-related advice, and Ethan Welty for providing a smoothing routine used in data processing.
Chapter 3

Modeling the WorldView-derived seasonal velocity evolution of Kennicott Glacier, Alaska

Abstract

Glacier basal motion generates diurnal to multi-annual fluctuations in glacier velocity and mass flux. Understanding these fluctuations is important for prediction of future sea level rise and for gaining insight into glacier physics and erosion. Here, we derive glacier velocity through cross-correlation of WorldView satellite imagery to document the evolution of ice surface velocity on Kennicott Glacier, Alaska, over the 2013 melt season. The summer speedup is spatially uniform over a \( \sim 12 \) km\(^2\) area, over which the spring velocity varies significantly. Velocity increases by 1.4 to 10-fold across the study domain, with larger values where spring velocities are low. To investigate the cross-glacier distribution of basal motion required to explain the observed surface speedup, we employ a two-dimensional cross-sectional glacier flow model. We find the model is insensitive to the spatial distribution of basal slip because stress gradient ice coupling diffuses the surface expression of the basal velocity field. While the temporal evolution of the subglacial hydrologic system is critical for predicting a glacier’s response to meltwater inputs, our work suggests that glacier and ice sheet models do not require a detailed representation of subglacial hydrology to accurately capture the spatial pattern of glacier speedup.

This chapter has been published:
3.1 Introduction

3.1.1 Background

Variations in glacier basal motion induce ice surface speedups on diurnal [e.g., *Iken and Bindschadler*, 1986], multi-day [e.g., *Das et al.*, 2008; *Bartholomaus et al.*, 2008], seasonal [e.g., *Mair et al.*, 2003], and multi-annual [e.g., *Kamb et al.*, 1985] timescales. The realization that velocity can fluctuate on seasonal timescales in some regions of the Greenland ice sheet (GrIS; *Zwally et al.* [2002]) suggested that changes in basal motion could allow the ice sheets can respond more quickly to climate warming than previously expected [*Parizek and Alley*, 2004]. In addition, basal sliding is required for a glacier to erode its bed [e.g., *Hallet*, 1979; *Iverson*, 1991], making understanding of this process important for modeling the evolution of both polar and alpine landscapes over geologic timescales [e.g., *Harbor*, 1992; *MacGregor et al.*, 2000; *Anderson et al.*, 2006; *Herman et al.*, 2011; *Beaud et al.*, 2014].

Variations in glacier surface velocity have been observed for decades using optical surveying of displacement stakes [e.g., *Iken and Bindschadler*, 1986] and, more recently, through on-glacier GPS monuments [e.g., *Anderson et al.*, 2004; *Flowers et al.*, 2011; *Tedstone et al.*, 2013; *Kehrl et al.*, 2015]. The logistics and expense associated with installing, maintaining, and resurveying monuments has meant on-glacier GPS and displacement stakes are typically sparse. Therefore, previous studies on individual glaciers have often focused more on temporal variation than spatial variations in glacier motion. Where spatial analyses are presented, they are often along a glacier flowline and resolve only along-flow velocity patterns [*Anderson et al.*, 2004; *Bartholomaus et al.*, 2008; *Sole et al.*, 2011]. However, several studies do investigate spatial patterns in basal motion in both planview dimensions [*Harbor et al.*, 1997; *Mair et al.*, 2001; *Nienow et al.*, 2005; *Riesen et al.*, 2010]. These studies have highlighted the limitations of attributing basal sliding to solely local variables such as subglacial water pressure at a point, local basal traction, or the locally defined driving stress.

Satellite remote sensing has facilitated the measurement of spatially distributed glacier velocity fields using optical image cross-correlation [*Scambos et al.*, 1992], but studies have
been limited by relatively coarse spatial and temporal image resolution [e.g., Scherler et al., 2008; Herman et al., 2011; Heid and Käab, 2012]. Satellite radar interferometry [Goldstein et al., 1993] provides another powerful tool for obtaining spatially distributed glacier velocity measurements, but it is better suited for the low relief and relatively stable surface elevations that are characteristic of ice sheets [Joughin et al., 2010a]. Recently, high resolution (0.5-2.5 m pixel) imagery acquired by the WorldView and SPOT satellites has enabled analysis of short timescale (2-4 weeks) glacier dynamics [e.g., Berthier et al., 2005].

To understand the spatiotemporal distribution of basal motion, we combine repeat WorldView-derived glacier velocity fields and a 2-D cross-sectional glacier flow model. We first document the spring and summer velocity fields using satellite imagery, and difference them to obtain a summer surface speedup pattern. We then employ the flow model to infer the distribution of basal motion required to match the observed speedup. In addition, we undertake a numerical experiment to determine how widely two regions of high basal velocity must be separated before they are resolved as separate features in the ice surface speedup by the flow model.

3.1.2 Study location

We investigate the Kennicott Glacier in the Wrangell Mountains of Alaska (Figure 3.1). The Kennicott Glacier trunk is approximately 40 km long by 3.5 km wide and ranges from 450 to 5000 m a.s.l. The Kennicott Glacier and its tributaries cover 375 km$^2$ of a 670 km$^2$ basin [Bartholomaus et al., 2011; Arendt et al., 2012]. Kennicott Glacier has retreated approximately half a kilometer since its Little Ice Age maximum extent [Rickman and Rosenkrans, 1997] and had a spatially averaged mass balance of -0.44 m w.e. a$^{-1}$ over 2000-2013 [Larsen et al., 2015]. Kennicott Glacier acts as a dam to Hidden Creek (Figure 3.1), which forms an ice marginal lake that drains catastrophically between late June and early August and releases $\sim 25 \times 10^6$ m$^3$ of water beneath the glacier [Rickman and Rosenkrans, 1997; Anderson et al., 2003].
Figure 3.1: Satellite image of Kennicott Glacier and its tributaries. The red box shows the approximate footprint of WorldView imagery used for velocity estimates. Yellow polygons show off-glacier locations used to quantify image mis-registration and uncertainty in velocity estimates. Donoho (Fireweed) is the northern (western) polygon. Location of GPS referenced in text and Figure 3.3 noted. Inset shows Alaska with box indicating approximate extent of main figure.
3.2 Methods

3.2.1 Image Acquisition and Preparation

We obtained \( \sim 0.5 \) m pixel WorldView1 and WorldView2 optical satellite imagery from the Polar Geospatial Center at the University of Minnesota. The images cover \( \sim 50 \) km\(^2\) of the terminal reach of Kennicott Glacier (Figure 3.1). We utilized 6 high-quality stereopair images acquired over spring-fall 2013 to generate a digital elevation model (DEM) and to estimate velocity fields from automated pixel tracking (Table 3.1). The shortest (longest) time between successive images is \( \sim 2 \) (5) weeks. Shorter period velocity fluctuations are averaged over these multi-week periods. The relatively high off-nadir angles (Table 3.1) of the image acquisitions mean that our imagery is sensitive to both vertical offsets and parallax (S. Leprince, personal communication, 2014). Also, as the glacier surface is rapidly evolving, a DEM created at one time may not accurately describe the glacier surface at another time. If unaccounted for, these issues would produce lower quality orthoimagery and subsequent image correlation. For these reasons, we employed the Leica Photogrammetry Suite (LPS) in ERDAS IMAGINE to orthorectify imagery before image correlation. For each time slice, we generate a 1-m pixel DEM from the stereopairs and orthorectified each image using the DEM from that timeslice. We coarsened the WorldView orthoimagery to 0.7 m pixels, which allowed us to maintain high spatial resolution while establishing a uniform pixel size for all images, which vary slightly from scene to scene.

<table>
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<th>Days elapsed</th>
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<td>2020011E772C8400</td>
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<td>23.0</td>
<td>26</td>
</tr>
</tbody>
</table>

Table 3.1: Information of WorldView imagery used. Days elapsed indicates time between successive images used for velocity correlations. N/A where image correlations were not performed. Glacier velocity is effectively averaged over these time periods.
3.2.2 Image Correlation and Error Analysis

We calculated glacier velocity using COSI-Corr, a free pixel-tracking program for use with the ENVI remote sensing software [Leprince et al., 2007]. The resulting velocity maps (Figure 3.4) have a spatial resolution of 5.6 m. We estimated image correlation error from analysis of calculated displacements over two flat-lying, static, off-glacier regions (Figures 3.1 and 3.2; Fireweed, to the west of the glacier, and Donoho, to the north). Any apparent motion at these locations is due to image mis-registration and distortion produced in the orthorectification process. We found that the apparent displacements in these regions were approximately normally distributed, and fit the data with a Gaussian curve (Figure 3.2). The spread of these Gaussian distributions, as measured by their standard deviations ($\sigma$), provides an estimate of random error in our image correlations. Standard deviations range from 0.19 to 0.34 m. We used the range of -1$\sigma$ of the Gaussian fit with a lower mean (e.g., ‘Fireweed’ in Figure 3.2a) to +1$\sigma$ of the Gaussian fit with a higher mean (e.g., ‘Donoho’ in Figure 3.2a) to conservatively estimate uncertainty in displacement magnitude. This corresponds to ± 0.5 m uncertainty in displacement magnitude (Figure 3.2). As the time elapsed between two successive images ranges from 17 to 38 days, our uncertainties correspond to ± 0.01 to 0.03 m d$^{-1}$ in estimated velocity. The ± 0.5 m uncertainty estimate addresses random error (as well as image distortion), but does not address systematic error (also known as measurement bias). Our georeferencing was not perfect, and as a result, we found mis-registration from pure translation of 2.38 to 3.25 m between successive images. We shifted the “slave” image such that the means of the Gaussians are centered about zero (Figure 3.2). After these corrections, high relief areas still have large apparent displacements due to shadowing effects and error in orthorectification, but low relief areas such as the glacier itself and the surrounding valley had negligible apparent displacements (bias). We did not correct for North-South striping in speedup maps, which originate from satellite sensor distortion (S. Leprince, personal communication, 2014). We extracted the apparent velocity across a transect perpendicular to striping on off-glacier terrain and estimated the error associated with striping to be ± 0.015 m d$^{-1}$ at worst. Our estimate of random error...
above incorporates error due to striping.

3.2.3 Extraction and processing of ice surface elevation and velocity

We extracted longitudinal and cross-glacier transects of glacier surface velocity and ice surface elevation using a Python-implemented swath profiler. We calculated mean ice velocity and elevation along the transect using a template rectangle measuring 10 m by 200 m in the along-transect and cross-transect directions, respectively. We averaged over 200 m (∼40 pixels) in the cross-transect direction to minimize the effect of spurious data due to random error and regions of low-quality velocity estimates. We sampled along-transect every 10 m
we only present mean attribute values, which reduces the effect of random noise. We extracted profiles across one down-glacier transect and five cross-glacier transects, although in this paper we only present the results from two representative cross-glacier transects (Figure 3.4). Despite the averaging applied in the swath profiler, there remain outliers in ice surface speed due to poor orthorectification, co-registration, and debris cover changes. These errors were mainly found in high-relief and actively changing moraines. We manually excluded several clearly erroneous values and smooth the data using robust locally weighted regression (MATLAB’s “rloess” method in the “smooth” function) using a 20-point (112 m) smoothing span. These smoothed velocity data were the targets for our modeling effort described in the following section.

3.2.4 Numerical model

We model cross-sectional velocity profiles using a 2D cross-sectional glacier flow model similar to that utilized in previous works [e.g., Nye, 1965; Amundson et al., 2006; Seddik et al., 2009]. Our approach employs an approximate force balance assuming a relatively uniform cross-sectional geometry in the longitudinal direction, in which we neglect all components of the deviatoric stress tensor except the shear stresses $\tau_{xy}$ and $\tau_{xz}$ (on the $x-z$ and $x-y$ planes, respectively, where $x$, $y$, and $z$ denote the longitudinal, cross-glacier and vertical coordinates). Correspondingly, we neglected horizontal velocity components and strain rate components other than $\dot{\epsilon}_{xy}$ and $\dot{\epsilon}_{xz}$. The approximate momentum equation is thus written as:

$$\frac{\partial \tau_{xy}}{\partial y} + \frac{\partial \tau_{xz}}{\partial z} = -\rho g \sin \alpha$$  \hspace{1cm} (3.1)

where $\rho$ is the density of ice, $g$ is gravitational acceleration, and $\alpha$ is the ice surface slope. The stresses are defined in terms of strain rates as:

$$\tau_{xy} = \mu \frac{\partial u}{\partial y} \quad \text{and} \quad \tau_{xz} = \mu \frac{\partial u}{\partial z}$$  \hspace{1cm} (3.2)

where $\mu$ is an effective ice viscosity, and $u(y, z)$ is the longitudinal velocity. Using Equa-
tion 3.2 in Equation 3.1 yields:

$$\frac{\partial}{\partial z} \left( \mu \frac{\partial u}{\partial z} \right) + \frac{\partial}{\partial y} \left( \mu \frac{\partial u}{\partial y} \right) = -\rho g \sin \alpha$$  \hspace{1cm} (3.3)

The effective viscosity is related to velocity field and strain rates using

$$\mu = \frac{1}{2} A^{-1/n} \dot{\epsilon}_e^{-1+1/n}$$  \hspace{1cm} (3.4)

where $A$ and $n$ are the Glen’s flow law rate factor and exponent, respectively. Here we assume $n = 3$. We discuss in detail our choice of the rate factor $A$ in the results section (Section 3.2.4). The effective strain rate in Equation 3.4 is defined as:

$$\dot{\epsilon}_e = \frac{1}{2} \sqrt{\left( \frac{\partial u}{\partial y} \right)^2 + \left( \frac{\partial u}{\partial z} \right)^2}$$  \hspace{1cm} (3.5)

The nonlinear dependence of $\mu$ on $u$ requires iterative solution of Equations 3.3-3.5 until the solution converges. We employed a Picard iteration in which we approximate $\mu$ in each iteration based on calculated using the velocity field from the previous iteration. We implemented the iterative solution of Equations 3.3-3.5 using the open-source finite element package FEniCS \cite{Logg et al., 2012}. Our finite element scheme used triangular Lagrange elements with a non-uniform grid, with 4500 elements in the domain. We used second order (quadratic) shape functions. Grid refinement studies confirmed that the numerical solutions converged with this grid resolution.

We applied a stress-free condition at the free surface and no regularization of the ice viscosity was needed. In various computations, we prescribed the boundary condition on the bed/valley walls as either specified velocities (e.g. zero velocity in the absence of basal sliding or a specified sliding velocity that may vary along the boundary) or as zero shear stress. Our approach for reconciling seasonal surface velocities with ice flow models was similar to that employed by Amundson et al. \cite{2006}. In their study, the basal geometry is well constrained and their main objective was to invert for the basal velocity variation using surface velocity and tiltmeter deformation data. However, in this study, the basal geometry
is not well known and for this reason we also inverted for the basal geometry. Amundson et al. [2006] employed a Dirichlet or specified velocity boundary condition at the bed in their inversions. As noted in their paper, the existence and uniqueness of the solution of a forward model based on Equations 3.1-3.5 above is well established, and a best-fit velocity field is uniquely determined regardless of the boundary condition employed. We followed a similar approach by prescribing basal velocities in most of our model simulations. However, we acknowledge that the inverse problem of estimating basal velocities based on surface velocity measurements is ill-posed.

We represent glacier valley geometry using power functions on either side of the maximum ice thickness ($H_{max}$), so that the bed elevation ($z_b$) is represented as:

$$
\begin{align*}
    z_b &= \begin{cases} 
        (H_{max} + z_{sl}) \left( \frac{y_c}{y} - \frac{y_c}{y_{max}} \right)^\beta - H_{max} & \text{if } y \leq y_c \\
        (H_{max} + z_{sr}) \left( \frac{y_c}{y} - \frac{y_c}{y_{max}} \right)^\gamma - H_{max} & \text{if } y > y_c 
    \end{cases}
\end{align*}
$$

(3.6)

where $y$ is the cross-glacier coordinate, $y_c$ is the location of maximum ice thickness ($H_{max}$), $z_{sl}$ and $z_{sr}$ are the measured surface elevations on the left and right side of the transect, respectively, $y_{max}$ is the cross-glacier location of maximum ice thickness, and $\beta$ and $\gamma$ are positive fitting parameters that determine the steepness of the valley walls. We verified the model solution against the shape factors provided by Nye [1965] for a variety of valley aspect ratios.

3.3 Results

3.3.1 Testing against GPS measurements

We maintained on-glacier GPS monuments in 2006, 2012, 2013, and 2014 that allow us to ground truth our satellite-derived glacier velocities (2006 data published in Bartholomaus et al. [2008, 2011]). We measure ice surface velocity from on-glacier GPS monuments that were in place during the July 15 to August 27 WorldView image periods. The GPS measurements agree well with image-derived velocities in both displacement magnitude and azimuth (Figure 3.3). The GPS data show 4.35 m of displacement between the July 15 and August
1 images used for velocity estimates in COSI-Corr. The mean COSI-Corr displacement in a 25 x 25 m box centered on the GPS location is 4.32 m, agreeing with the GPS displacement within 1%. The COSI-Corr estimated mean displacement azimuth is 125.2°, also agreeing with the GPS azimuth of 124.15° within 1% (Figure 3).

3.3.2 Seasonal evolution of glacier velocity

Glacier velocity is low during the early spring (Figure 3.4a), reaches a maximum in early summer (Figure 3.4b), and then slows through late summer (Figure 3.4c-d). In each velocity map, glacier velocity increases up-glacier and towards the glacier centerline (Figures 3.4 and 3.5). In spring (March 19 to April 26), maximum glacier velocity is 0.29 m d\(^{-1}\) (Figure 3.5a), which decreases almost linearly to near zero velocity 8 km down-glacier from the top of our study reach. The terminal 4 km as well as the lateral glacier margins exhibit negligible motion in spring. We are unable to produce a robust correlation from April to June because changing snow cover confounds the pixel-tracking software.

In early summer (June 19 to July 15), glacier velocity increases to peak observed values across the entire study area (Figure 3.4b). The maximum velocity (at the upstream end of the study reach) increases to 0.41 m d\(^{-1}\) and velocities at locations with near zero spring
motion (8-10 km down-glacier) increase to 0.05 m d$^{-1}$ (at 10 km) and 0.20 (at 8 km) m d$^{-1}$ (Figure 3.5a). Along the centerline transect, we find average velocity (over June 19 to July 15) increases by a factor of 1.4 relative to spring at the top of the study reach, growing to a factor of ~10 increase at ~9 km down-glacier, after which it monotonically decreases towards the glacier terminus (Figure 3.5a). Below ~11 km, we find spurious patches of high glacier velocity that likely reflect ice face retreat and debris cover changes. Along-flowline stripes of higher velocities in these maps (Figure 3.4) are debris-covered moraines, where ice face retreat (lateral retreat of inclined, bare ice walls), debris cover change, and high relief (which may cause lower quality orthorectification) add noise and spurious apparent motion to the correlation-based speed estimates.

The velocity maps from later in the summer (July 15 - August 1, August 1 - 27; Figure 3.4c-d) are qualitatively similar to the early summer pattern described above (Figure 3.4b). In the interval July 15 - August 1, the up-glacier reach (0-8 km) begins to slow from its peak during the June 19 - July 15 period, and continues to slow down over August 1 - 27. In late July, maximum velocity is 0.36 m d$^{-1}$. By late August, the maximum velocity decreases to 0.34 m d$^{-1}$ (Figure 3.5a).

Ice surface velocity varies significantly (0.15 m d$^{-1}$) between June 19 - August 27 in the middle of the study reach (4 km down-glacier, Figure 3.5a). In the terminal reach (8-12 km), however, velocity is more or less steady through the entire summer, varying in time by only 0.04 m d$^{-1}$ from June 19 to August 27. The observations between June 19 and August 27 indicate that, despite speeding up relative to spring, the marginal 500 m of the glacier cross-section show almost no variability in summer speed, while the centerline velocity varies by 0.10 m d$^{-1}$ (Figure 3.5b).

3.3.3 Spatiotemporal analysis of the summer velocity increase

We construct maps of the summer ice surface velocity increase relative to spring by subtracting the spring speeds from summer speeds (Figures 3.6-3.7). We find the summer speedup is much more uniform than the spring ice surface velocity (Figure 3.4a).

In early summer (June 19 - July 15), a 6 x 2 km swath of the glacier accelerates by
Figure 3.4: Maps of velocity of the terminal 15 km of Kennicott Glacier in a) spring, b) early-summer, c) late July, and d) August. Colorbar is the same in all figures. Panel a) shows the down-glacier coordinate system and transects used for analysis in Figures 3.5 and 3.7 and discussed in the text.
≥ 0.15 m d$^{-1}$ above the spring velocity in the same location (Figure 3.6; 1.7-8 km on Figure 3.7). The spring surface velocity ranges from 0.00-0.28 m d$^{-1}$ and the summer ice surface velocity varies from 0.20 to 0.38 m d$^{-1}$ over this same area (Figure 3.7). The summer velocity increase is relatively uniform from 1.7 to 8 km down-glacier, demonstrated by the broad peak in glacier speedup shown in Figure 3.7a. In addition, the central 1.5 km across the glacier exhibits a uniform speedup before decreasing in magnitude towards the glacier margins (Figure 3.7b). Despite this relative uniformity, we observe a smaller summer speedup in the first kilometer of the study transect, where total ice surface velocity is highest. At the top of the study reach, the summer velocity increase accounts for 40% of the total ice surface velocity. Near the terminus, the summer speedup accounts for 70-100% of the total summer ice surface velocity (shown as overlapping light and medium gray lines on Figure 3.7a). As stated above, along-flowline features reflect lower quality image correlation caused by the high relief, iceface melting, and debris cover change seen on moraines. As those velocities that exceed 0.20 m d$^{-1}$ on Figure 6a are associated with striping and other processing artifacts, we estimate that maximum early summer (June 19 - July 15) velocity increases are ≤ 0.20 m d$^{-1}$.

In late July (Figure 3.6b), the total ice surface velocity is lower, and thus the magnitude
of the speedup relative to spring is smaller. The maximum speedup at this time is $\sim 0.13 \text{ m d}^{-1}$, excluding spurious high velocities. The speedup velocity is still much more uniform than total ice surface velocity and we find smaller speedup velocities at the top of the study reach (Figure 3.8). In August a large portion of the glacier still moves $\sim 0.08 \text{ m d}^{-1}$ more quickly than it does in spring (Figure 3.6c). A 2 x 1.5 km patch of glacier centered 7 km down-glacier continues to flow $\sim 0.12 \text{ m d}^{-1}$ faster than in spring (Supplementary Figure 1), similar to the late July speedup magnitude in this area.

### 3.3.4 Modeling the summer speedup

We seek to explain the spatial pattern of the summer velocity increase using a 2D cross-sectional glacier model to capture cross-glacier variations in glacier velocity and to account for the importance of wall drag for ice dynamics on this relatively narrow (width/depth aspect ratio of $\sim 8$) valley glacier. The first step in this effort is to arrive at an acceptable representation of the conditions on the Kennicott Glacier that reproduce the observed spring (March 19 - April 26) surface speeds along the upper cross-section shown in Figure 3.4a. The major challenge in accomplishing this first step is that there are three poorly constrained factors to contend with: (i) the valley geometry, (ii) the winter/springtime sliding velocity, and (iii) the flow law rate factor $A$. Year-round “background” sliding on temperate glaciers
is possible [e.g., Raymond, 1971; Heinrichs et al., 1996; Amundson et al., 2006] and could explain a significant portion of our observed spring speeds. In addition, our estimate of ice thickness is poorly constrained due to sparse airborne radar coverage. On a temperate glacier the flow law parameter $A$ may vary due to water [Duval, 1977] or debris [Cohen, 2000] content, or spatial variation in ice fabric [Cuffey and Paterson, 2010]. The background sliding velocity may be estimated robustly by inverse modeling when basal geometry is well constrained [e.g., Amundson et al., 2006] and $A$ is known or assumed. However, in our less constrained situation, there is a tradeoff between the unknown ice thickness, spring sliding velocity, and rate factor that cannot be resolved uniquely by our inversion. For a fixed $A$, the combination of a higher background sliding velocity and a lower ice thickness (or vice versa) can produce the same surface velocity.

Thus, we begin with a sensitivity analysis to explore combinations of parameters that reproduce the observed spring surface velocity. We perform 54 simulations using a range of $A$ and $H_{\text{max}}$ values. For each pair of these values, we estimate the sliding velocity required to match the spring velocities (Figure 3.9). We hold valley geometry constant in all runs with the same $H_{\text{max}}$, and use an ice surface slope of 2.98%, the average slope over 2 km (approximately four ice thicknesses) measured from a DEM generated from a WorldView
Figure 3.8: (top) Partitioning July 15 - August 01 glacier velocity along the down-glacier transect. (bottom) Partitioning August 01 - 27 glacier velocity along the down-glacier transect. Medium gray lines represent observed summer ice surface velocities. Black lines are the March-April (spring) velocities. Light gray lines show the summer speedup, calculated as the difference of summer and spring ice surface speeds.
We consider three end-member scenarios, each of which reproduces the springtime surface velocities (Table 3.2; Figures 3.10-3.12. In each of these scenarios, the influence of one of the factors \((A, H_{\text{max}}, \text{springtime sliding})\) is emphasized. We do not claim that all these end-members are realistic, and present them only to demonstrate the sensitivity of inferred summer sliding to the assumed springtime conditions. All three scenarios considered reproduce the observed springtime surface velocities reasonably well, with RMSE values of 0.0155-0.0233 m d\(^{-1}\). The high rate factor case (Scenario B; Figure 3.11; Table 3.2) exhibited the lowest RMSE. The high spring basal speed case (Scenario A; Figure 3.10; Table 3.2) had the highest RMSE. The high ice thickness case (Scenario C; Figure 3.12; Table 3.2) produced an intermediate RMSE but we do not include figures corresponding to this scenario in the main body of the paper in the interest of space. In Scenario A, in which we impose \(H_{\text{max}} = 525\) m and the standard \(A = 2.4-24\) Pa\(^{-3}\) s\(^{-1}\), a springtime basal velocity of 0.175 m d\(^{-1}\) (70\% of the observed surface velocity) is required over 2/3 of the valley cross-section to match the surface speed. This percentage of spring surface velocity attributable to basal motion is not outside the range of previously reported values of 50-90\% [Raymond, 1971; Heinrichs et al., 1996; Truffer et al., 2001; Amundson et al., 2006], although data on the average fraction of spring motion attributable to sliding is not readily available, especially for a non-surge type valley glacier. We first sought to restrict \(H_{\text{max}}\) to 450 m for Scenario A to agree with radar estimates, but we were unable to explain the surface speedup without prescribing unrealistic basal velocity (e.g., basal velocity greater than the surface velocity or negative shear stresses); we have thus increased \(H_{\text{max}}\) to 525 m for this scenario.

Next, we seek to explain the observed summer velocities along the same transect by beginning from the three scenarios described above. We consider three schemes for specifying the basal velocity: a uniform sliding velocity along a fraction of the bed, high basal velocities in two “slippery” patches, and a basal velocity equal to the summer speedup (i.e. taken as the difference between the observed summer and springtime surface velocities). In the first two schemes, we estimated the spatial distribution and magnitude of the additional basal
velocity (i.e. over and above the springtime sliding velocity in Scenario A) required to match the observed summer velocities.

In the uniform sliding (i.e. single wide sliding patch) cases, additional basal velocities of 0.15-0.23 m d$^{-1}$ over 64-54% of the bed, respectively, are required to match surface velocities (Table 3.2; Figures 3.10-3.11). The width and magnitude of basal slip in the uniform sliding case is relatively insensitive to choice of scenario. Note that in Scenario A, which includes spring sliding, the total basal velocity is 0.33 m d$^{-1}$, the sum of the “base” and “uniform” cases on Table 3.2. While all of these three schemes produce a reasonable match to the maximum summer speeds, Scenarios A and C produce a superior fit (RMSE = 0.0260 and 0.0215, respectively; Table 3.2). Prescribing two narrower regions of high slip, basal velocities of 0.13-0.37 m d$^{-1}$ over 43-22% of the bed, respectively, are required to match surface velocities (Table 3.2; Figures 3.10-3.11). The model fits the observations slightly better for all scenarios in the two slippery patch case. In this case the width and magnitude of basal slip are sensitive to choice of scenario. The scenario with spring sliding (Scenario A) requires that a larger portion of the bed experience sliding than in Scenarios B and C to adequately explain the surface velocities. Prescribing the basal velocity as the surface speedup (i.e., difference between summer and spring surface velocities) under-predicts the peak surface velocity in all cases (Figures 3.10-3.11) and generally yields intermediate-to-high RMSE when compared to the other basal slip cases (Table 3.2). That the basal velocity does not equal the summer speedup reflects cross-glacier stress gradient coupling that reduces the magnitude and widens the surface expression of basal motion (Section 3.3.2).

We are aware that the abrupt transition from zero to high basal velocity, as prescribed above, is unrealistic and can induce a non-physical basal traction field [Hutter and Olunloyo, 1981; Schoof, 2004; Bueler and Brown, 2009]. We analyze the shear stress field for each slip distribution case of each scenario and find the maximum basal shear stresses are generally within a plausible range of 100-150 kPa. We do have cases in which a negative basal shear stress is calculated at the transition between slip and no-slip boundary conditions, but this is generally limited to a small number of elements near the transition. The high spring
sliding scenario (Scenario A) with two slippery patches, in which we impose the largest cross-glacier basal velocity gradients, produces the largest region of negative shear stresses. Previous authors have reported that this issue arises from a mathematical singularity at the transition from slip to no-slip [Hutter and Olunloyo, 1981; Schoof, 2004; Amundson et al., 2006]. We evaluate the model sensitivity to employing abrupt transitions in basal boundary condition by prescribing a smoothly varying basal velocity. We find the magnitude of the minimum basal shear stress increases (i.e., became less negative, or even positive), but observe no appreciable change in the overall pattern of ice deformation and surface velocity (Figure 3.13). For this reason, we use the simplest possible basal velocity patterns and do not parameterize the basal velocity with a continuous function.

3.4 Discussion

3.4.1 Advantages of optical satellite image correlation

Our weekly-to-monthly averaged WorldView-derived glacier velocities (Figure 3.4) agree remarkably well (to within 1% in both displacement magnitude and direction) with high-rate on-glacier differential GPS (Figure 3.3). This approach provides much greater spatial

<table>
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<th>Parameters</th>
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<th>Scenario B</th>
<th>Scenario C</th>
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<td>Width 1 [m]</td>
<td>$u_b$ 1 [m d$^{-1}$]</td>
<td>C. 2* [m]</td>
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<td>Speedup</td>
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</table>

Table 3.2: Parameter choices for each end-member parameter sensitivity scenario and associated best-fitting basal slippery patch center location, width and magnitude of sliding velocity. ‘Base’ refers to parameters best fitting the spring velocity, ‘Uniform’ refers to a uniform sliding velocity within a single patch, and ‘Two patch’ refers to two slippery patches. ‘Speedup’ refers to prescribing the observed summer speedup as the basal velocity. For (A) the basal motion described in ‘uniform’ and ‘two patch’ is superimposed on top of ‘base’. The RMSE between modeled and measured surface velocities is shown in the last column. *Center location of the second slippery patch. **Width of the second slippery patch.
Figure 3.9: Model parameter values that produce good fit between measured and modeled ice surface velocity along the modeled cross-section. We performed 54 model runs using fixed values for the maximum ice thickness and rate factor \( A \). Crosses show the parameters of each model run used to generate contours. The spatial extent of slip is fixed through all model runs. Parameters of three end-member base case scenarios discussed in text are shown as A (high spring basal motion), B (high rate factor), and C (high maximum ice thickness).
Figure 3.10: Results from Scenario A, which includes spring basal motion to match the observed spring surface speed. a) Modeled ice surface speeds under several prescribed basal velocity scenarios. Crosses indicate observed velocities in spring (black) and mid-summer (pink); dashed lines indicate prescribed basal velocities in our 2D cross-sectional flow model; solid lines show modeled ice surface velocities, where line colors indicate the corresponding basal velocity field. The basal velocity depicted by the red dashed line is equivalent to the measured summer speedup across the transect. The dotted lines show the summer increase in basal motion relative to spring. Model misfit and details are given on Table 3.2. b) Modeled glacier geometry and associated model parameters (values in Table 3.2). Colors indicate out-of-plane (longitudinal) velocity.
Figure 3.11: Results from Scenario B, which employs a high rate factor \( A \) to match the observed spring surface speed. a) Modeled ice surface speeds under several prescribed basal velocity scenarios. Crosses indicate observed velocities in spring (black) and mid-summer (pink); dashed lines indicate prescribed basal velocities in our 2D cross-sectional flow model; solid lines show modeled ice surface velocities, where line colors indicate the corresponding basal velocity. The basal velocity depicted by the red dashed line is equivalent to the measured summer speedup across the transect. Model misfit and details are given on Table 3.2. b) Modeled glacier geometry and associated model parameters (values in Table 3.2). Colors indicate out-of-plane (longitudinal) velocity.
Figure 3.12: Results from Scenario C, which employs a large ice thickness \( H \) to match the observed spring surface speed. 

a) Modeled ice surface speeds under several prescribed basal velocity scenarios. Crosses indicate observed velocities in spring (black) and mid-summer (pink); dashed lines indicate prescribed basal velocities in our 2D cross-sectional flow model; solid lines show modeled ice surface velocities, where line colors indicate the corresponding basal velocity. The basal velocity depicted by the red dashed line is equivalent to the measured summer speedup across the transect. Model misfit and details are given on Table 2. 

b) Modeled glacier geometry and associated model parameters (values in Table 3.2). Colors indicate out-of-plane (longitudinal) velocity.
Figure 3.13: a) Prescribed basal (dashed) and modeled surface (solid) velocities given continuous/smooth (blue) and discontinuous/sharp (red) basal boundary conditions. b) Cross-sectional plot of velocity difference between the continuous and discontinuous solutions. Note the \(10^{-3}\) scale on the colorbar, indicating that velocities agree to within 1 mm d\(^{-1}\) throughout most of the glacier. c) Velocity difference between the continuous and discontinuous solutions evaluated at the glacier surface. Again, note the \(10^{-3}\) scale on the y-axis.
coverage and resolution than GPS (5.6 m pixels over the terminal ~ 46 km\(^2\), compared to 1 GPS monument every 3 km along a 15 km centerline transect), but sacrifices temporal resolution. The WorldView satellite repeat time is 16 days, which places a limit on the best possible temporal resolution of the method. Our method’s coarse temporal resolution is counterbalanced by high spatial resolution and extent. Without large spatial coverage, for example, we would not be able to document the widespread uniformity of the summer speedup.

Unlike microwave-wavelength interferometric synthetic aperture radar (InSAR), extensive cloud cover precludes correlation of optical satellite imagery to produce velocity maps. As this is common in polar and alpine environments, velocity fields must often be averaged over longer timespans than the best-case 16-day window. Optical image correlation, however, appears to be more robust against temporal decorrelation, which is a difficulty for summertime InSAR velocity estimates [Burgess et al., 2013a]. In addition, velocity estimation from InSAR is difficult on high-relief and time-variable surface elevations typical of alpine glaciers [Tedesco, 2015]. Our method of orthorectifying optical imagery with respect to the DEM produced from the concurrent stereopair reduces errors associated with changing glacier geometry, allowing for high-accuracy velocity determination on alpine glaciers.

3.4.2 Interpreting the spatiotemporal pattern of the summer speedup

The temporal evolution of glacier velocity (Figures 3.4-3.5) is consistent with the cycle commonly observed on both alpine glaciers [e.g., Hooke et al., 1989; Mair et al., 2001; MacGregor et al., 2005; Bartholomaus et al., 2011] and Greenland outlet glaciers [Joughin et al., 2008; Sole et al., 2011; Moon et al., 2014] in which seasonal variability in both meltwater production and the evolving hydraulic system of the glacier result in fluctuations in subglacial water pressure and basal velocity [Iken and Bindschadler, 1986]. We find that the speedup (summer minus spring surface velocity) reaches a maximum of ~ 0.15 m d\(^{-1}\) during early summer (June 19 - July 15) and then declines through the end of our observation period (Figure 3.6). The spatial distribution of the speedup is remarkably uniform, with a 12 km\(^2\) area speeding up by 0.15-0.20 m d\(^{-1}\) during the early summer (June 19 - July 15;
Figures 3.6a and 3.7a). The total ice surface velocity, in both spring and summer, varies significantly over this area (Figures 3.4 and 3.7). At a given point, the early summer velocity is 1.4 to 10 times higher than the spring velocity, with areas of high spring speeds exhibiting smaller fractional velocity increases. The ten-fold speedup is found near the lateral glacier margins and terminus, likely reflecting the influence of stress gradient coupling, as we discuss in greater detail below. The 40% relative speedup (1.4-fold velocity increase) found in faster-moving ice is similar to that observed on Greenland outlet glaciers [Hoffman et al., 2011; Tedstone et al., 2014].

We consult our on-glacier GPS for the July 15 - August 1 and August 1 - 27 image correlation timespans to ensure that this apparent spatial uniformity does not result from temporally averaging over spatially variable dynamics (e.g., wave-like propagation of regions of high basal motion; [e.g., Kamb and Engelhardt, 1987; Anderson et al., 2004; Harper et al., 2007]. We see no evidence for wave-like behavior; GPS velocities fluctuate on diurnal and multi-day timescales, but all stations vary synchronously (Figure 3.14). We could not perform this analysis for the March 19 - April 26 timespan because we only had 1 active GPS monument at that time; the same is true for the June 15 - July 19 timespan due to GPS power loss.

As it is well recognized that the relationship between basal motion and its signature on the ice surface is complex [Balise and Raymond, 1985; Kamb and Echelmeyer, 1986; Gudmundsson, 2003], we resort to modeling to explore how the signal of basal motion will be transmitted through the body of the glacier. As noted in Section 3.2.4, there are significant uncertainties involved in inferring basal motion based on the presently available data on Kennicott Glacier. Uncertainty in ice thickness particularly confounds efforts to infer basal motion. Even if the valley geometry were well constrained, our modeling results suggest that there is a tradeoff between the magnitude of the prescribed basal velocity and the proportion of the bed that is slipping, such that two end-member basal slip patterns (one in which a large portion of the bed slides at a moderate speed, the other in which a small portion of the bed slides rapidly) can produce similar surface velocities (Figures 3.10a and 3.11a).
Figure 3.14: Down-glacier velocity at three on-glacier GPS monuments, offset by subtracting the pre-spring event velocity at each station. Dashed lines show mean velocity over two WorldView image correlations. The mean speedup is uniform over this reach, while the spring velocities are 0.17 m d$^{-1}$ at GPS3, 0.26 m d$^{-1}$ at GPS4, and 0.31 m d$^{-1}$ at GPS5. GPS3 is furthest down-glacier, with GPS4 and GPS5 located approximately 3 and 6 km up-glacier, respectively.
relatively small basal velocity is required if much of the bed slips; this might simulate an extensive zone of high water pressure associated with a distributed drainage system (blue lines on Figures 3.10a and 3.11a). A wide, relatively uniform cross-sectional distribution of basal slip has been observed or inferred in several studies [Raymond, 1971; Truffer et al., 2001; Amundson et al., 2006]. If, on the other hand, a channelized drainage system is in place and a much smaller portion of the bed slips (perhaps due to high water pressure when the capacity of the channel is overwhelmed, for example, as observed and modeled by Hubbard et al. [1995] and Werder et al. [2013]) then a much higher basal velocity is required to match the observed surface speedup (green lines on Figures 3.10a and 3.11a). It is interesting to note that the location of maximum summer speedup is offset from the glacier centerline (Figures 3.4 and 3.6b) and, regardless of the chosen pattern for basal slip, our modeling suggests a higher basal velocity on the side of the bed into which the ice-marginal Hidden Creek Lake (HCL) would drain during its annual outburst flood (Figures 3.10a and 3.11a). The drainage of HCL produces an annual speedup event [Anderson et al., 2005; Bartholomaus et al., 2008], that occurred around June 29 in 2013 and is thus captured by the June 19 - July 15 image correlation (Figures 3.4b and 3.6a). The higher basal sliding velocity in the left patch in Figures 3.10a and 3.11a could reflect HCL drainage altering basal conditions that in turn influence the speedup. Amundson et al. [2006] also inferred a region of high basal slip offset to one side on Black Rapids Glacier, which corresponds to the drainage pathway of an ice marginal lake.

The red lines on Figures 3.10a and 3.11a show the modeled surface velocity using the observed summer speedup as the input basal velocity distribution. Glaciological studies often interpret the difference between summer velocity and a minimum velocity observed in winter or spring as the “sliding speed” at a point [e.g., Zwally et al., 2002; Anderson et al., 2004; Bartholomaus et al., 2011; Tedstone et al., 2014]. We show here that while using the surface speedup as a proxy for the basal velocity may indeed account for the speedup field, there are many other plausible basal velocity patterns that can generate the same speedup pattern (Figures 3.10a and 3.11a). Previous studies have highlighted this
point, demonstrating that neither the glacier surface velocity nor the basal velocity field are defined by local quantities [Kamb and Echelmeyer, 1986; Blatter et al., 1998; Price et al., 2008]. Prescribing the basal velocity based on the summer speedup leads to a slightly higher RMSE in fitting the surface velocity, compared to the two other prescribed sliding patterns. Stress gradient coupling reduces the magnitude and both spreads and smooths the surface expression of the prescribed basal sliding velocity; this creates the difficulty in distinguishing between these slip patterns and confounds interpretation of the difference between summer velocity and a winter/spring velocity at a point as the “sliding speed” at that location.

The insensitivity of our modeled speedup to the prescribed basal velocity pattern likely reflects strong cross-glacier stress gradient coupling in the ice. The ice thickness at the location of the modeled transect is approximately 450 m (M. Truffer, unpublished data, personal communication, 2014), meaning that our observed speedup is uniform over approximately 12 (4) ice thicknesses in the down-glacier (cross-glacier) direction. This would be reduced to 10 (3) ice thicknesses if the large ice thickness Scenario C (Table 3.2; Figure 3.12) is closer to reality. While 4 ice thicknesses is the theoretically expected stress gradient coupling length scale for an alpine glacier, larger coupling length scales are expected when basal motion comprises a large fraction of ice surface displacement, potentially reaching 12 ice thicknesses for surging glaciers [Kamb and Echelmeyer, 1986]. While Kennicott Glacier is not surging, its surface velocity may reflect a strong influence of basal motion (Figure 3.10; Sections 3.2.4 and 3.3.2). If this is true, stress gradient coupling may largely explain the uniformity of the speed-up. If background basal motion is less important (Figure 3.11) and the coupling length scale is smaller, this leaves us with the alternative that there may exist a relatively uniform distribution of effective pressure at the bed driving uniform basal motion across this broad terminal region.

3.4.3 The difficulty in inferring basal conditions from the pattern of surface speedup

Motivated by the results presented in Section 3.2.4, we ask the question: when would we expect to be able to distinguish between several narrow slippery patches and one wide region of slip? The answer to this question is important for remotely sensed as well as field-based
studies in which the goal is to infer basal slip and associated subglacial hydraulics from surface observations. We address this question using a numerical experiment in which we progressively widen the distance between two identical basal slippery spots and identify the separation distance required to resolve the two distinct slippery spots (Figure 3.15).

We conducted these numerical experiments by non-dimensionalizing the problem, which collapses results from trials with disparate ice thicknesses, rate factors, aspect ratios, and prescribed basal velocities. The results can be readily converted into dimensional units to consider a particular system of interest. We non-dimensionalize all lengths by the mean ice thickness ($H$),

$$y_\ast = y/H; \lambda_\ast = \lambda/H; \text{ and } W_\ast = W/H$$

where $y$ is the cross-glacier coordinate, $\lambda$ is the slippery patch width, and $W$ is the glacier width. We non-dimensionalize surface and basal velocities by the expected deformation velocity calculated using the one-dimensional shallow ice approximation without a shape factor (i.e. infinitely wide glacier),

$$u_\ast = u_{s|b}/u_{def} = u_{s|b} \left[ \frac{2A}{n+1} \left( \rho g \sin \alpha \right)^n H^{n+1} \right]$$

where $u_{s|b}$ is the surface or basal velocity, respectively, $u_{def}$ is the expected deformation velocity, and the rest of the terms are defined in Section 3.2.4. We model the glacier using a nearly rectangular valley with wall steepness exponents of $\beta = \gamma = 10$ (Equation 3.6), a surface slope of $0.03 \text{ m m}^{-1}$, and a rate factor of $A = 2.4 \times 10^{-24} \text{ Pa}^{-3} \text{ s}^{-1}$ [Cuffey and Paterson, 2010].

In all trials, we assign $\lambda = H/2$, $W = 40H$, and $u_b = u_{def}/2$. Between trials, we increase the separation distance between slippery patches from $0H$ to $16H$ and numerically model the surface velocity (Figure 3.15). First, we note that at any separation distance, the surface expression of basal slip is distributed over a much wider distance than the region of active basal slip due to stress gradient coupling. At small separation distances, the basal drag
associated with the intervening sticky spot is negligible and increasing the separation distance increases the maximum surface speedup. We find the maximum surface velocity occurs when the separation distance is equal to $H/2$ but note that the surface speedup is only 35 percent of the prescribed basal velocity (the ratio of modeled surface speedup to prescribed basal velocity increases to 1 as $\lambda$ approaches $W$). As the separation distance is increased further, the maximum speedup decreases but a greater fraction of the surface experiences elevated velocity. Only when the separation distance is greater than $4H$ do the individual slippery spots express themselves in a “double humped” speedup profile (Figure 3.15). At this point, although the two slippery spots are resolvable in our model output, they would be extremely difficult to detect in noisy field observations of temporally varying surface velocities, until the separation distance is yet larger. Even when the separation distance is $20H$ the velocity difference between peak and trough is only $\sim 0.1$, which in dimensional terms would be $0.05 \text{ m d}^{-1}$ for a glacier with a deformation velocity of $0.5 \text{ m d}^{-1}$. This difference would be difficult to detect without very precise velocity measurements.

Adding to this difficulty, wall drag further damps any heterogeneity in the surface speedup in narrower glacier valleys (Figure 3.16). In the “infinitely wide” model ($W/H = 40$), the “double humped” speedup pattern is just barely distinguishable at a separation distance of $4H$. When $W/H$ is decreased to 20, the two slippery spots are less recognizable in the surface velocity pattern and are indistinguishable when $W/H = 10$, most similar to that found in a confined valley glacier.

In additional sensitivity studies, we found that the spatial distribution and magnitude of the surface speedup is sensitive to the width of the slipping region, as well as the magnitude of the prescribed basal velocity relative to the glacier’s deformation velocity, but not to the rate factor $A$, per se. That is, in model runs with differing $A$, the surface speedup in response to a given basal velocity field will be exactly the same if the surface slope increases to compensate for a lower $A$ such that the deformation velocity (term in brackets in Equation 3.8) is constant (Figure 3.17). By rearranging Equations 3.3-3.4 such that both $A$ and $\alpha$ appear in the driving stress term, it is readily seen that the velocity solution is
unchanged for identical values of the product of $A^{1/n}$ and $\sin \alpha$.

The above results indicate that we cannot expect to resolve variations in basal motion from surface observations until those variations occur over distances many times the ice thickness. These findings echo those of previous authors who found basal perturbations are not faithfully transmitted to the surface until those perturbations occur on a length scale equivalent to many times the ice thickness [Balise and Raymond, 1985; Kamb and Echelmeyer, 1986; Raymond and Gudmundsson, 2005]. We therefore argue that the influence of cross-glacier stress gradient coupling between regions of low and high effective pressure [Blatter et al., 1998; Price et al., 2008] can explain the widespread uniformity of surface speedups in valley glaciers (and ice sheets, although to a lesser extent due to their closer similarity to the infinitely wide case) as well as other alternative mechanisms that have been proposed, such as uniform hydrology [e.g., Tedstone et al., 2014]. This is true for any glacier but in situations where basal slip contributes significantly to surface velocity, the effective ice coupling length scale increases [Kamb and Echelmeyer, 1986; Gudmundsson, 2003], and slippery spots must be even further separated before they are resolved in the ice surface speedup.

3.4.4 Implications for glacier models

A major confounding factor in unraveling the mechanisms controlling basal motion on Kennicott Glacier is the poorly constrained basal geometry. As a result, there is a tradeoff between ice thickness, flow law parameter $A$, and springtime background basal velocity; and springtime motion may be explained by several end-member scenarios (Figures 3.10-3.12, Table 3.2). However, we also find that several end-member basal velocity configurations can explain the summer speedup equally well (Figures 3.10a and 3.11). One model run is meant to simulate widespread basal slip associated with a distributed cavity drainage system, while another simulates high localized slip that may occur near overwhelmed channelized drainage. Our third test case does nearly as well (although it generally produces the highest RMSE) by employing a common glaciological practice of estimating the basal velocity as the difference between summer and early spring surface velocity. While this non-uniqueness frustrates
Figure 3.15: Location of prescribed basal slip (dashed) and modeled surface velocities (solid) in a glacier with an aspect ratio of $W/H = 40$ (effectively infinitely wide). Surface velocity in the absence of basal motion is shown in black. Basal slip magnitude ($u_b/u_{def} = 0.5$) and extent (two patches $\lambda = H/2$ wide) is constant in all trials. We are unable to resolve individual slippery spots until they are separated by $\geq 4H$. 

\[ u_\text{max} \text{ at } 0.5H \text{ separation} \]

\[ 4H \text{ separation} \]

\[ \text{Surface velocity in no-slip case} \]

\[ \text{Modeled surface velocity} \]

\[ \text{Location of prescribed basal slip (dashed)} \]

\[ \text{Basal velocity not to scale; } u_b = 0.5u_{def} \]
Figure 3.16: Comparison of the influence of prescribed basal velocities (dashed) on modeled surface speed (solid) on glaciers with varied aspect ratios. A double-humped speedup is evident in the $W/H = 40$ (infinitely wide case) and just barely in the $W/H = 20$ case. The surface expression of basal slip is muted in the $W/H = 10$ case (similar to Kennicott) due to wall friction and requirement of zero velocity at the glacier margins. Basal slip magnitude ($u_b/u_{def} = 0.5$) and extent (two patches $\lambda = H/2$ wide) is constant in all trials. The vertical offset between surface velocities is due to wall friction reducing the maximum surface velocity as $W/H$ decreases.
Figure 3.17: Model results demonstrating insensitivity of results from our separation distance experiment to choice of rate factor, $A$. Dimensional prescribed basal (dashed) and modeled surface (solid) velocities using four different values of $A$. Note that all curves plot on top of each other and have been offset for visibility.
the exercise of probing basal phenomena using surface data, it offers promise for numerical modeling efforts of glaciers and ice sheets. If disparate basal slip patterns can explain our observed summer speedup equally well, then it is less important to know the detailed slip pattern to accurately predict integrated glacier behavior. Indeed, Tedstone et al. [2014] found that proximity to an inferred subglacial conduit is unimportant for determining the proportion of annual motion that is accomplished in summer, and suggest that a high-complexity description of subglacial hydraulics may not be required for accurate ice flow models. Our findings echo these results. Presumably basal hydraulic conditions vary considerably over our study reach in both space and time, and yet we find a uniform speedup magnitude over a length scale equal to 12 (4) ice thicknesses in the down- (cross) glacier direction. Our modeling results show that strong cross-glacier stress gradient coupling effectively acts to smooth variations in basal velocity, resulting in similar surface speedup patterns despite varied basal velocity configurations (Figures 3.10-3.12, 3.15-3.16). This may allow for simpler parameterizations of the increased ice flux associated with hydraulically-induced speedup without requiring a detailed description of basal conditions.

3.5 Conclusions

We document the evolution of ice surface velocity on Kennicott Glacier from March 19 to August 27, 2013 using cross-correlation of WorldView optical satellite imagery. We find the satellite-derived velocity estimates agree with on-glacier GPS motion to within 1% in both displacement azimuth and magnitude. The high spatial resolution of WorldView imagery allows analysis of glacier dynamics over intra-seasonal timescales. We document the often-observed cycle in which glacier velocity reaches a maximum in early summer and slows until fall. The summer speedup is much more uniform than the spring ice surface speed, indicating that the pattern of the summer speedup cannot simply be assumed to be a scaled version of the deformation velocity field, at least in the terminal reach of this alpine glacier. We employ a 2D cross-sectional numerical flow model in an attempt to explain the observed spring surface velocity field but are unable to distinguish between widely distributed moderate-speed basal motion and narrower high basal velocity zones from surface
observations. Our modeling shows that ice surface velocities cannot reveal distinct regions of high basal velocity until they are separated by at least four ice thicknesses, and that this problem is exacerbated by the relatively low width-to-depth ratios of alpine glaciers such as Kennicott. The insensitivity of the glacier surface velocity to the details of the basal velocity field reflects the influence of strong transverse and longitudinal stress gradient ice coupling in determining the ice surface velocity, but may provide an excuse to employ simpler representations of glacier basal hydraulics in glacier models.

3.6 Acknowledgements

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Chapter 4

Spatial patterns of summer speedup on south-central Alaska glaciers

Abstract

Basal motion of glaciers is attributable to seasonally varying hydrology, and both increases ice flux and promotes erosion of the glacier bed. Tracking features in Landsat 8 imagery, we document differences between spatial patterns of summer and winter glacier ice surface speed, which reflect changes in the distribution of basal motion. Of 64 glacier profiles from ice divide to terminus across the Wrangell-St Elias Ranges of Alaska, we identify summer speedup on 13 glaciers. Where observed, the speedup is relatively uniform over much of the ablation zone, and varies by only a factor of $\sim 2$ between glaciers whose velocities span an order of magnitude. Summer speedup extends up to $\sim 30$ km up-glacier from the terminus, and often ends at the base of an icefall. These data suggest the possibility of simple parameterizations of enhanced summer basal motion in glacier models and hint at a mechanism for the formation of icefalls.

This chapter is in revision:
4.1 Introduction

Variations in glacier motion on sub-annual timescales are due to hydraulically-forced variations in glacier basal motion [Iken, 1981; Willis, 1995; Iken and Truffer, 1997; Mair et al., 2002; Anderson et al., 2004; Bartholomaus et al., 2008; Das et al., 2008; Shepherd et al., 2009; Andrews et al., 2014] and observations of these surface variations can be used to infer the spatial distribution of basal motion. When meltwater inputs to the glacier bed exceed the ability of the subglacial hydrologic system to transport that melt, the subglacial hydrologic system pressurizes, lubricating the ice-bed interface, inducing basal slip and causing a speedup at the ice surface [Bartholomaus et al., 2008; Das et al., 2008; Schoof, 2010]. Typically, glacier motion varies seasonally, with high speeds in spring with the onset of surface melt, intermediate mean velocity with large amplitude diurnal and few-day velocity fluctuations in summer, slowest speeds in late summer/fall, and gradual recovery to a relatively steady slow-intermediate velocity in winter.

Over short length scales (on the order of several ice thicknesses), stress gradient coupling [Kamb and Echelmeyer, 1986; Blatter et al., 1998] obscures spatial variability in basal motion when observed at the ice surface [Balise and Raymond, 1985; Gudmundsson, 2003; Raymond and Gudmundsson, 2005; Armstrong et al., 2016], especially in the low width-to-depth aspect ratio valley geometries common in alpine glaciers [Armstrong et al., 2016]. Examination of longitudinal profiles of velocity variation over many 10s of km (\(\sim 10^1 \text{ to } 10^2\) ice thicknesses), ensures that our observations reflect local variations in basal motion, rather than variations associated with ice coupling [Howat et al., 2008; Price et al., 2008; Flowers et al., 2016]. Such observations provide constraints on our understanding of the relationship between basal motion and controlling mechanisms, such as subglacial hydrology [Schoof, 2010; Hewitt, 2013; Hoffman and Price, 2014], subglacial substrate properties [Alley et al., 1986; Truffer et al., 2001], and glacier geometry [Weertman, 1957]. In addition, as basal motion can comprise a large fraction of a glacier’s surface speed [Raymond, 1971; Truffer, 2004; Amundson et al., 2006], particularly on fast flowing glaciers, understanding patterns
and controls on basal motion is critical for understanding spatially heterogeneous patterns of glacier response to climate warming [Joughin et al., 2010a; Bjørk et al., 2012; Moon et al., 2012]. Finally, as basal motion drives erosion of glacier beds [Hallet, 1979; Iverson, 1991; Beaud et al., 2014], knowing its distribution is key for understanding the evolution of arctic and alpine landscapes over geologic time [Harbor, 1992; MacGregor et al., 2000; Anderson et al., 2006; Herman et al., 2011; Egholm et al., 2012; Anderson, 2014].

Advances in computing power and satellite technology have led to a boom in large spatial and temporal scale glacier velocity mapping from cross-correlation of optical satellite imagery [Dehecq et al., 2015; Jeong and Howat, 2015; Rosenau et al., 2015; Fahnestock et al., 2016]. Analyses of such records have typically focused on decadal trends in glacier velocity [e.g., Heid and Kääb, 2012]. Systematic analysis of sub-annual variations in glacier velocity over a large spatial scale are more rare, and where they do occur, they have generally focused on temporal variations in glacier velocity at a point in space [e.g., Moon et al., 2014]. Such studies have noted marked spatial heterogeneity in temporal patterns of glacier speedup [Moon et al., 2012], but little data exist on the spatial patterns of glacier speedup that would allow exploration of the physical mechanisms that control these patterns. In this study, we document spatial patterns in glacier speedup across nearly 50 south-central Alaska glaciers. This work builds upon previous large-scale remote sensing work characterizing Alaska glacier surge dynamics [Abe and Furuya, 2015] and interannual velocity variability [Burgess et al., 2013b; Waechter et al., 2015] to provide observational constraint on conceptual and numerical modeling of the link between subglacial hydrology and glacier sliding.

4.2 Methods

4.2.1 Study area

The Wrangell and St Elias ranges (WRST) of south-central Alaska (Figure 4.1b) are one of the most heavily glacierized sub-polar regions. Changes in WRST glaciers are responsible for approximately half of Alaska ice mass loss [Berthier et al., 2010], which is accelerating in the 21st century [Das et al., 2014]. In order to facilitate data interpretation, we exclude
from analysis glaciers that underwent an active surge phase during our study. Our study glaciers comprise 10,600 km² of glacierized area (12% of Alaska glacierized area); individual glaciers range 4 - 91 km in length, with a median length of 25 km. The glaciers are all land-terminating, with a median terminus elevation of 1050 m a.s.l.

4.2.2 Seasonal glacier velocity from image cross-correlation

We estimate glacier velocity across WRST glaciers using cross-correlation of 15 m pixel Landsat 8 panchromatic satellite imagery. This method finds the pixel offsets required to maximize the spatial correlation of small image subsets in two images of the same geographic location separated in time [Scambos et al., 1992]. We use the Python-implemented image correlation program PyCorr described in Fahnestock et al. [2016]. The high geolocation accuracy and high radiometric resolution of Landsat 8 [Jeong and Howat, 2015; Fahnestock et al., 2016], allow determination of glacier velocity over large spatiotemporal domains, extending from the glacier headwall to the terminus. The velocity determinations used here are an early version of an Alaska subset of the global GoLIVE Landsat 8 ice velocity data archived at the National Snow and Ice Data Center (http://nsidc.org/data/NSIDC-0710).

The derived velocity field has 150 m postings and is a spatial average over the template chip and a temporal average over the time between two images (Section 4.7.1). The majority of image pairs we employ are separated by ≤ 48 days, and range from 16 to 160 days (Figure 4.2). We group velocity fields into summer (01 May to 30 September, DOYs 121-273) and winter (01 October to 30 April, DOYs 274-120) seasons (Section 4.7.1). The start and end dates of the summer group approximately correspond to when mean daily air temperature is above freezing at local weather stations. We then aggregate seasonal glacier velocity profiles across years. While this approach assumes glacier behavior in a given season in one year is similar to that in other years, which is not strictly true [Burgess et al., 2013a], the procedure is required to increase sample sizes for the statistical approaches we employ to characterize seasonal velocity profiles.

We then extract longitudinal profiles of glacier velocity along 64 centerlines (Figure 4.1c). We employ swath profiling to reduce image noise (Section 4.7.1), which are then filtered
Figure 4.1: a) Median ice surface velocity across WRST glaciers. b) Map of Alaska with red box extent of a and c. c) Spatial distribution of glacier speedup types across the study region. Red glaciers show a summer speedup; blue glaciers do not appear to a speedup; white glaciers lack sufficient data to determine speedup behavior; yellow glaciers have one branch speeding up while the other does not; gray glaciers were excluded from the study for ice dynamical reasons. Thin black lines show longitudinal profile locations. Heavy red lines show upglacier speedup extent. Where two red lines exist, the lower line indicates the down-glacier speedup extent. Heavy blue lines show approximate late summer snowline location. Arrow: location of Kennicott Glacier, profiled in Figure 4.5. d) Glacier speedup type shown in length-area space.
Finally, we determine median longitudinal velocity profiles and their interquartile range for both winter and summer seasons. We then subtract winter from summer velocity profiles to produce “summer speedup” profiles (Section 4.7.1). To determine the up-glacier distance to which the summer speedup extends (the “speedup length”), we employ the Mann-Whitney-Wilcoxon rank sum test (Section 4.7.2). We calculate the best linear fit to the down-glacier speedup; speedup magnitude is defined as the mean value of this fit line and speedup gradient is its slope.
Figure 4.3: Effect of filtering on mean velocity profiles and data noise level. Colored dots show each velocity observation used to calculate median (solid lines) and interquartile range (filled areas). Yellow data are filtered only on correlation output; red data are filtered on swath profiler results; blue are final data that have removed temporal outliers at each point.
4.3 Results

We develop an algorithm to characterize glacier surface velocity profiles from headwall to terminus. This improves upon previous large-scale Alaska glacier velocity mapping efforts [e.g., Burgess et al., 2013b] by extending velocity fields into these high reaches and including summer velocity fields. The general velocity profile is as expected (Figure 4.1a): minimum velocities are found at the glacier headwalls and termini and reach maximum values at mid-glacier locations that vary due to individual glacier geometry. The maximum velocities on the long, low-slope glacier tongues of non-surge type WRST glaciers are generally between 0.50-1.0 m d\(^{-1}\), but can reach 2.3-2.7 m d\(^{-1}\) in icefalls (Figure 4.1a).

4.3.1 Profiles of glacier summer speedup

For glaciers with small within-season velocity variability (Figure S2), we can detect seasonal velocity changes as small as 0.05-0.10 m d\(^{-1}\). Our median glacier-wide within-season interquartile velocity range is 0.075 m d\(^{-1}\) on glaciers for which we identify speedup behavior (“select glaciers”; colored glaciers on Figure 4.1c). We can detect seasonal velocity changes smaller than this where smooth spatial variation in the speedup field indicates a systematic change.

We find that 13 of 64 centerline profiles display a clear summer speedup (e.g., Figure 4.5b-c; Figure 4.1c; Table 4.1). Employing the rank sum test (Figure 4.5e; Sections 4.2.2 and 4.7.2), we find the median speedup length is 21 km, with an interquartile range of 9 to 32 km. This includes glaciers of different lengths; when normalized by total glacier length (\(L\)), the speedup length increases as a nearly constant fraction of glacier length (\(\sim 0.6 - 0.8L\); Figure 4.6c). However, on glaciers longer than \(\sim 40\) km, the speedup length appears to reach a maximum at 30 km, and thus comprises a declining fraction (\(\sim 0.3 - 0.4L\)) of total glacier length (Figure 4.6c).

Generally, for the glaciers that do speed up in summer, we find the terminal 60-80% of the glacier experiences a relatively uniform speedup of \(\sim 0.10\) m d\(^{-1}\), with an abrupt up-glacier transition to no difference in seasonal velocity (Figure 4.5c). We observe the median
Table 4.1: Attributes of study glaciers: part 1.  

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1 Many glaciers are unnamed in the Randolph Glacier Inventory. Tributary profiles are also unnamed. ²N/E/S/W = north/east/south/west/; B = branch; CL = centerline. ³Distance is relative to the terminus. First (second) number indicates speedup start (end) location. ⁴Units of speedup gradient are m d⁻¹ m⁻¹. ⁵Indicates profiles where rank sum speedup length was manually modified for speedup magnitude and gradient calculations. Rows are bolded where speedup behavior is identified.
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Table 4.2: Attributes of study glaciers: continued. ¹Many glaciers are unnamed in the Randolph Glacier Inventory. Tributary profiles are also unnamed. ²N/E/S/W = north/east/south/west/; B = branch; CL = centerline. ³Distance is relative to the terminus. First (second) number indicates speedup start (end) location. ⁴Units of speedup gradient are m d⁻¹ m⁻¹. ⁵Indicates profiles where rank sum speedup length was manually modified for speedup magnitude and gradient calculations. Rows are bolded where speedup behavior is identified.
mean speedup over a down-glacier reach to be 0.11 m d\(^{-1}\), with a total range of 0.07 to 0.18 m d\(^{-1}\) (Figure 4.7; Table 4.1). The speedups are relatively uniform, with the interquartile range of linear best-fit “speedup gradient” ranging from \(-2.0 \times 10^{-6}\) d\(^{-1}\) to \(-4.8 \times 10^{-6}\) d\(^{-1}\), with a median of \(-3.2 \times 10^{-6}\) d\(^{-1}\). This speedup gradient is the average strain rate in the enhanced summer sliding field; the median value corresponds to the speedup magnitude varying by -0.3 cm d\(^{-1}\) per kilometer of longitudinal distance. The negative magnitude indicates compression, implying that enhanced summer basal motion generally increases compression in the down-glacier reaches of these glaciers. The individual speedup data points are somewhat noisy, with a median root mean squared error of 0.048 m d\(^{-1}\) compared to their respective linear best fits. These relatively uniform speedups are superimposed on non-uniform winter velocity fields. The speedup magnitude does not appear to scale with total ice surface velocity; speedup magnitudes range from \(\sim 20 - 200\%\) of the wintertime velocity, with larger relative speedups for slower-moving portions of glaciers. Despite the general uniformity of speedup, several glaciers have relatively steep \((-1.8 \times 10^{-5}\) d\(^{-1}\); Figure S6) speedup gradients or shallower gradients that persist over long reaches \(\sim 30\) km; Figure 4.9), resulting in variable speedup across the profile.

On most glaciers, the speedup begins very near the glacier terminus (Figure 4.5c). However, on two extensively-debris covered glaciers, the summer speedup begins within the slow-moving debris-covered ice, approximately 5-10 km from the terminus/tributary junction. On one these glaciers, the speedup reaches 0.20 m d\(^{-1}\), extending through 10 km of ice moving \(\sim 0.07\) m d\(^{-1}\) in winter (Figure 4.8), with the surface strain rate 20-fold larger in summer than winter. Enhanced summer basal motion therefore accomplishes much of the ice discharge in these slow-moving reaches.

Six profiles appear to have steady velocity year-round (Table 4.1), or speed up less than our method is able to resolve given that glacier’s noise level. Five of these profiles sample quiescent surge-type glaciers, as evidenced by loop moraines.

Thirty-four profiles lack data of sufficient quality to conclusively determine whether the glacier speeds up in summer. These are typically slow-moving glaciers with velocities that
are nearly indistinguishable from measurement noise (Section 4.7.3). We exclude 11 profiles due to other glacier dynamics, such as active surge or iceberg calving, which obscure the melt-induced seasonal velocity signal (Figure 4.1c; Table 4.1).

4.3.2 Up-glacier limit to speedup extent

In elevation space, the speedup is relatively tightly constrained. Speedups terminate at a median elevation of 1800 m a.s.l., with an interquartile range of 1450 to 2400 m a.s.l. On many glaciers, this roughly corresponds to the elevation of the late summer snowline (Figure 4.6a), which, on the basis of manual inspection of Landsat 8 imagery, varies from 1800-2400 m (Figure 4.10). However, on several glaciers the speedup terminates many kilometers down-glacier of the snowline (Figure 4.6b).
Figure 4.5: Example data along a centerline profile on Kennicott Glacier (location shown in Figure 4.1c). a) Number of temporal observations at each point. b) Seasonal velocities in summer (red) and winter (blue). Each observation is shown as a circle. Seasonal median (heavy solid lines) and interquartile range (filled area) are shown. c) Summer speedup (positive where summer is faster than winter). Median speedup (circles) is calculated by subtracting the median winter speed from median summer speed (Section 4.2.2). Data are colored based on the rank sum test results (Section 4.2.2; panel e). Solid black line shows best linear fit to down-glacier speedup. d) Glacier surface elevation extracted from the National Elevation Dataset. e) Wilcoxon-Mann-Whitney test statistic for difference between up-glacier and down-glacier samples. The vertical dashed line shows the dividing point that statistically maximizes difference between up-glacier and down-glacier samples.
Figure 4.6: a) Inset showing locations of glaciers shown in b. b) False color composite (R,G,B: bands 7,5,3) Landsat 8 images showing speedup extent and snowline location on (left) Nabesna, (middle) Chisana, and (right) Root Glaciers. Images are overlain on Arctic DEM hillshades. c) Summer speedup length as a function of glacier length. The 1:1 line is dashed.
Figure 4.7: Longitudinal profiles of glacier speedup along 13 profiles with clear summer speedups. Points show median speedup. Solid lines show best linear fit to the speedup in the down-glacier reach. Bottom histogram shows frequency of speedup length in 3.5 km bins. Profile handles defined in Table 4.1. Sample speedup gradients are shown on right for reference.
Figure 4.8: Example data along a centerline profile on Barnard Glacier West Branch. a) Number of temporal observations at each point. b) Seasonal velocities in summer (red) and winter (blue). Each observation is shown as a circle. Seasonal median (heavy solid lines) and interquartile range (filled area) are shown. c) Summer speedup (positive where summer is faster than winter). Median speedup (circles) is calculated by subtracting the median winter speed from median summer speed. Maximum and minimum speedup bound the filled area. Maximum speedup is calculated as the 75th percentile summer speed minus the 25th percentile winter speed. Minimum speedup is calculated as the 25th percentile summer speed minus the 75th percentile winter speed. Vertical dotted line shows the speedup length as determined by automated rank sum test (panel e). Dashed vertical lines and colored data show start and end of summer speedup based on manual evaluation. Data are orange in the down-glacier portion that experiences summer speedup, and appear gray in the up-glacier portion that does not speed up in summer. Solid black line shows best linear fit to down-glacier speedup. d) Topographic profile from the National Elevation Dataset. e) Wilcoxon-Mann-Whitney test statistic for difference between up-glacier and down-glacier samples. The vertical dashed line shows the dividing point that statistically maximizes difference between up-glacier and down-glacier samples. This provides an objective measure of the up-glacier distance to which seasonal velocity fluctuations propagate.
Figure 4.9: Example data along a centerline profile on Nabesna Glacier West Branch. a) Number of temporal observations at each point. b) Seasonal velocities in summer (red) and winter (blue). Each observation is shown as a circle. Seasonal median (heavy solid lines) and interquartile range (filled area) are shown. c) Summer speedup (positive where summer is faster than winter). Median speedup (circles) is calculated by subtracting the median winter speed from median summer speed. Maximum and minimum speedup bound the filled area. Maximum speedup is calculated as the 75th percentile summer speed minus the 25th percentile winter speed. Minimum speedup is calculated as the 25th percentile summer speed minus the 75th percentile winter speed. Vertical dotted line shows the speedup length as determined by automated rank sum test (panel e). Dashed vertical lines and colored data show start and end of summer speedup based on manual evaluation. Data are orange in the down-glacier portion that experiences summer speedup, and appear gray in the up-glacier portion that does not speed up in summer. Solid black line shows best linear fit to down-glacier speedup. d) Topographic profile from the National Elevation Dataset. e) Wilcoxon-Mann-Whitney test statistic for difference between up-glacier and down-glacier samples. The vertical dashed line shows the dividing point that statistically maximizes difference between up-glacier and down-glacier samples. This provides an objective measure of the up-glacier distance to which seasonal velocity fluctuations propagate.
For many glaciers, the up-glacier limit of the summer speedup occurs at the base of an icefall (Figure 4.6b; Figure 4.11). Where icefalls exist, the late summer snowline often lies within the icefall (Figure 4.6b).

4.4 Discussion

4.4.1 Physical interpretation of speedup patterns

As our method requires averaging of velocities within seasons and across several years, these speedup profiles reflect the seasonally-integrated effect of enhanced basal motion. The times of our summer observations largely fall between expected spring speedup and fall slowdown events (Section 4.7.1, 4.7.3) and thus reflect average seasonal velocities rather than displacements associated with these shorter-lived events. The speedup profile at any given
Figure 4.11: Longitudinal elevation profiles along select glaciers. Portions below the summer speedup limit are colored red, whereas portions above are colored blue. Glaciers that do not show a seasonal speedup are shown in gray. Box and whisker plots on the plot margins show the median (yellow line), interquartile range (red box), and limits (caps) of the elevational (right axis) and longitudinal (top axis) distribution of the speedup lengths. The dashed line shows the median elevation to which speedups extend, with interquartile range shown as the filled area. Numbers indicate glacier profile number shown in Table 4.1.
time will likely be more complicated than the quasi-uniform season-averaged speedups we report here, and may partly explain the scatter about the linear best fit lines (Figure 4.7). For example, a constant magnitude speedup that propagated up-glacier as a wave [e.g., Colgan et al., 2012; Van De Wal et al., 2015] would appear to be a spatially uniform speedup if measured at a coarse temporal resolution (Section 4.7.3).

The uniformity of the observed speedup over large reaches of the ablation zone requires explanation. Seasonal variations in ice surface velocity are driven by non-steady basal motion induced by variations in subglacial water pressure. Water pressure histories in turn reflect the interplay between surface melt production and the ability of the subglacial hydrologic system to transmit that melt to the proglacial environment. One potential explanation for the uniformity of the speedup is that the magnitude and duration of enhanced summer basal motion could trade off in a manner that produces a relatively uniform summer speedup. For example, the down-glacier reach speeds up to a greater peak speed due to abundant meltwater availability, but basal motion wanes earlier in the year due to earlier establishment of efficient drainage [e.g., Kessler and Anderson, 2004]. Further up-glacier, a lower magnitude speedup due to less meltwater input and weaker surface-bed connectivity may persist for a greater fraction of the melt season. While it is tempting to invoke strong longitudinal stress-gradient coupling wherein speedups in one location pull and push those in other up- and down-glacier locations, we argue that this is not the case over such long distances in narrow valley glaciers, unless previous estimates of stress coupling length scales (4-10 ice thicknesses) [Kamb and Echelmeyer, 1986; Enderlin et al., 2016] are much too small.

The scaling of the speedup length with glacier length (Figure 4.6c) also merits discussion. The speedup length increases as a near-constant fraction (60-80%) of glacier length for glaciers < 40 km, and appears to reach a maximum of ~ 30 km for yet longer glaciers. The scaling may reflect the division of glacier accumulation and ablation areas. However, the snowline is 10-20 km up-glacier of the speedup limit on these four long profiles where the scaling breaks down, suggesting meltwater availability is not the sole prerequisite for basal motion. This could possibly be explained by weaker surface-bed hydraulic connections
through the presumably thick ice of these long low-gradient glaciers. The up-glacier limit of summer speedup on one of these glaciers (Figure 4.9) occurs at a large tributary junction and could therefore reflect a step increase in basal meltwater flux as these tributaries merge (Figure 4.6b).

Our data readily identify the order-of-magnitude velocity changes associated with the initiation and multi-annual evolution of active glacier surges. Between surges, however, quiescent surge-type glaciers display minimal seasonal velocity variation (Figure 4.5c). These findings are in line with those of Raymond and Harrisson [1988], who observed an order-of-magnitude increase in seasonal velocity variation during the surge of Variegated Glacier relative to its quiescent state. This suggests a fundamentally different coupling between meltwater inputs to the subglacial hydrologic system and basal motion may exist on quiescent surge type glaciers.

4.4.2 Implications for glacier modeling

The simple spatial geometry and relatively limited range of speedup magnitude we have documented suggests a simpler parameterization of enhanced summer basal motion. This could be particularly attractive for models of long-term glacier motion in which the very short time steps required to resolve subglacial hydraulics are certainly undesirable and often prohibitive. Simply applying a uniform seasonal speedup over the entire ablation area would capture the essence of the observed increase in ice flux on most of the glaciers we have studied. Further exploration of what sets the speedup magnitude, as well as what limits speedup length on long glaciers, is required before this approach could be applied with confidence.

The quasi-uniform speedup we observe scales with neither ice discharge, which reaches a maximum at the equilibrium line altitude (ELA) for a glacier without tributaries, nor winter surface velocity. Previous modeling studies have suggested that sliding may peak around the ELA [e.g., MacGregor et al., 2000; Anderson et al., 2006]; this does not appear to be the case.

It is difficult to find studies with output directly comparable to our seasonally-averaged
rate of basal motion. However, Beaud et al. [2014] modeled spatial patterns of glacier erosion with hydraulically-induced sliding and shear-stress dependent erosion rules and output a related metric. They predict quasi-uniform sliding over the terminal \(\sim 50\%\) of their modeled glacier, which appears to capture the essence of the summer enhanced sliding field, and hence implies that their model captures well the dynamics of the connection between glacier motion and hydraulics.

4.4.3 Feedback for the generation of icefalls

The summer speedup often extends to the base of an icefall (Figure 4.6b; Figure 4.11). Despite the difficulty of estimating velocities within icefalls (Section 4.7.3), we observe clear speedups below the icefall and not above. In addition, the seasonal velocity variation often declines toward the icefall over distances of several kilometers. Icefalls could affect the magnitude of seasonal speedup by: a) rising to elevations at which melt is no longer available, or b) changing the glacier dynamics in a such a way that the ice surface velocity responds insensitively to seasonal variations in subglacial water pressure, or c) disallowing water pressure variations over this timescale. As the icefalls span large ranges in elevation (up to \(\sim 1000\) m), the snowline often lies within the icefall (Figure 4.6b). Meltwater is therefore typically abundant below the icefalls, negligible above, with a rapid transition between. Several plausible conceptual models could explain minimal seasonal variation in velocity within an icefall (Section 4.7.4). Regardless of the physical mechanism, however, the observation of relatively uniform speedups below icefalls and minimal velocity variability above suggests that sliding and hence erosion of the glacier bed occurs below the icefall and not above, implying a positive feedback that could generate the icefalls. Assuming that glacier erosion is proportional to the rate of basal motion [Hallet, 1979; Iverson, 1991; Hallet, 1996], the speedup pattern would result in quasi-uniform erosion below the icefall and near-zero erosion above. This erosion pattern would increase the relief of an icefall once it begins to form, and could explain their widespread distribution without appealing to breaks in lithology. The seasonal speedup below icefalls also reduces compression and could partly account for the formation of wave ogives.
4.5 Conclusions

We document spatial patterns of seasonal glacier surface velocity variability, which is the surface expression of hydraulically-induced enhanced glacier basal motion. Across the subset of glaciers that show pronounced summer speedups, we find a relatively narrow range of summer speedup magnitude (median = 11 cm d\(^{-1}\)) that is relatively uniform (median speedup gradient = \(-3 \times 10^{-6} \text{ d}^{-1} = 0.3 \text{ cm d}^{-1} \text{ km}^{-1}\)) across the lower 60-80% (9-32 km) of a single glacier longitudinal profile. Over these large length scales, ice surface speedup likely reflects local basal motion rather than longitudinal stress-gradient ice coupling. This relatively simple spatial pattern and narrow range of speedup magnitude may allow simple parameterization of enhanced summer basal motion for glacier models in which the short time steps required for resolving subglacial hydrology are not feasible. The speedup does not scale with ice surface velocity, resulting in large fractional summer velocity increases in regions of slow-moving ice, such as debris-covered glacier tongues. Several quiescent surge-type glaciers appear not to speed up in summer, suggesting a fundamental difference in the coupling of seasonal meltwater inputs and glacier basal motion on these glacier types. The up-glacier limit of the summer speedup is often encountered at the base of an icefall, which suggests a feedback for icefall generation over millennia of glacier erosion.

4.6 Acknowledgements

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4.7 Supplements

4.7.1 Image cross-correlation, swath profiling, and velocity filtering

Cross-correlation of optical satellite imagery [Scambos et al., 1992] is often better suited than satellite-borne radar data for velocity estimation on alpine glaciers because it is less prone to decorrelation on the steep, rough, rapidly evolving surfaces often found on valley glaciers [Joughin et al., 2010b]. Despite these limitations, pixel-offset tracking (also called feature, speckle, or pixel tracking) of radar imagery has yielded valuable insight into Alaska glacier surge dynamics [Abe and Furuya, 2015] and interannual velocity variability [Burgess et al., 2013a; Waechter et al., 2015].

The image correlation nomenclature used here is consistent with that of Fahnstock et al. [2016]. We refer to the image sub-scene that is searched for as the template chip and the sub-scene in which it is sought as the source chip. We use a template chip size of 24 pixels (360 m), a source chip size of 40 pixels (600 m), with a posting of 10 pixels (150 m). Using this parameter set, PyCorr processes an entire Landsat scene in less than 10 minutes.

We apply swath profiling to reduce measurement noise. In swath profiling, the down-glacier transect is interpolated to an even spacing to generate centroid coordinates. Rectangular boxes of user-specified down-glacier and cross-glacier extent are generated for each centroid point. A raster is then sampled within the extent of these rectangles, from which summary statistics are calculated. This process is repeated for all velocity rasters that intersect the profile to produce a time series of glacier velocity at each rectangle centroid. We calculate the median velocity and interquartile range over rectangles measuring 500 (1000) m in the longitudinal (transverse) directions, which corresponds to \( \sim 22 \) velocity measurements.

The majority of our summer imagery is acquired between DOYs 175-210 (June 24-July 29; Figure 4.2), whereas our winter imagery spans DOYs 25-100 (January 25-April 10). Thus, our data unevenly sample the annual glacier velocity cycle and weight summer observations
toward times when glacier velocity is likely to be high. The few velocity fields in late summer, a time when velocity is likely to be slowest, do not sufficiently lower the average speeds to produce negative speedup magnitudes (i.e., summer slower than winter) on any profile. We exclude most springtime imagery due to poor correlations on rapidly changing snow cover (Section 4.7.3) and thus likely miss at least the peaks of the spring speedup event on many glaciers; in this range these typically occur in late May to early June [e.g., Anderson et al., 2004; MacGregor et al., 2005; Bartholomaus et al., 2011; Darling, 2012]. Therefore, our observations mainly reflect average seasonal velocities rather than displacements associated with dramatic, short-lived events.

The maximum speedup profile is calculated by subtracting the 25th percentile winter velocity from the 75th percentile summer velocity, and vice-versa for the minimum speedup. The interquartile range characterizes within-season velocity variability as well as measurement uncertainty.

Spurious velocities are found where the template chip is only weakly or ambiguously correlated to any feature within the source chip, or where chips are highly correlated but apparent displacement is related to features whose motion does not reflect ice surface motion. These non-physical velocities can be removed by filtering based on cross-correlation quality and by using information on spatial coherence provided by the swath profiler. We remove both pixels with weak correlation peaks and pixels with multiple correlation peaks of near equal magnitude. We find using a threshold correlation strength ($corr$) of 0.20 and threshold correlation strength difference between the two strongest peaks ($delcorr$) of 0.05 are sufficient to remove spurious data. In addition, we filter data on output from the swath profiling routine. The swath profiler-based approach filters based upon the spatial coherence of the output. We remove any median velocity within a swath profiler box if the box contains fewer than 10 pixels contributing to that median, or if the standard deviation of pixel velocity is greater than 1 m d$^{-1}$. Over the small area of a swath profiler box, such large variance is typically due to spurious data, which is often associated with a small number of valid pixels. Further, we employ an iterative approach to filter based upon the temporal variation. This
approach incorporates all velocity data (after filtering based on correlation and swath profiler output) to define a temporal mean ($\mu$) and standard deviation ($\sigma$) for each point along a longitudinal profile. Velocity data outside $\mu \pm 2\sigma$ are removed, and the process is repeated twice. Data lying outside $\mu \pm 2\sigma$ before these two iterations are due to spurious correlations and are not caused by physical temporal variability in glacier motion. However, the method does select for “sameness” and we therefore may reduce the magnitude of seasonal speedup as a result. On average, after filtering, we incorporate $\sim 10$ velocity observations at each point along a glacier centerline profile in our computation of average seasonal velocity profiles (Figure 4.12).
4.7.2 Speedup length determination

The rank sum test evaluates whether two samples are drawn from the same population and is a nonparametric equivalent to the t-test. Using a nonparametric test is advantageous because it relies on fewer assumptions about data distribution and independence, although it is less powerful than the t-test in detecting a shift if the data are indeed independent and identically distributed [Helsel and Hirsch, 2002]. The test statistic \(W\), is given by

\[
W = \sum_{i=1}^{n} R_i \tag{4.1}
\]

where \(n\) is the number of observations in the smaller sample, and \(R_i\) is the rank of the \(i\)-th observation. A rank of 1 represents the largest value and \(N = n + m\) represents the smallest value, where \(m\) is the length of the larger sample [Helsel and Hirsch, 2002]. To identify the transition between up-glacier and down-glacier groups, we iteratively change \(n\) between the bounds of 2 and \(N - 1\), and identify the speedup length as the distance associated with the value of \(n\) with the largest absolute value of \(W\), which is the dividing point that maximizes difference between the up-glacier and down-glacier samples.

4.7.3 Data limitations

Clouds, and changing snow cover and shadow distributions reduce the useful portions of images. This can introduce temporal sampling bias in our data. Autumn is particularly cloudy in WRST and results in poor sampling of time periods when glacier velocity may be slowest. Changing surface appearance as the snowline retreats up-glacier in spring and early summer can also generate poor correlation at the time of year when basal motion is likely to be highest. The seasonally-integrated and multiyear averaged nature of the velocity profiles reduces noise but also effectively decreases the temporal resolution of our observations. We therefore cannot interpret glacier dynamics on sub-seasonal scales, such as pulses of basal motion that propagate as waves. In addition, we are unable to sample small glaciers, which commonly move too little on 48 day timescales to escape our measurement error. Glaciers must have low signal-to-noise ratios to confidently detect seasonal velocity fluctuations. For
this reason, glaciers that do not appear to speed up in summer may just have higher noise levels that mask systematic seasonal velocity variability.

East-west flowing glaciers tend to have higher noise levels because topographic shadowing strongly varies throughout the year, resulting in a greater frequency of spurious correlations. In addition, debris-covered termini often result in high noise levels, likely because complicated topography, ice cliff backwearing, and supraglacial lake dynamics cause low correlation strength or well-correlated chips that do not reflect ice advection.

Icefalls present a distinct challenge for velocity estimation from cross-correlation \( [N. \text{Anderson et al., in prep; Altena et al., in prep}] \). Rough, disorganized ice texture and persistent surface features that are linked to bedrock geometry and do not translate with ice motion result in spurious near-zero velocities. Velocity estimates through icefalls tend to be more coherent in winter than in summer. Happily, while this limitation prevents velocity estimates in the steepest portions of icefalls, we are able to characterize velocity immediately above and below these zones. Finally, we note that, from these data of seasonal variations in ice surface velocity, we can resolve only the relative variations in basal motion. Without borehole deformation measurements or detailed knowledge of ice thickness distributions, we are unable to resolve any “background” year-round basal motion, but acknowledge that it likely exists in places.

4.7.4 Conceptual model of glacier sliding through an icefall

There are several plausible physical mechanisms for why basal motion could be altered by the presence of an icefall. In one conceptualization, the fast rates of ice deformation and regelation sliding through the icefall associated with their steep slopes could inhibit the formation of efficient subglacial drainage, as in glacial surges \( [Kamb \text{ et al., 1985}] \). Conversely, the steep surface slope would result in steep hydraulic potential gradients \( [Shreve, 1972] \), which encourages the growth of efficient subglacial drainage \( [Rothlisberger, 1972] \). This may in turn result in the year-round maintenance of efficient subglacial drainage beneath icefalls. Finally, ice motion through the icefall may be rapid enough that glacier motion is limited not by local basal shear stress but by stress-gradient coupling with up- and down-glacier ice.
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