1  **Observations of seasonal and diurnal glacier velocities at Mount**

2  **Rainier, Washington using terrestrial radar interferometry**

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9  **Abstract**

10  We present surface velocity maps derived from repeat terrestrial radar interferometry (TRI)

11  measurements, and use these time series to examine seasonal and diurnal dynamics of alpine

12  glaciers at Mount Rainier, Washington. We show that the Nisqually and Emmons glaciers have

13  small slope-parallel velocities near the summit (<0.2 m/day), high velocities over their upper and

14  central regions (1.0-1.5 m/day), and stagnant debris-covered regions near the terminus (<0.05

15  m/day). Velocity uncertainties are as low as ±0.02-0.08 m/day. We document a large seasonal

16  velocity decrease of 0.2-0.7 m/day (-25 to -50%) from July to November for most of the

17  Nisqually glacier, excluding the icefall, suggesting significant seasonal subglacial water storage

18  under most of the glacier. We did not detect diurnal variability above the noise level. Simple 2D

19  ice flow modeling using TRI velocities suggests that sliding accounts for 91% and 99% of the

20  July velocity field for the Emmons and Nisqually glaciers with possible ranges of 60 - 97% and

21  93 - 99.5%, respectively, when considering model uncertainty. We validate our observations

22  against recent in situ velocity measurements and examine the long-term evolution of Nisqually

23  glacier dynamics through comparisons with historical velocity data. This study shows that repeat

24  TRI measurements with >10 km range can be used to investigate spatial and temporal variability

25  of alpine glacier dynamics over large areas, including hazardous and inaccessible areas.

## 1 Introduction

26

27  Direct observations of alpine glacier velocity can help improve our understanding of ice

28  dynamics. Alpine glacier surface velocities are typically dominated by basal sliding, which is

29  tightly coupled to subglacial hydrology (Anderson et al., 2014; Bartholomaus et al. 2008).

30  However, the spatial extent and spatial/temporal resolution of direct velocity measurements are

31  often limited to short campaigns with sparse point measurements in accessible regions (e.g.

32  Hodge, 1974; Driedger and Kennard, 1986).  Remote sensing can help overcome many of these

33  limitations. Radar interferometry, a form of active remote sensing, detects mm- to cm- scale

34  displacements between successive images of the same scene and can see through clouds and fog.

35  In the past few decades, satellite interferometric synthetic aperture radar, or InSAR (e.g.

36  Massonnet and Feigl, 1998; Burgmann et al., 2000) has emerged as an invaluable tool for

37  quantifying glacier dynamics (e.g., Joughin et al, 2010). However, limited data availability and

38  revisit times limit the application of InSAR for the study of many short-term processes.

39  Terrestrial radar interferometry (TRI), also referred to as ground-based radar interferometry, has

40  recently emerged as a powerful technique for observing glacier displacement that is not prone to

41  the same limitations (Caduff et al., 2014). Sets of radar data acquired at intervals as short as ~1

42  minute from up to several km away allow for observations of velocity changes over short

43  timescales and large spatial extents. Stacking these large numbers of interferogram pairs over

44  longer timescales can significantly reduce noise. Here, we employ this relatively new technique

45  to provide spatially- and temporally-continuous surface velocity observations for several glaciers

46  at Mount Rainier volcano in Washington State (Fig. 1).  Though Rainier's glaciers are among the

47  best-studied alpine glaciers in the U.S. (Heliker et al., 1984; Nylen, 2004), there are many open

48  questions about diurnal and seasonal dynamics that TRI can help address.  Specifically, many

49  aspects of subglacial hydrology and its effects on basal sliding are poorly constrained, especially

50  for inaccessible locations like the Nisqually icefall and ice cliff. Our observations provide new

51  insight into these processes through analysis of the relative magnitude and spatial patterns of

52  surface velocity over diurnal and seasonal timescales.  To our knowledge, no other studies have

53  investigated seasonal changes to glacier dynamics using TRI.

54  Mount Rainier offers an excellent setting for TRI, with several accessible viewpoints offering a

55  near-continuous view with ideal line-of-sight vectors for multiple glaciers, and well-distributed

56    static bedrock exposures for calibration.  The ability to image the velocity field of entire glaciers

57    from one viewpoint with minimal shadowing sets this study area apart. Most previous studies

58    only image part of the glaciers under investigation, usually due to less favorable viewing

59    geometries (e.g. Noferini et al., 2009; Voytenko et al., 2015; Riesen et al., 2011). However, the

60    steep topography and local climatic factors at Mount Rainier result in strong atmospheric

61    variability and turbulence – a major source of noise for radar interferometry techniques

62    (Goldstein, 1995). Atmospheric noise is a particular issue for the long ranges (>10 km)

63    associated with accessible viewpoints at Mount Rainier. To overcome this limitation, we

64    successfully combine, expand on, and evaluate noise reduction techniques such as stacking

65    interferograms (e.g. Voytenko et al. 2015) and deriving atmospheric noise corrections over static

66    control surfaces (bedrock exposures) (e.g. Noferini et al. 2009). We demonstrate that these

67    techniques offer significant uncertainty reductions using a novel bootstrapping approach.

68    In the following sections, we provide background on Mount Rainier's glaciers, and detail our

69    sampling methodology and data processing techniques.  We then present TRI results

70    documenting seasonal and diurnal velocity variations for the Nisqually, Wilson, and Emmons

71    Glaciers, and quantify measurement uncertainty. Next we examine the partitioning of observed

72    surface velocities between deformation and basal sliding at different times of year using a simple

73    2D flow model, and compare our observations to other recent and historical velocity

74    measurements.  These comparisons provide ground truth for TRI measurements and new insight

75    into the evolution of the Nisqually glacier since the late 1960s.

76    **2   Study area**

77    With a summit elevation of 4392 m, Mount Rainier (Fig. 1) is the largest stratovolcano in the

78    Cascades and is considered the most dangerous volcano in the United States (Swanson et al.,

79    1992). It also holds the largest concentration of glacial ice in the mainland United States

80    (Driedger and Kennard, 1986) - 87 km$^2$ was covered with perennial snow and ice in 2008 (Sisson

81    et al., 2011). The steep upper sections of the major glaciers are relatively thin, with typical

82    thicknesses of ~30 to 80 m (Driedger and Kennard, 1986). Thickness increases at lower

83    elevations, with a maximum of ~200 m for the Carbon glacier, although these estimates likely

84    provide an upper bound, as these glaciers have experienced significant thinning in recent

85    decades, losing 14% of their volume between 1970 and 2008 (Sisson et al., 2011).  Mass balance

86    stake measurements from 2003-2010 show that the average winter balance for Nisqually was 2.4

87    m water equivalent (m.w.e.), average summer balance was -3.5 m.w.e., and cumulative net

88    balance was -8.6 m.w.e. from 2003-2011 (Riedel, 2010; Riedel and Larrabee, 2015).

89    The glaciers of Mt. Rainier have been of interest to geoscientists for over 150 years and have a

90    long record of scientific observation (Heliker, 1984). In this study, we focus on large, accessible,

91    well-documented glaciers in the park: the Nisqually and Wilson glaciers on the southern flank,

92    and Emmons and Upper Winthrop glaciers on the northeastern flank (Fig. 1).

93    The Nisqually Glacier is visible from several viewpoints near the Paradise Visitor Center, which

94    is accessible year-round. The terminus location has been measured annually since 1918, and

95    three transverse surface elevation profiles have been measured nearly every year since 1931

96    (Heliker, 1984). Veatch (1969) documented a 24-year history of Nisqually's advances and

97    retreats and other dynamic changes through a meticulous photographic survey from 1941-1965.

98    Hodge (1974) conducted a detailed 2-year field study of the seasonal velocity cycle for the lower

99    Nisqually. He found that velocities varied seasonally by about 50%, with maximum velocities in

100    the spring (June) and minimum in the fall (November). This finding, and the lack of correlation

101    between runoff and sliding speeds, advanced the idea that efficient conduits close as meltwater

102    input decreases in the fall, leading to distributed subglacial storage through the fall, winter and

103    spring. Increased surface melting in spring and summer leads to increased subglacial discharge

104    and the opening of a more efficient network of conduits capable of releasing some of this stored

105    water (Hodge, 1974). More recently, Walkup et al. (2013) tracked the movements of supraglacial

106    rocks with high precision from 2011-2012, yielding velocity vectors for a wide network of points

107    over the lower parts of Nisqually glacier.

108    The Emmons glacier, visible from the Sunrise Visitors Center, has received less attention than

109    Nisqually, despite the fact that it is the largest glacier by area on the mountain (Driedger and

110    Kennard, 1986), mainly because it is not as easily accessible as Nisqually. A large rock fall

111    ($\sim 1.1 \times 10^7 \, m^3$) from Little Tahoma in December 1963 covered much of the lower Emmons

112    glacier with a thick debris layer (Crandell and Fahnestock, 1965).  The insulating debris cover

113    likely contributed to the advance and thickening of the Emmons Glacier from 1970-2008, while

114    all other glaciers on Mount Rainier experienced significant thinning (Sisson et al., 2011).

115    Average 2003-2010 winter balance for Emmons was 2.3 m.w.e., average summer balance was -

116  3.2 m.w.e , and cumulative net balance was -7.7 m.w.e. from 2003-2011 (Riedel, 2010; Riedel

117  and Larrabee, 2015).

118  The National Park Service's long-term monitoring protocols include both the Nisqually and

119  Emmons glaciers and involve regular photographs, annual mass balance measurements,

120  meltwater discharge rates, plus area and volume change estimates every decade (Riedel, 2010;

121  Riedel and Larrabee, 2015).

## 3   Methods

### 3.1 Instrument description

124  For this study, we used a GAMMA portable radar interferometer (GPRI) (Werner et al., 2008;

125  Werner et al., 2012) - a ground-based, frequency-modulated continuous waveform (FMCW)

126  radar that can capture mm-scale surface displacements. The instrument includes three 2-m

127  antennas mounted on a vertical truss, with one transmit antenna 35 cm above the upper of two

128  receiving antennas, spaced 25 cm apart (Fig. 2). The transmit antenna produces a 35° vertical

129  beam with 0.4° width that azimuthally sweeps across the scene to build a 2D radar image as the

130  truss rotates.  The radar operates at a center frequency of 17.2 GHz, with selectable chirp length

131  of 2-8 ms and bandwidth of 25 to 200 MHz. The radar wavelength is 17.6 mm with range

132  resolution of ~0.75 cm and one-way interferometric change sensitivity of 8.7 mm/cycle of phase

133  providing <1 mm line-of-sight precision.  Line-of-sight interferograms are generated by

134  comparing phase differences in successive acquisitions from the same viewpoint. The interval

135  between acquisitions can be as short as ~1 min, allowing for high coherence even in rapidly

136  changing scenes.

### 3.2     Survey Description

138  We performed four data collection campaigns in 2012 (Table 1). The first campaign occurred on

139  6-7 July 2012. This timing corresponds to just after the expected peak seasonal glacier velocities

140  at Mount Rainier (Hodge, 1974).  Following the success of this study, three subsequent

141  deployments were performed during the late fall and early winter, which should capture near-

142  minimum seasonal velocity (Hodge, 1974).  These campaigns were timed to occur before,

143  immediately after, and a few weeks after the first heavy snowfall of the season (2 Nov 2012, 27

144  Nov 2012 and 10 Dec 2012, respectively).

145    Three viewpoints were selected for data collection: GLPEEK and ROI, which overlook the

146    Nisqually and Wilson glaciers, and SUNRIZ, which overlooks the Emmons and upper Winthrop

147    glaciers (Fig. 1). ROI and SUNRIZ were directly accessible from park roads, which greatly

148    facilitated instrument deployment, and GLPEEK was accessed on foot. ROI was occupied

149    during all campaigns, while SUNRIZ and GLPEEK were only occupied during the July 2012

150    campaign because of access limitations. Figures A1-A3 show the field of view from each

151    viewpoint.

152    Distances from the GPRI to the summit were 6.7, 7.6, and 10.8 km from GLPEEK, ROI and

153    SUNRIZ, respectively. Radar images were continuously collected with a 3-minute interval for all

154    surveys. Total acquisition time at each site was dictated by logistics (weather conditions,

155    personnel), with ~24 hour acquisitions at SUNRIZ and ROI to capture diurnal variability.

156    The instrument was deployed on packed snow during the 6 July 2012 GLPEEK and 27 Nov and

157    10 Dec 2012 ROI acquisitions. Over the course of the GLPEEK survey, we noted limited snow

158    compaction and melt beneath the GPRI tripod with total displacement of ~2-4 cm over ~6 hours.

159    However, this instrument motion proved to be negligible for the interferogram interval used (6

160    min). We did not note significant snow compaction under the tripod during the fall/winter

161    surveys.

162    Weather conditions during the July 2012 surveys were clear with light/variable wind. The 2 Nov

163    2012 survey involved high-altitude clouds, passing showers and brief interruptions in data

164    collection. Weather conditions were clear with sun for the 27 Nov 2012 campaign, and fog with

165    limited visibility on 10 Dec 2012.

166    **3.3    Data Processing**

167    All radar data were processed with the GAMMA SAR and Interferometry software suite.

168    Interferograms were generated from single-look complex SLC products with a time separation of

169    6 minutes, though sometimes longer if acquisition was interrupted. For example images see Fig.

170    A4. Interferograms were multi-looked by 15 samples in the range direction to reduce noise. A

171    correlation threshold filter of 0.7 and an adaptive bandpass filter (ADF) with default GAMMA

172    parameters were applied to the interferograms to improve phase unwrapping. Phase unwrapping

173    was initiated in areas with high correlation scores and negligible deformation, such as exposed

174    bedrock or stagnant ice.

### 3.3.1 Atmospheric noise corrections

175

176    Slight changes in the dielectric properties of the atmosphere between the GPRI and target

177    surfaces can lead to uncertainty in the interferometric displacement measurements (Zebker et al.

178    1997; Werner et al., 2008).  Changes in atmospheric humidity, temperature, and pressure can all

179    affect radar propagation velocity (Goldstein, 1995).  These variations are manifested as phase

180    offsets in the received radar signal, which must be isolated from phase offsets related to true

181    surface displacements.

182    This atmospheric noise proved to be significant for the long range (i.e., ~22 km two-way

183    horizontal path at SUNRIZ), mountainous terrain (i.e., ~2.4 km vertical path from SUNRIZ to

184    summit), and turbulent atmosphere involved with this study, with the magnitude of this noise

185    often exceeding that of surface displacement signals.  The scale of the atmospheric noise features

186    we observed in the data was typically much wider than the width of the glaciers, so in order to

187    minimize this atmospheric noise in the individual interferograms, we interpolated apparent

188    displacement values over static control surfaces (e.g. exposed bedrock). To do this, we fit a

189    surface using Delaney triangulation to a subset (5%) of pixels over exposed bedrock. The subset

190    of pixels was resampled randomly for each unwrapped interferogram and the interpolated result

191    was smoothed to reduce artifacts and then subtracted from the interferogram. The corrections

192    were applied to all individual interferograms, and the resulting products were stacked to further

193    reduce noise. To stack, we took all the images for a given time period and computed the mean

194    and median at each pixel. This has the effect of augmenting signal and canceling out noise. The

195    median is less affected by outliers and is our preferred result. The median line-of-sight (LOS)

196    velocities from this stack provide a single measurement with a high signal to noise ratio for the

197    entire sampling period.

198    In addition to computing the median LOS velocities for the entirety of each sampling period, we

199    also computed a running mean of the LOS velocities to characterize any short-term velocity

200    variations in the extended occupation datasets: 7-8 July SUNRIZ (24 hours) and 1-2 November

201    ROI (21 hours).  The running mean was computed every 0.3 hours with a 2-hour centered

202    (acausal) window, with standard error used to estimate uncertainty.

203  Interferograms with significant phase unwrapping errors, low correlation, or anomalous noise

204  were excluded from stacking. We only excluded a few images for each site with the exception of

205  SUNRIZ, which produced many images with anomalous noise and unwrapping errors, possibly

206  due to instrument noise and/or the extended range through significant atmospheric disturbance.

207  For this reason, more than half of the data from SUNRIZ was excluded from the analysis (Table

208  2). For GLPEEK and ROI, interferograms with occasional localized unwrapping errors were

209  preserved during stacking, as they have little influence on the final stack median. However,

210  localized areas with persistent unwrapping errors in the SUNRIZ data were masked using a

211  threshold standard deviation filter of 0.6 m/day.

212  We estimated median LOS velocity uncertainties using a bootstrapping approach (Efron, 1979).

213  This involved resampling the set of images used in the stack with replacement 1000 times for

214  each campaign. Then, for each pixel, the 25th and the 975th ordered values were set as the lower

215  and upper bounds of the 95% confidence interval.

### 3.3.2 Conversion from radar coordinates to map coordinates

217  We developed a sensor model and tools to terrain-correct the stacked GPRI data (in original

218  azimuth, range coordinates) using an existing 2 m/pixel airborne LiDAR digital elevation model

219  (DEM) acquired in September 2007/2008 (Robinson et al., 2010).  While some elevation change

220  has undoubtedly occurred for glacier surfaces between September 2008 and July 2012, the

221  magnitude of these changes (<20 m) is negligible for orthorectification purposes given the GPRI

222  acquisition geometry. A single control point identified over exposed bedrock in the LiDAR DEM

223  and the multi-look image (mli) radar data was used to constrain absolute azimuth orientation

224  information for each campaign.  A ~10 m/pixel (mean of azimuth and range sample size) grid in

225  UTM 10N (EPSG: 32610) was created for each campaign, with extent computed from the GPRI

226  GPS coordinates, min/max range values, and min/max absolute azimuth values.  Each 3D pixel

227  in this grid was then populated by extracting the radar sample with corresponding range and

228  azimuth.

### 3.3.3 Correction to slope-parallel velocities

230  While the line-of-sight vectors for these surveys are roughly aligned with surface displacement

231  vectors (median incidence angles for glacier surfaces are ~22° for GLPEEK, ~25° for SUNRIZ

232  and ~26° for ROI), glaciological analyses typically require horizontal and vertical velocity

233  components relative to the glacier surface.  As each GPRI survey offers only a single look

234  direction, this is not possible.  However, we can assume that displacement is dominated by

235  surface-parallel flow, and use the 2007/2008 LiDAR DEM to extract surface slopes needed to

236  estimate 3D displacement vectors (e.g., Joughin et al., 1998).

237  This approach is intended for relatively smooth, continuous surface slopes over length scales >2-

238  3x ice thickness.  It is therefore possible that the slope-parallel correction can overestimate

239  velocity for steep, high relief surfaces with significant high-frequency topographic variability

240  (e.g. icefalls).  The slope-parallel assumption also begins to break down where the vertical flow

241  velocity component becomes significant.  This is expected in the upper accumulation and lower

242  ablation zones, where the submergence and emergence velocities become more significant,

243  respectively, but is less important near the equilibrium-line altitude (ELA) or locations where

244  sliding dominates surface motion.  The latter is expected for much of the Nisqually Glacier at

245  least (Hodge, 1974).

246  We implement a slope-parallel correction by first downsampling the 2007/2008 LiDAR DEM to

247  20 m/pixel and smoothing with a 15x15 pixel (~300 m), 5-sigma Gaussian filter.  The slope-

248  parallel velocity ($V_{sp}$) is defined as:

249  $$V_{sp} = V_{LOS}/(\hat{S} \cdot \hat{L}) \qquad\qquad (1)$$

250  where $\hat{S} \cdot \hat{L}$ is the dot product between the unit vector pointing directly downslope from each

251  grid cell ($\hat{S}$) and the unit vector pointing from each grid cell to the sensor ($\hat{L}$).  Regions where

252  the angle between these two vectors exceeded 80° were masked to avoid dividing by numbers

253  close to zero which could amplify noise.

## 3.4   2-D glacier deformation modeling

255  Surface flow velocity can be partitioned into internal deformation and basal sliding components.

256  We present a simple, 2-D plane-strain ice deformation model for a preliminary assessment of the

257  importance of basal sliding for the glaciers in our study area. The deformation model uses the

258  shallow ice approximation (SIA) – an approximate solution of the Stokes Equations (Greve and

259  Blatter, 2009; Cuffey and Paterson, 2010). The expected surface velocity $u_s$ due to internal

260  deformation from the SIA model is:

261 $$u_s = \frac{2\,A(sin(\alpha)\rho_i g)^n\,H^{n+1}}{n+1} \tag{2}$$

262 where $\rho_i$ represents ice density, $g$ represents gravitational acceleration, $\alpha$ represents local surface

263 slope, $H$ represents local ice thickness, $A$ represents an ice softness parameter and $n$ represents a

264 flow rate exponent. The coordinate system is vertically aligned.

265 The SIA is not well-suited for narrow mountain glaciers, so we modify it to simulate the effect of

266 non-local conditions, such as lateral sidewall drag and longitudinal stretching. The ice thickness

267 $H$ and surface slope $\alpha$ are smoothed using a weighting function based on Kamb and Echelmeyer

268 (1986). Kamb and Echelmeyer (1986) calculated a longitudinal coupling length $l$ using a 1-D

269 force balance approach, for each point in their domain. They calculated $l$ to be in the range of

270 one-to-three ice thicknesses for valley glaciers. We simplified this by using a single value for $l$

271 over the domain of model. The longitudinal couple length $l$ is used in a weighting function to

272 smooth $\alpha$ and $H$. The weighting function has the form:

273 $$W(x,y) = e^{-\frac{\sqrt{(x-x')^2+(y-y')^2}}{l}} \tag{3}$$

274 where $x$ and $y$ represent the horizontal coordinates of the weight position, and $x'$ and $y'$ represent

275 the horizontal coordinates of the reference position. Weights are calculated at each point in the

276 model domain, over a square reference window (side length of $A_w$). $H$ and $\alpha$ are smoothed at the

277 reference position by normalizing weights over the reference window. We choose a coupling

278 length $l$ of ~1.5 ice thickness and an averaging window size of ~3 ice thicknesses, consistent

279 with the usage in Kamb and Echelmeyer (1986). We use a spatially uniform and temporally

280 constant ice softness parameter suitable for ice at the pressure melting point of $2.4 \times 10^{-24}$ Pa$^{-3}$ s$^{-1}$

281 (Cuffey and Paterson, 2010, pg. 75). Ice softness can be affected by several factors (e.g.,

282 englacial water content and impurities), so we also consider an ice softness parameter up to twice

283 this best estimate in accounting for model uncertainties, as described below. Our best estimates

284 of model input parameters are summarized in Table 3.

285 Surface slope (Fig. A1B) was estimated from the 2007/2008 LiDAR DEM (Robinson et al.

286 2010). Surface velocities $u_s$ are the TRI-derived median slope-parallel velocities. Ice thicknesses

287 $H$ (Fig. A1A) were estimated by differencing the 2007/2008 LiDAR DEM surface elevations and

288 the digitized and interpolated bed topography from Driedger and Kennard (1986). The Driedger

289  and Kennard (1986) bed topography contours were derived from ice-penetrating radar point
290  measurements and surface contours from aerial photographs.  The published basal contours for
291  Nisqually/Wilson, Emmons, and Winthrop Glaciers were digitized and interpolated to produce a
292  gridded bed surface using the ArcGIS Topo to Raster utility.  The gridded bed elevations have
293  root mean squared error (RMSE) of 11 m when compared with the 57 original radar point
294  measurements.  A point-to-plane iterative closest point algorithm (implemented in the NASA
295  Ames Stereo Pipeline pc_align utility (Shean et al., 2015)) was used to coregister the 1986 bed
296  topography to the 2007/2008 LiDAR topography over exposed bedrock on valley walls.  Mean
297  error over these surfaces was 7.6 m following coregistration, although some of this error can be
298  attributed to actual surface evolution near glacier margins (e.g., hillslope processes) from 1986-
299  2008. In addition to these interpolation and coregistration errors, there were likely small changes
300  in ice thickness during the 4-5 years between the 2007/2008 DEM data collection and the 2012
301  TRI observations, as mass balance measurements suggest that both the Nisqually and Emmons
302  Glaciers experienced net mass loss during this time period (Riedel and Larrabee, 2015).
303  Propagation of these uncertainties results in estimated ice thickness uncertainties of ~5-25%. In
304  order to account for this large uncertainty, we ran the model with ±25% ice thickness as well as
305  2x ice softness in order to estimate the possible range of expected deformation velocities.

306  More sophisticated ice flow models (e.g. Gagliardini et al., 2013; Le Meur et al., 2004; Zwinger
307  et al., 2007) could potentially offer a more realistic picture of the spatial and temporal variability
308  of glacier sliding. However, given the poorly-constrained model inputs and observational
309  emphasis for this study, we proceed with the SIA model to obtain approximate estimates for the
310  deformation and sliding components of observed velocities.

## 4   Results

312  The median stacks of surface-parallel velocity for all viewpoints and their respective uncertainty
313  estimates are shown in Fig. 3-6. Overall, our results show that repeat TRI measurements can be
314  used to document spatial and temporal variability of alpine glacier dynamics over large areas
315  from >10 km away. The atmospheric noise removal approach was successful in extracting a
316  glacier displacement signal for all campaigns, with excellent results for Nisqually Glacier due to
317  the shorter range from ROI and GLPEEK viewpoints and limited glacier width between control
318  surfaces. Stacking alone was very effective; the velocities of the mean and median stacks with

319  and without the atmospheric noise correction were very similar. The main benefit of the extra

320  step of using stable rock points to subtract an estimate of the atmospheric noise was to

321  significantly reduce the uncertainties and to reduce the noise where velocities are slow. The

322  uncertainties before and after atmospheric correction are compared on Table 2. The median

323  width of the 95% confidence interval for each corrected, stacked pixel is plotted in Fig. 3B and

324  Fig. 5. Note near-zero values over exposed bedrock surfaces used to derive atmospheric noise

325  correction. We were able to reduce uncertainties (half the median confidence interval width) to

326  about ±0.02 to ±0.08 m/day over glacier surfaces for some campaigns, with uncertainty

327  dependent on the total number of stacked images, weather conditions, and target range (Table 2).

328  For example, the 6 July 2012 ROI survey had a final confidence interval width of 0.11 m/day

329  (~±0.06 m/day) while the 10 Dec 2012 ROI survey had a final confidence interval width of 0.15

330  m/day (~±0.08 m/day) despite a 50% increase in stack count. This is likely due to increased local

331  atmospheric variability, as low-altitude clouds obscured the surface during 10 Dec 2012 survey.

332  The 2 Nov 2012 ROI survey had the highest stack count (359) with the lowest uncertainty values

333  of ±0.02 m/day (Table 2).

334  **4.1 July 2012 Surface Velocities**

335  The 6-7 July 2012 observations show slope-parallel velocities that range from ~0.0-1.5 m/day for

336  both the Nisqually and Emmons glaciers (Fig. 3A, 4, 6). Both display high velocities over their

337  upper and central regions that taper into essentially stagnant (<0.05 m/day) debris-covered

338  regions near the terminus. In general, slope-parallel velocities near the summit are small (<0.2

339  m/day).

340  On the Nisqually Glacier, a series of local velocity maxima (>1.0 m/day) are associated with

341  increased surface slopes between local surface highs. Local velocity maxima are also observed

342  for the fast-flowing Nisqually icefall (western branch of Upper Nisqually, see Fig. 3) and above

343  the Nisqually ice cliff (eastern branch). A relatively smooth velocity gradient from slow- to fast-

344  moving ice is present upstream of the icefall, while the velocities above the ice cliff display a

345  steep velocity gradient (Fig. 3).

346  The main (south) branch of the Emmons glacier displays generally increasing velocity from the

347  summit to lower elevations. A large high velocity region (>0.7-1.1 m/day) is present over central

348  Emmons, downstream of the confluence of upper branches. These elevated velocities decrease

349 at lower elevations, where ice thickness increases and surface slopes decrease (Fig. A5).  A

350 central "core" of exposed ice displays slightly elevated velocities relative to surrounding debris-

351 covered ice within ~1-1.5 km of the terminus.

352 Velocities exceed 1 m/day over the "central" branch of the Upper Emmons Glacier, where flow

353 is restricted between two parallel bedrock ridges, with local maxima similar to Nisqually.

354 Velocities at higher elevations within the "central" branch appear slower (<0.1-0.5 m/day),

355 separated from the fast downstream velocities by a small area that was excluded due to phase

356 unwrapping errors.  Photographs show that this area appears heavily fractured with many large

357 blocks indicative of rapid, discontinuous flow (Fig. A3).

358 Smaller, relatively thin glaciers, such as the Fryingpan, Upper Kautz, and Inter Glacier (labeled

359 on Fig. 1), also display nonzero surface velocities of <0.1-0.2 m/day, but with limited spatial

360 variability.

361 **4.2    Seasonal variability**

362 The repeat observations from the ROI viewpoint provide time series that capture seasonal

363 velocity variability for the Nisqually, Wilson and Upper Kautz Glaciers.  We observe significant

364 velocity changes during the summer to winter transition and more subtle changes within the

365 winter period. These changes are shown in map view on Fig. 4 and in profile view with

366 corresponding slope and ice thickness on Fig. 6.

367 These data show a velocity decrease of 0.2-0.7 m/day (-25 to -50%) from July to November 2012

368 for most of the Nisqually Glacier.  This includes central and lower Nisqually and the ice above

369 the ice cliff.  The greatest velocity decreases are observed near the crest and lee of surface rises

370 (downstream of data gaps from radar shadows, Fig. 4), where some of the highest velocities were

371 observed in July.  In contrast, the area immediately downstream of the ice cliff and the area

372 surrounding the icefall both display an apparent velocity increase for the same time period (Fig.

373 4, 6).  While the increase is less than the 95% confidence interval for most areas, we can

374 confidently state that the icefall and area below the ice cliff do not display the significant

375 decrease in velocity observed elsewhere.

376 The majority of the Wilson Glacier displays a similar ~0.3-0.7 m/day (-40 to -60%) velocity

377 decrease from July to November.  Interestingly, the steep transition where the Wilson merges

378 with the Nisqually displays an apparent velocity increase of ~0.1 m/day during this time period

379 (Fig. 4). These data also reveal subtle velocity increases in the debris-covered ice near the

380 Nisqually terminus and the Upper Kautz glacier (Fig. 4), though these increases are statistically

381 insignificant.

382 The repeat winter observations of Nisqually show relatively constant velocities with some

383 notable variability. Analysis of the 2 Nov. to 10 Dec. observations reveals a statistically

384 significant -0.1 m/day (-50%) velocity decrease ~1 km upstream of the terminus (centered on

385 ~0.7 km in Fig. 6A profile), a +0.1 to +0.2 m/day (+20 to +30%) increase over central Nisqually

386 centered on ~3.5 km in the Fig. 6D profile, and an apparent +0.2 m/day (+130%) increase over

387 the Upper Wilson. In the latter case, the 10 Dec. velocities are actually higher than those

388 observed in July. The slowdown over lower Nisqually appears robust, but other trends have

389 amplitudes that are mostly below the 95% confidence interval for the 27 Nov. and 10 Dec.

390 observational campaigns (Fig. 4).

391 **4.3 Diurnal variability**

392 We collected ~21 and ~24 hour time series for the Emmons and Nisqually/Wilson Glaciers

393 (Table 1) in July and November, respectively, and look at changes throughout the day. Although

394 uncertainties are large, we present the time series on Fig. 7.

395 In general, velocities for these regions remain relatively constant during their respective

396 sampling periods. The Emmons time series shows an apparent decrease in velocity over the

397 central, fast-flowing regions (B, C, D in Fig. 7A) from ~18:00 to 21:00 local time, and an

398 apparent increase between ~07:00 to 09:00 local time (Fig. 7A). The Nisqually time series

399 shows an apparent decrease from ~06:00 to 11:00 local time for the icefall and ice cliff, and an

400 apparent decrease for several areas of the glaciers followed by an increase (Fig. 7B). However,

401 uncertainties are large and none of these are statistically significant.

402 **4.4 Comparison with independent velocity measurements**

403 We now compare our TRI results with independent velocity measurements for an overlapping

404 time period. Walkup et al. (2013) performed repeat total station surveys to document the

405 location of sparse supraglacial cobbles and boulders on the lower Nisqually glacier from 2011-

406 2012. While measurement errors (e.g., cobble rolling/sliding) for these observations are difficult

407    to document, the large sample size and relatively long measurement intervals allow for accurate

408    surface velocity estimates.

409    Figure 8 shows average velocity vectors measured by Walkup et al. (2013) for the period

410    between 19 July and 11 October 2012, with corresponding surface-parallel velocity vectors from

411    the 7 July and 2 November TRI surveys. This comparison is summarized on Table 4. In general,

412    the velocity magnitudes are similar, with the overall mean of the Walkup et al. (2013)

413    measurements slightly higher on average, but often falling between the 7 July and 2 November

414    GPRI magnitudes, as would be expected of a mean velocity spanning approximately the same

415    period.  The velocity directions are also relatively consistent, with a median angular difference of

416    12°. The greatest deviations are observed near the ice margins and over small-scale local

417    topography (e.g. ice-cored moraine near western margin), where surface-parallel flow

418    assumptions break down.  In general, the two techniques provide similar results and offer

419    complementary data validation. However, since the Walkup et al. (2013) measurements were

420    limited to accessible areas, they cannot be used to validate TRI observations for heavily

421    crevassed areas, icefalls, and other hazardous dynamic areas generally higher on the mountain.

**4.5 2-D flow modeling**

423    Figure 9 shows modeled deformation, sliding velocity residual (observations - deformation

424    model), and sliding percent (sliding velocity residual as percentage of total velocity) with best

425    estimate model parameters for Nisqually glacier in July and November.  Figure 10 shows

426    corresponding output for Emmons.  The SIA deformation models suggest that most areas of both

427    glaciers are moving almost entirely by sliding. The modeled glacier deformation alone is unable

428    to account for the observed surface velocity during any of the observation periods. Only a

429    median of 1% of the velocity field over the Nisqually glacier area can be explained by internal

430    deformation in July, and only 2% in November. If we consider ±25% ice thickness and up to 2x

431    the ice softness, the possible range of the median deformation contribution is still small, 0.5 –

432    7% in July and 0.5 – 8 % in November. If we consider only ±25% ice thickness and do not

433    change the ice softness, the range narrows to 0.5 – 4% in both cases. Using stake measurements,

434    Hodge (1974) estimated deformation contributed ~5-20% of the velocity for the upper third of

435    the ablation area of the Nisqually glacier. He did not study any areas above the equilibrium line,

436    so to compare directly to Hodge's (1974) numbers, we take the median deformation percentage

437    over approximately the upper third of the ablation area and find a best estimate of 1% (range 0.3

438    – 5%) for July and 2% (range 0.5 – 7%) for November. These numbers suggest that sliding is

439    even more dominant than Hodge (1974) estimated in this area, though it is difficult to say if the

440    differences are real (i.e. sliding was higher in 2012 than it was five decades ago) or just due to

441    differences in methods and assumptions.

442    The model results for Emmons suggest that deformation is more important for the Emmons

443    glacier than for Nisqually. A median of 9% of the July velocity field of Emmons can be

444    explained by deformation, with a possible range of 3 – 40% when considering ±25% ice

445    thickness and up to 2x the ice softness. If we consider only ±25% ice thickness, the range

446    narrows to 3 – 20%.

447    There are a few regions where the observed surface velocity can be explained entirely or nearly

448    entirely by internal deformation.  These include the area within ~1-2 km of the Nisqually and

449    Emmons Glacier terminus, where ice is relatively thick and observed velocities are small.

450    **5. Discussion**

451    The continuous coverage of the TRI provides information about the spatial distribution of

452    surface velocities and strain rates.  Several local velocity maxima are apparent along the

453    centerline of the Nisqually glacier and the central branch of the Emmons glacier. These velocity

454    maxima are associated with surface crevasses and increased surface slopes, with peak velocities

455    typically observed just upstream of peak slope values (Fig. 6).  They are likely related to

456    accelerated flow downstream from local bedrock highs,

457    However, the local velocity maxima at ~2.1 km in Fig. 6 corresponds to a region of decreased

458    surface slopes and increased ice thickness. This location also displayed significant seasonal

459    velocity change, which could be related to variations in local subglacial hydrology (e.g. reservoir

460    drainage) during this time period.

461    **5.1 Icefall and ice cliff dynamics**

462    Terrestrial radar interferometry offers new observations over dynamic, inaccessible areas that

463    have received limited attention in previous studies (e.g., icefalls, ice cliffs).  For example, the

464    velocities above the Nisqually ice cliff display an abrupt transition from slow- to fast-moving ice

465    (Fig. 4).  The high strain rates associated with this transition are suggestive of crevasse opening

466 and "detached slab" behavior rather than continuous flow, which is reflected in the heavily

467 crevassed surface at this location.

468 Our results show that the Nisqually icefall and the icefall at the convergence of the Wilson and

469 Nisqually glaciers show a slight increase in velocity from July to the winter months. This

470 suggests that the icefalls may not be susceptible to the same processes that caused the seasonal

471 velocity decrease over much of the rest of the glacier. This may indicate that there is a lack of

472 local continuity through icefalls, which appears to prevent or dampen propagation of

473 downstream seasonal velocity decreases. It could also indicate that the icefall is relatively well-

474 drained year-round, and is not significantly affected by seasonal changes in subglacial

475 hydrology. A potential explanation for the observed minor increase in velocity could be early

476 winter snow accumulation on blocks within the icefall.

477 Interestingly, in contrast to the icefall, the hanging glacier above the Nisqually ice cliff displayed

478 a significant velocity decrease from July to November, despite similar steep surface slopes and

479 crevasse density. This could potentially be related to the lack of backstress from downstream ice

480 and an increased sensitivity to minor fluctuations in subglacial hydrology. Hanging glaciers are

481 also thought to be the source of some of the repeating glacial earthquakes that are triggered by

482 snow loading (Allstadt and Malone, 2014), which highlights their sensitivity to minor

483 perturbations.

484 **5.2    Lack of significant diurnal variability**

485 We expected to see significant variability over the 24-hour July time series for Emmons, as

486 atmospheric temperatures varied from 16°C to 27°C at Paradise Visitors Center (~1600 m.a.s.l),

487 and skies remained cloud-free during data collection. We hypothesized that the resulting

488 increase in meltwater input from late morning through late afternoon might produce an

489 observable increase in sliding velocity. While the results potentially show a slight velocity

490 decrease at higher elevations overnight, and a slight velocity increase in the morning (Fig. 7A,

491 A-D), these changes are not statistically significant, nor coincident with times expected to have

492 highest melt input. The lack of a significant diurnal speedup suggests that the subglacial conduits

493 are relatively mature by July, and are capable of accommodating the diurnal variations in

494 meltwater flux without affecting basal sliding rates.

495    We did not expect to see significant diurnal changes in the 21-hour November time series for

496    Nisqually (Fig. 7A), as atmospheric temperatures ranged between 2°C and 6°C at Paradise

497    Visitors Center (~1600 m a.s.l.) and skies were partly-cloudy to overcast during data collection,

498    so surface meltwater input should have been minimal.  Our results show only a minor velocity

499    decrease higher on the glacier in the morning hours but it is not statistically significant and does

500    not occur at times when we would expect increased meltwater.

501    Though some of the subtle changes in the extended time series may reflect actual diurnal

502    velocity variability, we cannot interpret these with confidence.  This suggests that the magnitude

503    of diurnal variability, if it exists, during these time periods is minor when compared to the

504    observed seasonal changes.  It also implies that other stacks derived from a subset of the day can

505    be considered representative of the daily mean, and can be compared for seasonal analysis.

**5.3 Seasonal velocity changes**

507    The observed seasonal velocity changes from July to November can likely be attributed to

508    changes in glacier sliding, which in turn are driven by evolving englacial and subglacial

509    hydrology (Fountain and Walder, 1998).  During the spring-summer months, runoff from

510    precipitation (i.e. rain) and surface snow/ice melt enters surface crevasses, moulins, and/or

511    conduits near the glacier margins.  This water travels through a series of englacial fractures,

512    reservoirs and conduits, and eventually ends up in a subglacial network of channels and

513    reservoirs between the ice and bed.  Storage time and discharge rates within the subglacial

514    system are variable, with water finally exiting the system through one or more proglacial streams

515    at the terminus.  This dynamic system is continuously evolving due to variable input, storage

516    capacity, and output. In early July, ongoing snowmelt should produce high meltwater discharge

517    that travels through a relatively efficient network of mature conduits.  As discharge decreases

518    later in the summer, these subglacial conduits/reservoirs close due to ice creep without high flow

519    to keep them open through melting due to heat from viscous dissipation.  By November, there

520    should be little or no surface meltwater input and we would expect to see a minimum in basal

521    sliding velocity (Hodge, 1974).  This is consistent with the observed velocity decrease in Fig. 4.

522    However, the deformation modeling results (Fig. 9) show that a significant sliding component is

523    still present for most of the Nisqually glacier in November and December, when minimum

524    surface velocities are expected.

525    The spatial patterns of the velocity change observed between July and November can be used to

526    infer the extent of basal sliding.  This may provide some insight into subglacial water storage,

527    since the deformation component of surface velocity should remain nearly the same year-round.

528    Fig. 4 indicates that almost the entire Nisqually glacier slows down significantly between July

529    and November, suggesting that storage is occurring under most of the glacier below the icefall

530    and ice cliff. Significant velocity decreases are observed near local surface rises (Fig. 4), where

531    some of the highest velocities were observed in July. This suggests that there are likely

532    subglacial cavities downstream of these areas with high basal water pressures that can support

533    enhanced sliding during the summer.

534    Hodge (1974) interpreted a delay in both the maximum summer velocity and minimum winter

535    velocity between the terminus and ELA as a propagating "seasonal wave" traveling ~55 m/day.

536    While our sampling is limited, the continued November 2 to November 27 slowdown over the

537    lower Nisqually near the terminus (Fig. 4F) could represent a delayed response to the significant

538    slowdown over central Nisqually. This might be expected, as surface velocities near the terminus

539    are dominated by internal deformation and should respond more slowly than areas dominated by

540    basal sliding.

541    **5.4 Comparison with historical velocity measurements**

542    As described earlier, Hodge (1972, 1974) measured surface velocity for a network of centerline

543    stakes on the lower Nisqually from 1968-1970.  He documented a significant seasonal cycle with

544    minimum velocities in November and maximum velocities in June.

545    To put our velocity data in historical context, we digitized Hodge's (1972) July and November

546    1969 surface velocity data at 19 stake locations along a profile of the lower half of the Nisqually

547    glacier.  We then sampled the 2012 TRI slope-parallel velocities at these locations (Fig. 11).

548    Remarkably, in spite of significant terminus retreat of up to ~360 m and surface elevation

549    changes of approximately -20 m (Sisson et al., 2011), the November 1969 and November 2012

550    surface velocities are almost identical at stakes 12-20, suggesting that bed properties and local

551    geometry have greater influence over sliding velocity than ice thickness or relative distance from

552    the terminus.  In contrast, the July 2012 velocities at stakes 12-20 are 8-33% faster than the July

553    1969 velocities.  The ice is mostly sliding at these locations, so the change could be related to a

554    difference in the timing of the peak summer velocities, or potentially enhanced sliding in 2012.

555 The nearly identical surface velocities in November 1969 and 2012 suggests that the discrepancy

556 between Hodge's sliding percentage estimates and our estimates (section 4.5) is likely related to

557 different methodology and assumptions rather than actual changes in sliding since 1969.

558 The most notable difference between the profiles is observed closer to the terminus at stakes 7-

559 12. At these locations, the July and November 2012 velocities are both <0.05 m/day, whereas

560 July and November 1969 velocities are ~0.2 and ~0.1 m/day, respectively, with significant

561 seasonal variability. This suggests that the ice near the present-day terminus is essentially

562 stagnant and no longer strongly influenced by changes in subglacial hydrology.

## 563 6. Conclusions

564 In this study, we used repeat TRI measurements to document spatially continuous velocities for

565 the Emmons and Nisqually glaciers at Mount Rainier, WA. We produced surface velocity maps

566 that reveal speeds of >1.0-1.5 m/day over the upper and central regions of these glaciers, <0.2

567 m/day near the summit, and <0.05 m/day over the stagnant ice near their termini. Novel data

568 processing techniques reduced uncertainties to ±0.02-0.08 m/day, and the corrected, surface-

569 parallel TRI velocities for Nisqually display similar magnitude and direction with a set of sparse

570 interannual velocity measurements (Walkup et al., 2013).

571 Repeat surveys show that Nisqually glacier surface velocities display significant seasonal

572 variability. Most of the glacier experienced a ~25-50% velocity decrease (up to -0.7 m/day)

573 between July and November. These seasonal variations are most likely related to changes in

574 basal sliding and subglacial water storage. Interestingly, the steep icefall displays no velocity

575 change or even a slight velocity increase over the same time period. We documented no

576 statistically significant diurnal velocity variations in ~24-hour datasets for Nisqually and

577 Emmons, suggesting that subglacial networks efficiently handled diurnal meltwater input.

578 Comparisons with 1969 velocity measurements over the Lower Nisqually (Hodge, 1972; 1974)

579 reveal similar November velocities in both 2012 and 1969, and faster July velocities in 2012.

580 Using a simple 2D ice flow model, we estimate that basal sliding is responsible for most of the

581 observed surface velocity signal except in a few areas, mainly near the termini. The model

582 suggests that about 99% of the July velocity field for the Nisqually glacier is due to sliding. Even

583 when we account for the large uncertainties in ice thickness and ice softness, the possible range

584    of sliding percentage is still narrow: 93 – 99.5% Deformation is more important for the Emmons

585    glacier, where we estimate 91% of the observed motion is due to sliding, with a much wider

586    possible range of 60 – 97% when accounting for uncertainties.

587    In summary, TRI presents a powerful new tool for the study of alpine glacier dynamics. With

588    just a few hours of fieldwork for each survey, we were able to document the dynamics of several

589    glaciers at Mount Rainier in unprecedented extent and detail from up to 10 km away. TRI is

590    particularly well suited for examining diurnal and seasonal glacier dynamics, especially for areas

591    that are difficult to access directly (e.g., icefalls), like many parts of the glaciers at Mount

592    Rainier. Repeat surveys provide precise surface displacement measurements with unprecedented

593    spatial and temporal resolution, offering new insight into complex processes involving subglacial

594    hydrology and basal sliding.  Future studies involving coordinated, multi-day TRI occupations

595    during critical seasonal transition periods could undoubtedly provide new insight into these and

596    other important aspects of alpine glaciology.

597    **Figure Captions**

598    Figure 1. Glaciers at Mount Rainier and locations of viewpoints used for ground based radar

599    interferometry. Instrument view angle ranges are indicated by arrows extending away from each

600    viewpoint location. Boxes A-C show zoom areas for later figures. Inset map shows regional

601    location of Mount Rainier. Glacier outlines in this and subsequent figures are from Robinson et

602    al. (2010).

603    Figure 2. GPRI equipment setup during 27 Nov 2012 campaign at ROI viewpoint.

604    Figure 3. A) Median slope-parallel velocity derived from TRI for GLPEEK and SUNRIZ

605    viewpoints taken on July 6-7, 2011. B) Width of 95% confidence interval (high minus low limits

606    for slope parallel flow field) of slope parallel velocities for July 6-7, 2011 computed by

607    bootstrapping after performing atmospheric noise corrections and stacking. Area shown is

608    indicated by Box A on Fig. 1.

609    Figure 4. A-D) Median slope-parallel velocities for Nisqually and Wilson glaciers for four

610    different time periods taken from ROI viewpoint. Dashed lines on top left panel show locations

611    of profiles taken to create Fig. 6, markers indicate distance in km.  E-G) Percent change in

612    median slope-parallel velocity for the Nisqually and Wilson glaciers between time periods. Blue

613    indicates a velocity decrease and red indicates a velocity increase relative to the earlier time

614    period, gray polygons indicate areas where velocity change is significant with 95% confidence.

615    Area shown is indicated by Box B on Fig. 1.

616    Figure 5. Width of 95% confidence interval (high minus low limits for slope parallel velocity)

617    over Nisqually glacier computed by bootstrapping. Shown for four sampling periods from the

618    ROI viewpoint. Note that the color bar is scaled differently than Fig. 3B.

619    Figure 6. A and C: Slope parallel velocity profiles along the two branches of Nisqually glacier

620    (profile lines shown in map view on Fig. 4A) for all sample time periods and viewpoints. B and

621    D: Surface slope and ice thickness along each profile line. Surface slope is smoothed identically

622    to that used for slope parallel corrections (see text), ice thicknesses are estimated from digitized

623    basal contours from Driedger and Kennard (1986) and surface elevations from the 2007/2008

624    LiDAR (Robinson et al., 2010). Refer to Figure 5 and Table 2 for uncertainty estimates.

625    Figure 7. LOS velocity time series for areas outlined on maps to the right. Shaded region around

626    each line represents ± one standard error for a 2-hour running mean.  a) 24-hour timeseries at

627    SUNRIZ on July 7-8, 2012, gray box indicates the period with poor data quality (see text for

628    details).  b) 22-hour timeseries at ROI on Nov 1-2, 2012.

629    Figure 8. Comparison of average azimuth and velocities measured by Walkup et al. (2013)

630    between 19 July 2012 and 11 October 2012 (black) compared to TRI slope-parallel velocities

631    derived from this study at the same locations for two time periods that bracket the time period

632    measured by Walkup et al. (2013). See Table 4 for comparison statistics and Box C on Fig. 1 for

633    context.

634    Figure 9. Model results for summer (6 July 2012) and a late fall (2 Nov 2012) time period for

635    Nisqually and Wilson glaciers. A, D) Modeled surface velocity for internal deformation, B, E)

636    Sliding residual (observed slope parallel velocity minus the modeled deformation velocity), C, F)

637    Estimate of the sliding percentage (sliding residual divided by total slope-parallel velocity).

638    Figure 10. Same as Fig. 9 but for Emmons Glacier.

639    Figure 11. July and November 1969 surface velocities measured by Hodge (1974, digitized from

640    Hodge, 1972) at 19 stake locations along lower Nisqually profile (circles), compared with

641    sampled 2012 slope-parallel velocities for corresponding locations/seasons (triangles). Stake

locations are labeled and indicated with dotted lines and are shown in map view at right (same map extent as Fig. 8).

Figure A1. Photomosaic acquired from ROI viewpoint on July 5, 2012. Approximate glacier outlines shown in red.

Figure A2. Photomosaic acquired from GLPEEK viewpoint on July 6, 2012. Approximate glacier outlines shown in red.

Figure A3. Photomosaic acquired from SUNRIZ viewpoint on July 7, 2012. Approximate glacier outlines shown in red.

 Figure A4. Pair of multi-look intensity (MLI) radar images from ROI viewpoint (left and center) generated from original single-look complex (SLC) images multi-looked by 15 samples in range and multi-looked interferogram generated from the SLC images (right).

Figure A5. A) Filtered ice thickness and B) filtered slope used as model inputs.

**Appendix A**

Appendix A contains supplementary figures.

**Author Contribution**

K. E. Allstadt coordinated the effort, developed methods, performed data acquisition and processing, made the figures and prepared the manuscript. D.E. Shean developed methods, performed data acquisition, processing, analysis, and interpretation of results, and contributed significantly to the manuscript. A. Campbell performed modeling experiments and contributed the related section of the manuscript. M. Fahnestock and S. Malone contributed significantly to experiment design, establishment of objectives, data acquisition, and manuscript review.

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783

784    Table 1 Survey parameters

| Sampling Location | Start Time (UTC) | End Time (UTC) | Survey length (hr) | Latitude | Longitude | Elev (m) | Glaciers in view | Sampling Interval (mins) | Number of Scans | Azimuth sweep | Chirp length (ms) | Chirp Band width (Hz) | Antennae angle |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| GLPEEK | 7/6/12 17:20 | 7/6/12 23:37 | 6.3 | 46.7924 | -121.7399 | 1788 | Nisqually, Wilson | 3 | 105 | 75° | 4 | 50 | +15° |
| SUNRIZ | 7/7/12 19:50 | 7/8/12 19:56 | 24.1 | 46.9157 | -121.6492 | 1929 | Emmons, Winthrop | 3 | 436 | 29° | 8 | 25 | +5° |
| ROI | 7/6/12 0:32 | 7/6/12 5:23 | 4.8 | 46.7836 | -121.7502 | 1564 | Nisqually, Wilson | 3 | 62 | 68° | 4 | 50 | +15° |
| ROI | 11/2/12 1:20 | 11/2/12 23:14 | 21.9 | 46.7837 | -121.7502 | 1559 | Nisqually, Wilson | 3 | 377 | 52° | 4 | 50 | +15° |
| ROI | 11/27/12 18:47 | 11/28/12 0:29 | 5.7 | 46.7836 | -121.7502 | 1563 | Nisqually, Wilson | 3 | 107 | 60° | 4 | 50 | +15° |
| ROI | 12/10/12 20:50 | 12/11/12 1:32 | 4.7 | 46.7836 | -121.7502 | 1562 | Nisqually, Wilson | 3 | 91 | 70° | 4 | 50 | +15° |

785
786

787    Table 2 Summary of uncertainty estimates of median stacks

| Sampling Location | Start Time (UTC) | End Time (UTC) | Total interferograms used/total collected | Median confidence interval width over ice* (m/day) | |
|---|---|---|---|---|---|
| | | | | Before correction^ | After correction |
| GLPEEK | 7/6/12 17:20 | 7/6/12 23:37 | 93/105 | 0.23 | 0.07 |
| SUNRIZ | 7/7/12 19:50 | 7/8/12 19:56 | 215/436 | 0.14 | 0.09 |
| ROI | 7/6/12 0:32 | 7/6/12 5:23 | 56/62 | 0.33 | 0.11 |
| ROI | 11/2/12 1:20 | 11/2/12 23:14 | 359/377 | 0.16 | 0.04 |
| ROI | 11/27/12 18:47 | 11/28/12 0:29 | 100/107 | 0.44 | 0.10 |
| ROI | 12/10/12 20:50 | 12/11/12 1:32 | 76/91 | 0.43 | 0.15 |

* derived from bootstrapping, 95% confidence, line of sight velocities

^ correction refers to removing displacements due to atmospheric noise (interpolated over static control surfaces)

788

789    Table 3 Constants used in modeling analysis

| Name | Symbol | Value | Units |
|---|---|---|---|
| Ice softness parameter | $A$ | $2.4 \times 10^{-24}$ | $Pa^{-3}\,s^{-1}$ |
| Side length of reference window | $A_w$ | 120 | m |
| Acceleration of gravity | $g$ | 9.81 | $m\,s^{-2}$ |
| Coupling length | $l$ | 60 | m |
| Flow law exponent | $n$ | 3 | dimensionless |
| Density of ice | $\rho_i$ | 900 | $kg\,m^{-3}$ |
| Density of water | $\rho_w$ | 1000 | $kg\,m^{-3}$ |

790

791    Table 4 Comparison between Walkup et al. (2013) and TRI velocities at Walkup et al. (2013)

792    sample locations (Figure 8)

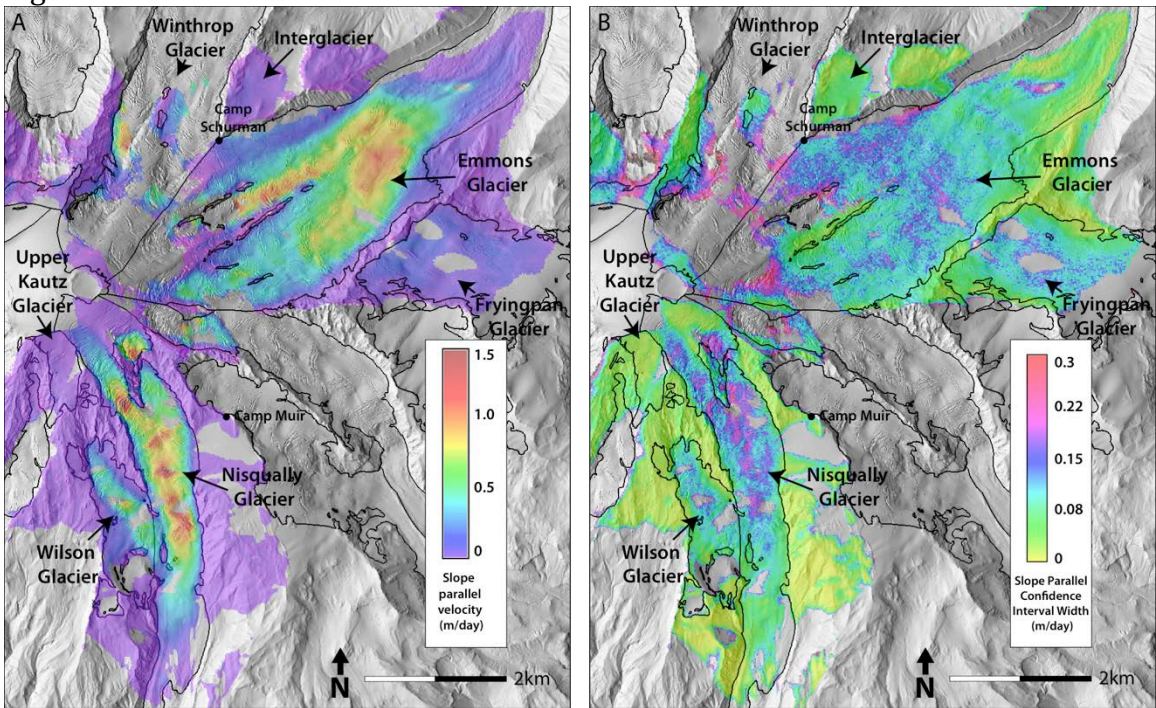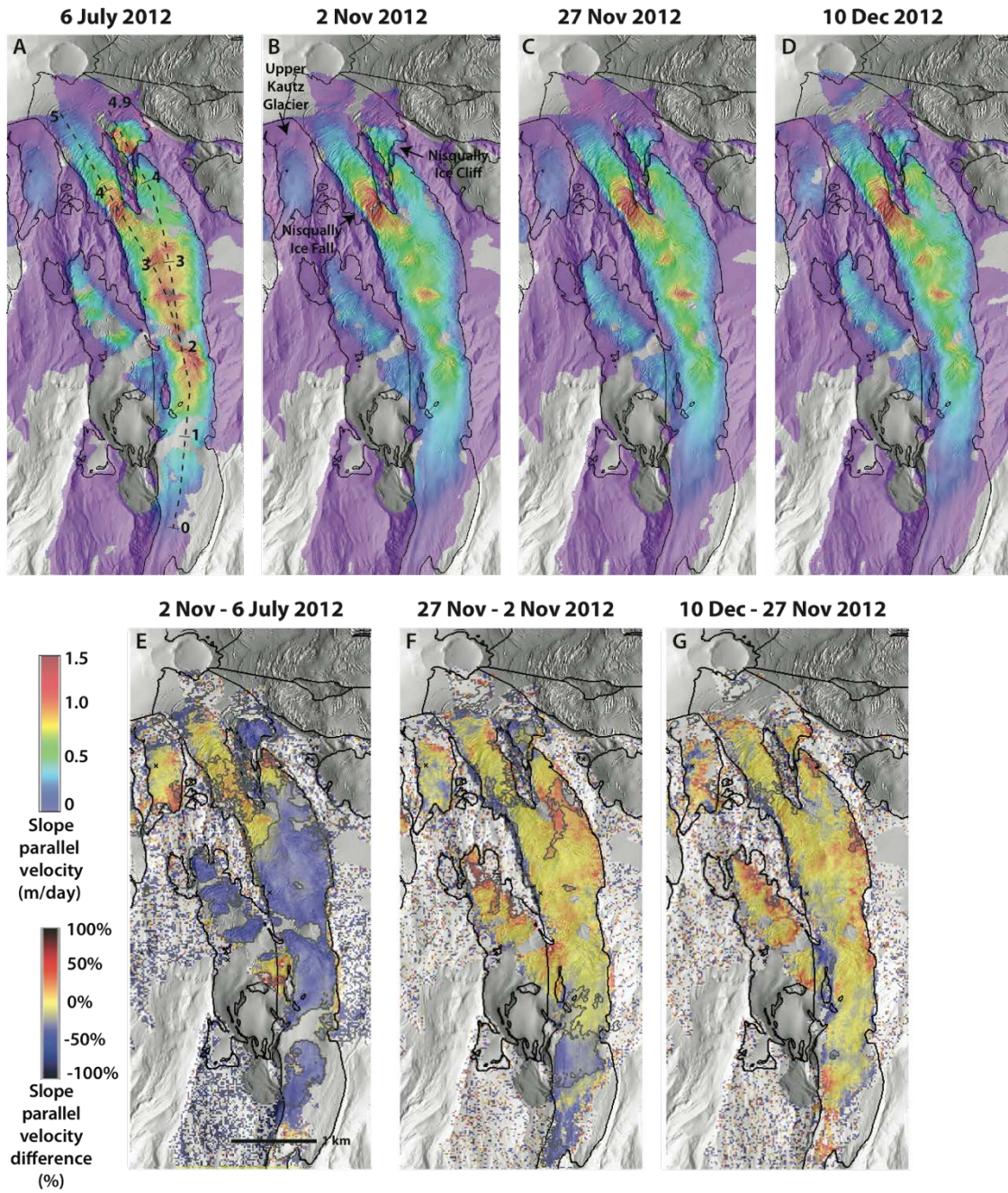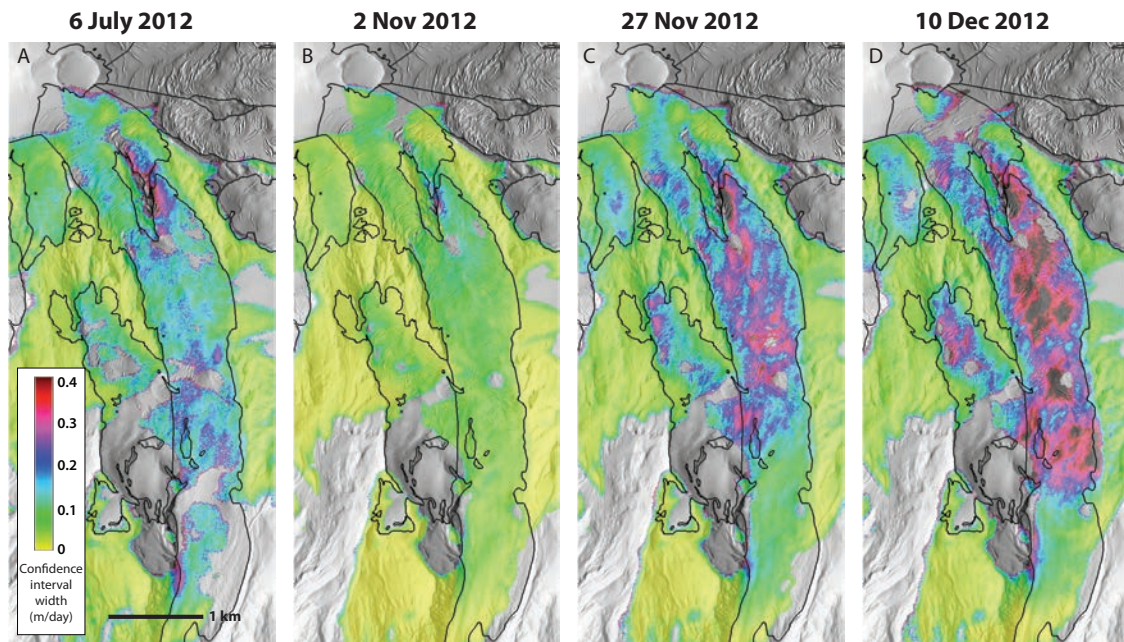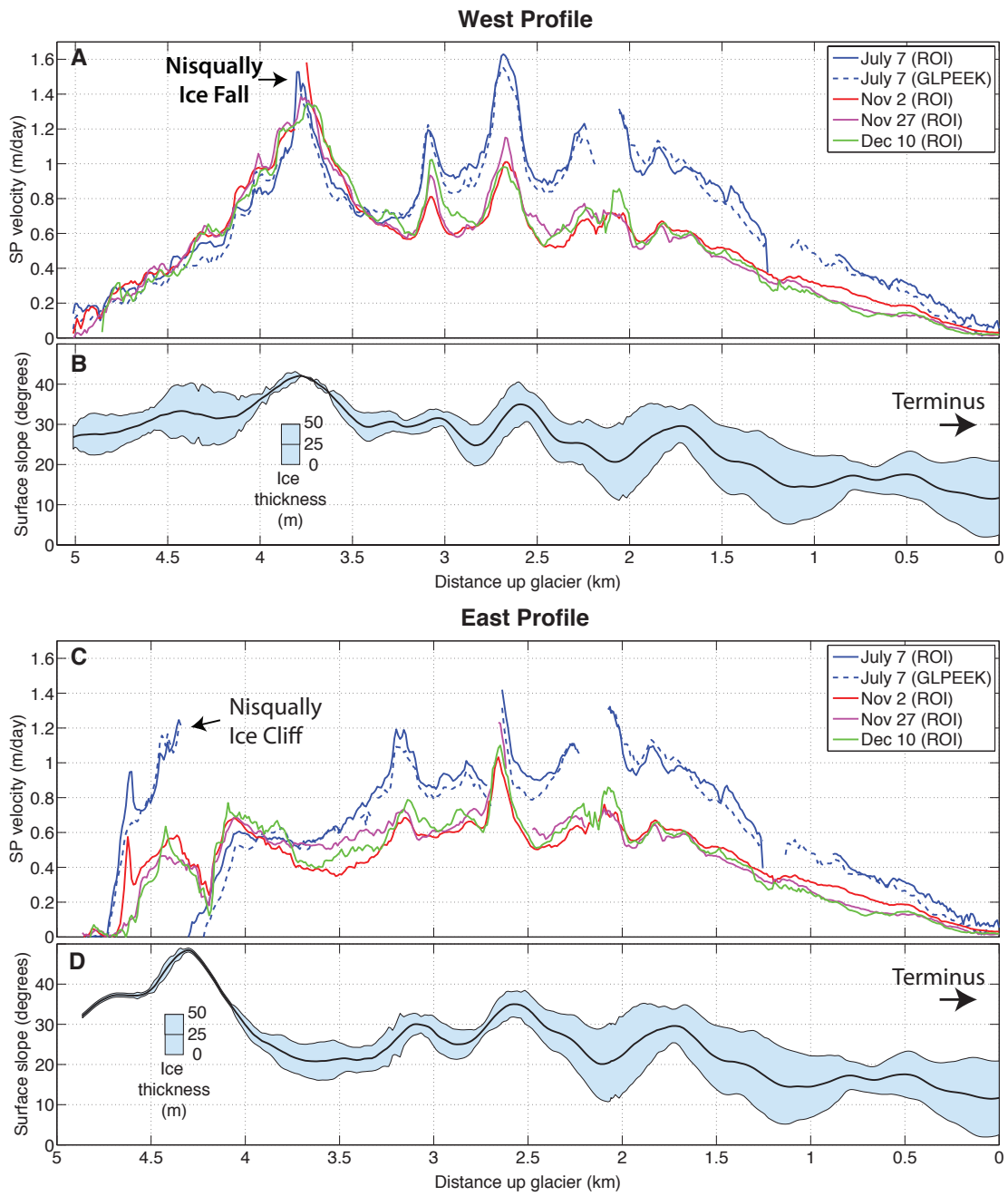| | Velocity Magnitude (cm/day) | | | | Angular Difference from Walkup et al. 2013 (degrees) | | | |
|---|---|---|---|---|---|---|---|---|
| Source | Mean | Median | Max | Min | Mean | Median | Max | Min |
| Walkup et al. 2013 | 22.3 | 16.6 | 64.4 | 1.8 | - | - | - | - |
| GLPEEK July | 20.8 | 10.5 | 82.9 | 0.1 | 15.8 | 12.0 | 55.8 | 0.7 |
| ROI Nov | 14.6 | 10.4 | 51.4 | 0.3 | 15.8 | 12.0 | 55.8 | 0.7 |

793

Figure 1
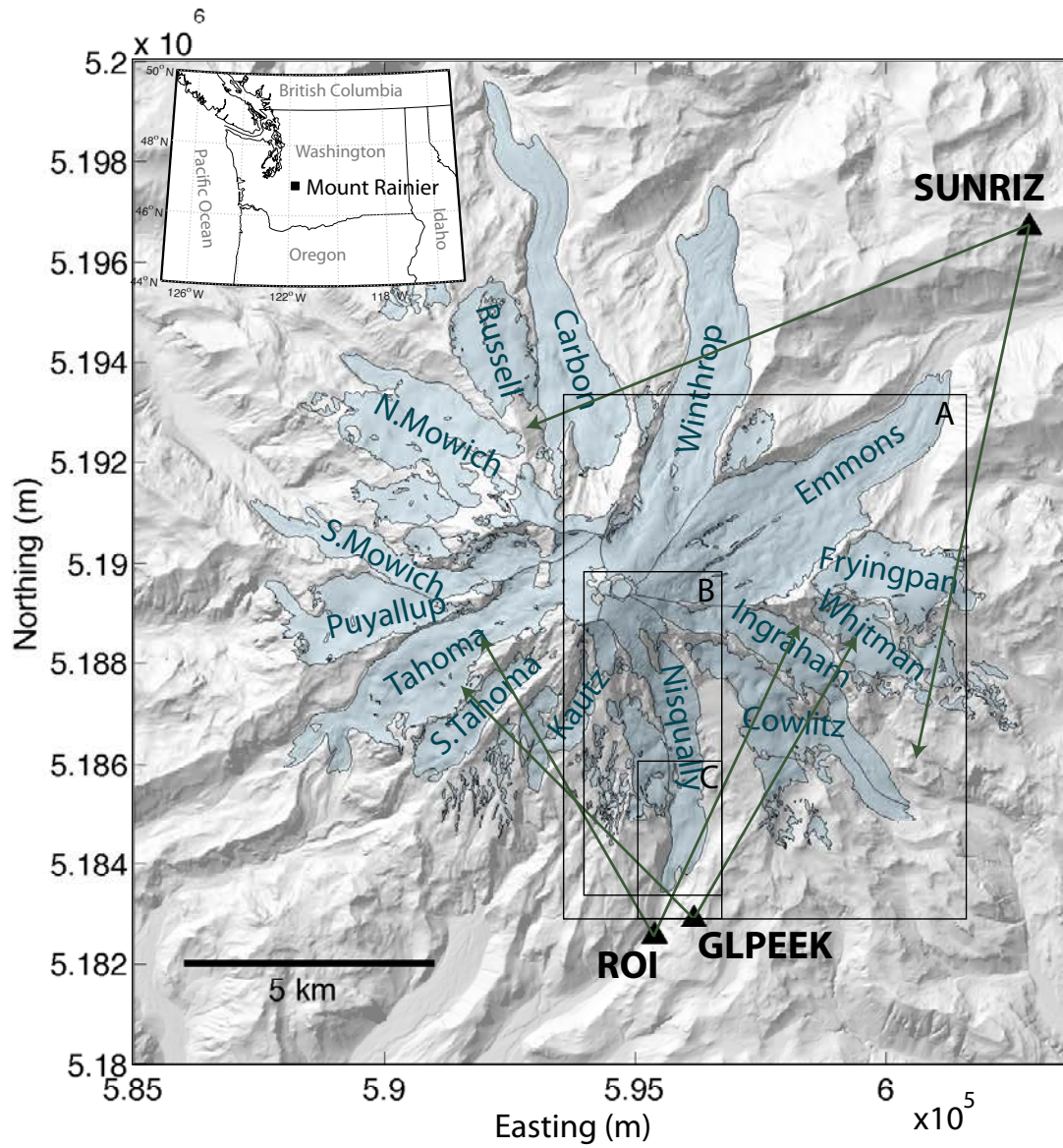


Figure 2

Figure 3

Figure 4



**6 July 2012**    **2 Nov 2012**    **27 Nov 2012**    **10 Dec 2012**

A

B
Upper
Kautz
Glacier

Nisqually
Ice Cliff

Nisqually
Ice Fall

C

D

4.9
5
4
4
3    3
2
1
0

**2 Nov - 6 July 2012**    **27 Nov - 2 Nov 2012**    **10 Dec - 27 Nov 2012**

E    F    G

1.5
1.0
0.5
0

Slope
parallel
velocity
(m/day)

100%
50%
0%
-50%
-100%

Slope
parallel
velocity
difference
(%)

1 km

Figure 5



| 6 July 2012 | 2 Nov 2012 | 27 Nov 2012 | 10 Dec 2012 |

Confidence interval width (m/day)

1 km

Figure 6



## West Profile

**A**

Nisqually Ice Fall

Legend:
- July 7 (ROI)
- July 7 (GLPEEK)
- Nov 2 (ROI)
- Nov 27 (ROI)
- Dec 10 (ROI)

SP velocity (m/day)

**B**

Surface slope (degrees)

Ice thickness (m): 50 / 25 / 0

Terminus

Distance up glacier (km)

## East Profile

**C**

Nisqually Ice Cliff

Legend:
- July 7 (ROI)
- July 7 (GLPEEK)
- Nov 2 (ROI)
- Nov 27 (ROI)
- Dec 10 (ROI)

SP velocity (m/day)

**D**

Surface slope (degrees)

Ice thickness (m): 50 / 25 / 0

Terminus

Distance up glacier (km)

# Figure 7



Emmons Glacier (SUNRIZ) July 7-8, 2012

Nisqually and Winthrop (ROI) Nov 1 - 2, 2012

Figure 8
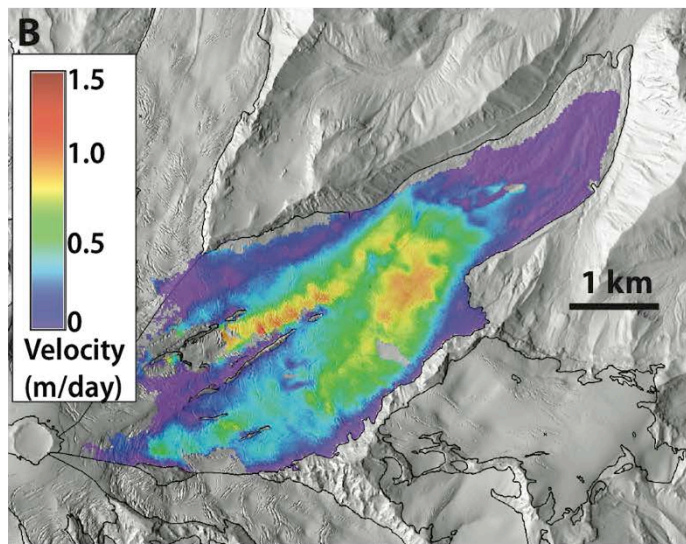
Figure 9



| Deformation model d_model | Sliding residual (observations -d_model) | Sliding residual/ observations (%) |
|---|---|---|

July 6, 2012

A — Velocity (m/day): 1.5, 1.0, 0.5, 0

B — Velocity (m/day): 1.5, 1.0, 0.5, 0

C — Percent Sliding: 100%, 75%, 50%, 25%, 0%

Nov 2, 2012

D — 1 km

E

F

Figure 10

**A** Deformation model
d_model

Velocity
(m/day)

1.5
1.0
0.5
0

1 km

**B** Sliding residual
(observations - d_model)

Velocity
(m/day)

1.5
1.0
0.5
0

1 km

**C** Sliding residual/
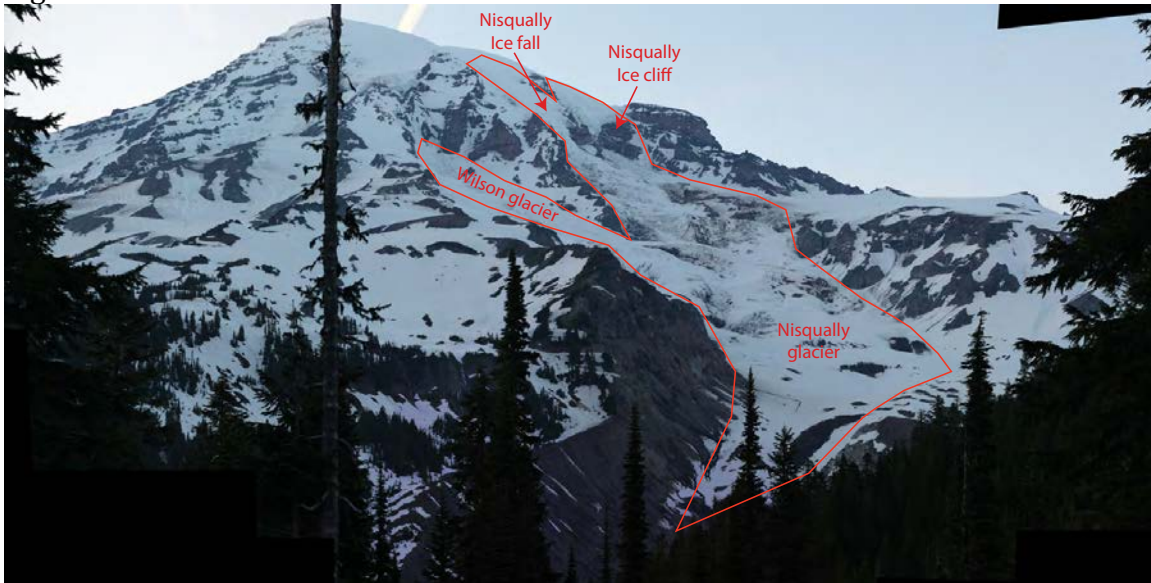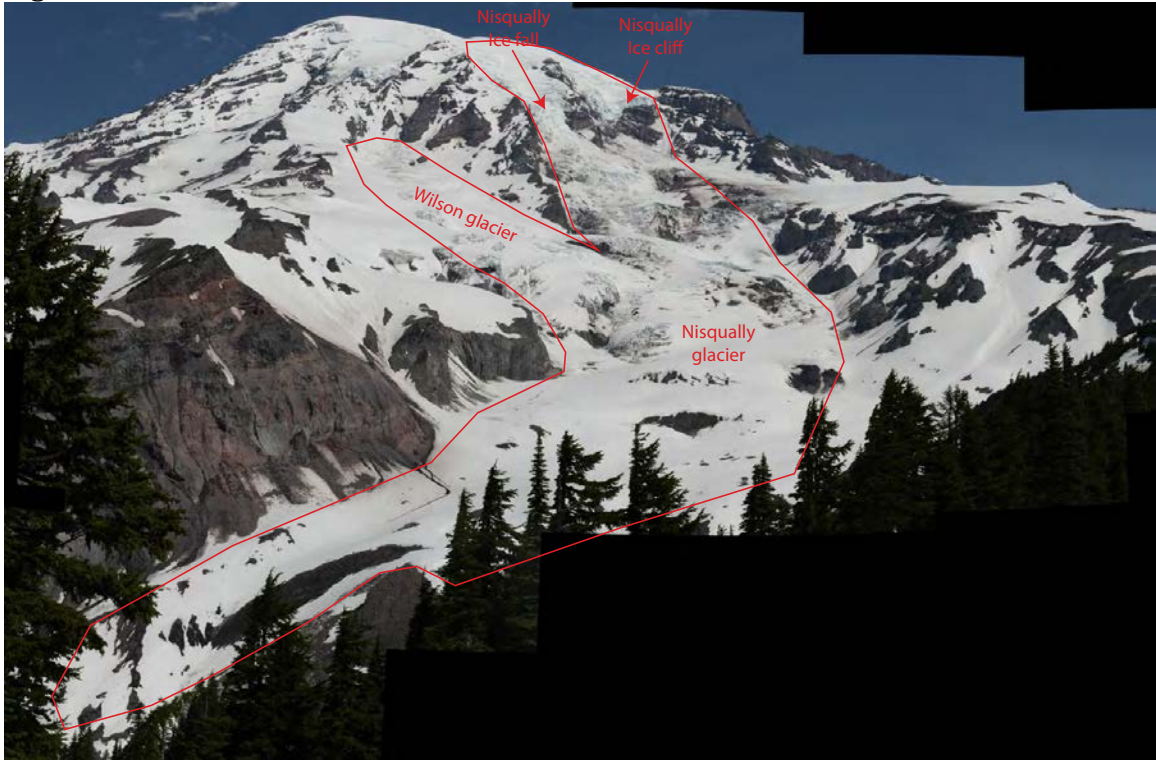observations (%)

Percent
Sliding

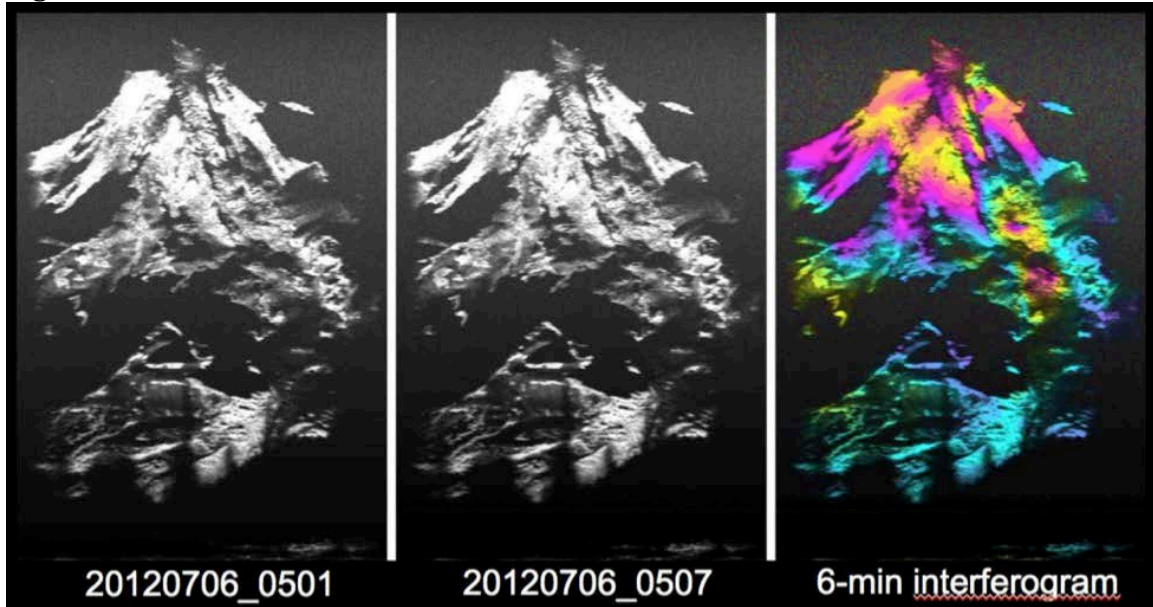100%
75%
50%
25%
0%

1 km
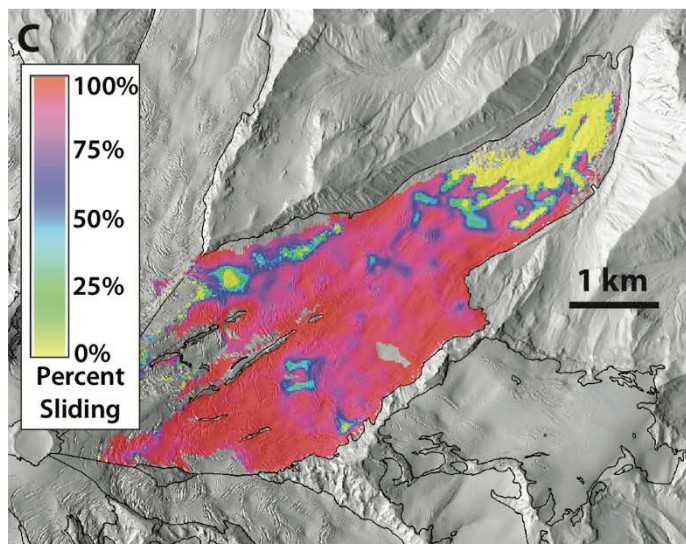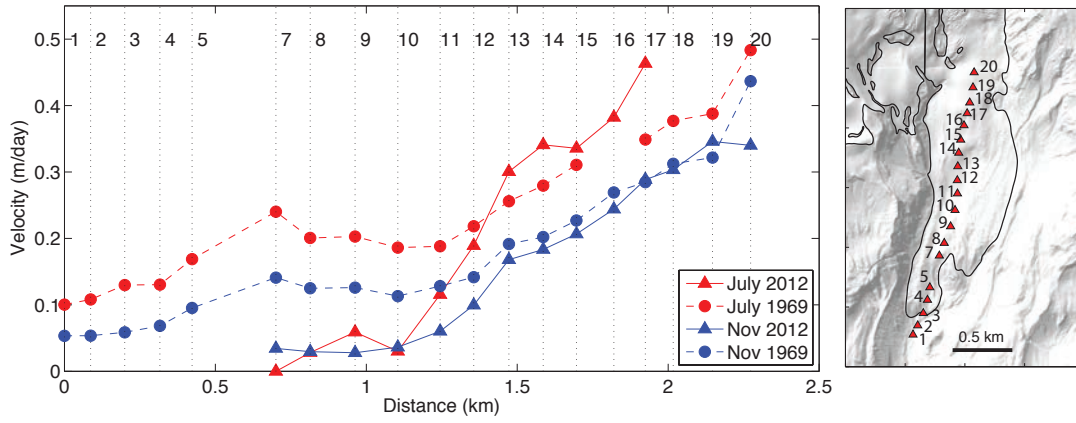
Figure 11



Figure A1

Figure A2



Figure A3

Figure A4



Figure A5