Response to review of "Observations of seasonal and diurnal glacier velocities at Mount Rainier, Washington using terrestrial radar interferometry"
K. Allstadt, D. Shean, A. Campbell, S. Malone, M. Fahnestock

We thank the reviewers for their comments. We have incorporated most of their suggested changes and explain the specifics in response to each comment below. The vast majority of the comments from both reviewers were regarding the modeling, which was mainly included to add additional interpretation of the observations, it was not the main focus of the paper. We recognize that we needed to modify how we incorporated modeling so that it didn't distract from the main point of the study but still contributed. Therefore, we removed the sliding model, which was too simple and problematic in the eyes of both reviewers and did not add much to the paper scientifically anyway, and we introduce uncertainty estimates for the deformation model to address the reviewers' comments regarding that model. We also made some minor modifications to the text to further improve clarity.

Response to comments:
Original comments in black, author response and changes in red

Response to M. Luthi comments

On page 4074, line 11, it is stated that the interferograms were created from MLIs. They are not (since MLI are just signal strength without phase information) but they are created from the SLC data. I would have assumed that this is a typo, but then in Figure A4 the same statement reappears, and is even illustrated. This looks like a serious misunderstanding of the radar data analysis process. In the Gamma software the call signature of the program creating an interferogram is SLC intf <SLC-1> <SLC-2R> ."
Thank you for catching this. We have updated the lines you reference to now say: "Interferograms were generated from single-look complex SLC products with a time separation of 6 minutes, though sometimes longer if acquisition was interrupted (for example images, see Fig. A4). Interferograms were multi-looked by 15 samples in the range direction to reduce noise."

And we changed the caption of Figure A4 to: "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)."

The noise correction with interpolation from bedrock looks interesting, but how robust is it? Atmospheric disturbances are often blob-like and not linear with distance, so it is not immediately clear how useful the method is to reduce noise. It would be interesting to elaborate somewhat more in this.
Indeed, the atmospheric noise is often "blob-like", we see this in the data from Mount Rainier. We spent time looking at the atmospheric noise characteristics (which could be a study on its own), and determined that, qualitatively, the "blobs" are usually larger in scale than the width of Nisqually glacier (~500-900 m across). Bedrock points on either side are typically at distances smaller than the scale of the "blobs" and so the geometry is well-suited for our noise removal method. The geometry isn't quite as favorable for all of the Emmons glacier (~700-2100 m wide), but still acceptable, due to ridges of exposed rock in the middle of the upper Emmons. Our results are quite robust, the velocities of the median stack for each sampling period were very similar whether or not we applied the atmospheric noise correction. The main improvement of the correction was to significantly reduce the uncertainties (reflected as the median confidence interval width - Table 2) and reduce the noise over regions with slow velocities.

We addressed this comment by adding the following to the description of the atmospheric noise correction methods: "Even though atmospheric noise is not necessarily linear with distance, the scale of the atmospheric noise features we observed in the data were typically much wider than the width of the glaciers so we expect the method we use does a reasonable job of approximating the atmospheric noise directly over the glaciers."

And we also added a few sentences to the first paragraph of the Results section, which now reads: "Stacking alone was very effective; the velocities of the mean and median stacks with and without the atmospheric noise correction were very similar. The main benefit of the extra step of using stable rock points to subtract an estimate of the atmospheric noise was to significantly reduce the uncertainties and to reduce the noise where velocities are slow. The uncertainties before and after atmospheric correction are compared on Table 2."

The section 5.3 (p 4084) on flow modeling should be split, with the introductory part moved into the "Methods" section, and the results in the "Results" section. Here one would expect only the discussion of the model results.
We made this change.

The authors use a SIA model which is not well suited for the problem at hand (steep geometry). The authors are fully aware of the problem and even cite three papers using better methods, but do not rely on them at all. Full models in glaciology have been used since the 1980s (e.g. Iken, Echelmeyer, Gudmundsson etc) and have become very easy to use nowadays. Writing this section which sounds like an excuse probably has taken longer than just installing Elmer and modifying one of their examples for the investigated glacier (not that I am advocating a specific code here).
We could have used Elmer here, but we did not feel it was appropriate to use a more complex, full 3D model. The uncertainty in ice thickness would be problematic regardless of model complexity, so we decided to employ a simple model. Furthermore, this is not a modeling paper, it is an observational paper and we invoke the modeling only to aid in interpretion of observed results.

We modified the explanation here to sound less like an excuse, and added uncertainty estimates of sliding percent by assuming a wide range of uncertainty in the thickness and ice softness estimates that go into the deformation model (+-25% thickness and 2x ice softness). Even with these large uncertainties, the deformation for Nisqually still contributes <10% - deformation was so much smaller than the observed velocities in most places that even doubling or tripling deformation didn't change the median percentages much. The possible range for Emmons is much wider than for Nisqually - when we account for the range of uncertainties in the inputs, we get sliding contributions of 60 to 97%.

The implementation of sliding seems cumbersome. Since nothing is known about the process anyway, why formulate it like Equation (B3), and not just formulate it as $u_b = C \tau_b$ (1) with a spatially and temporally varying slipperiness C? This would also alleviate the problem with negative N_{eff} which are probably not as unphysical as the authors think, especially given the serious limitation of the code (no surface evolution, no full stresses).
We agree with the reviewer's suggestion here, but this is no longer an issue because we have decided to remove the sliding model from the paper. The fit to the data was poor and the model perhaps too simple, for many of the reasons discussed both in our text and by the reviewers, and as a result it did not add much scientifically.

The discussion of velocity changes (p 4087, l 20ff) is oversimplistic. It seems to be based on the assumption that N_{eff} is somehow directly related to meltwater supply, and that basal motion is somehow directly controlled by N_{eff}. There are some hardbed sliding theories where these assumptions might hold true, but given that the glaciers reside on a volcano it is likely that their beds consist of sediment, which has a very different rheology and dynamics. With the given data it is impossible to discern between different sliding regimes, but papers like e.g. Clarke (1987) and Clarke (2005) give an idea of the complexity and nonlinearity of possible processes.

We removed the sliding model so this paragraph is no longer in the paper.

4084, 3 The model is "planestrain", not "planview".
We made this change.

4084, 9 Ice thickness and bedrock topography are basically the same (if the surface is known).
We deleted "bedrock topography."

4090, 2 A better reference for the SIA would be Hutter (1983) or Greve and Blatter (2009).
This suggestion may be due to the reviewer having more familiarity with European authors, but we added Greve and Blatter (2009) and keep the citation for Cuffey and Paterson.

4090, 2ff In the formulation of the problem it is very important to consistently specify the coordinate system. Is z pointing vertically up, or perpendicular to mean slope? According to Equation (B1) it is the latter (given the sin term), but then H has to be measured accordingly (i.e. not vertically).
We added "The coordinate system is vertically aligned"

4090,3 In glaciology only the Stokes equations are usually considered, since all acceleration and momentum advection terms are vanishingly small (as proven by scaling arguments).
We changed this to Stokes instead of Navier-Stokes.

4081, 10 It is very important to be clear about the coordinate system (is z vertical up, or perpendicular to mean slope). Depending on this the calculation of overburden stress and Ne_ is different.
It seems the reviewer is referring actually to 4091, 10? We addressed this in a previous response by stating that the coordinate system is vertical.

4090, 11 longitudinal stretching cannot be simulated with SIA, also not by smoothing surface topography.
As described in the manuscript, we follow the approach of Kamb and Echelmeyer (1986), which demonstrates that this is, in fact, possible.

Fig 11 the symbols are too small.
We increased the size of the symbols on the plot and the font size of the stake location labels. The stakes are too close together to increase their symbol size.

Fig A1 , A2, A3: What do we see here? I see mountains with some snow-covered areas. It would be very helpful to mark the glacier outlines with red lines.
Good suggestion, we added rough outlines and labels of each glacier.

Response to A Vieli comments

I am a bit critical about the method and consequently the results regarding the quantification of basal sliding, in particular in relation to ice deformation and I think the derived ratios are subject to very large uncertainties that should be better discussed. The above 90% sliding to ice deformation ratio seems to me a very high estimate and could well be lower. I briefly outline my points below: 1. Ice deformation is highly dependent (linearly) on the rate factor A which itself is (for isotropic ice) dependent on ice temperature, water content and impurities and is in general not that well know. Even for ice at 0 degrees (temperate ice) literature values vary by a factor of 2 (higher than used here, see also Paterson) and impurities and high water content (probably to expect for a relatively warm and moist climate regime) may lead to even higher rate factors. This means the ice deformation could easily be a factor 2 to maybe 3 bigger which results in substantially lower sliding ratios (factor 2-3 higher ice deformation). I agree that the chosen value for A is probably the best guess but it is not in stone.

This is a good point and we have taken your suggestion and reran the deformation model for the maximum and minimum thicknesses and the maximum realistic ice softness parameter. Actually, even accounting for the maximum uncertainty of +-25% thickness and an ice softness parameter 2x higher, the sliding percentage for Nisqually glacier is still above 90% because the sliding is so much greater than the deformation in most places that even a several-fold increase in deformation doesn't change the percentages much. When we perform a similar test for Emmons glacier, however, the sliding contribution can be as low as 60%, so this was a valuable addition to the paper.

2. Bed topography and therefore ice thickness are not that well known (as clearly stated on p. 4092 line 1-2) which potentially impacts very strongly on the inferred ice deformation velocities. In particular in areas without radioechosounding data, which I assume includes that fast flowing areas of ice falls, thicknesses are interpolated and may well be off by more than the given +/-11m RMSE. Even if we assume just 11m uncertainty in thickness for this relatively thin glacier of 30m to 80m we get thickness uncertainties of 25% to 12% which (due to the non-linearity between ice flow and thickness) result in and over- or under-estimation of ice flow by a factor 5 (30m) to 1.8 (80). I guess for the thin ice fall regions uncertainties in ice thickness likely will be higher, and as the ice is thin there it will turn into even higher uncertainties in ice flow estimates (more than factor 5). This means the calculated velocities due to ice deformation and in particular the spatial variations will be strongly affected by uncertainties in bed topography and consequently weaken the conclusions on basal sliding and its spatial patterns.
See response to previous comment.

3. Further the used DEM is from 2008 and thinning (in average) from 2003-2011 is 8m. Has this been taken into account? If not, thicknesses to calculate flow may in places well be overestimated by about 4m which actually overestimate ice flow due to deformation (which is in favour of the conclusion of flow dominated by sliding) between a factor of 2 (for 30m) and 1.3 (for 80m).
This is lumped in the uncertainty of thickness uncertainty of 25%. Given the uncertainties involved with the bed and deformation model, we feel that using the 2008 surface is appropriate. We added a clarifying sentence.

4. The approach to calculate velocity fields for ice deformation (using the shallow-iceapproximation) is also questionable, in particular in areas of large changes in surface (bed) gradients such as around ice falls. The spatial smoothing (Echelmeyer method) certainly improves results compared to pure SIA, but I still think large uncertainties remain which are currently just assumed to be basal sliding (residuals packed into basal sliding). I agree that not too much modelling effort should be done if the bed (and ice thickness) are not well known, but in such a case maybe one should rather not try to derive accurate basal sliding rates at all and keep the modelling and interpretation on sliding simple.

Thus, overall the basal sliding analysis/modelling part (and its spatial variation) seems to suffer from over-interpretation in particular regarding the large uncertainties attached to the modelling. I would expect a less narrow consideration of these modelling results (% in sliding ratios) and that modelling uncertainties related to flow parameters, model choice and geometry data are taken into account and communicated. This would actually strengthen the case. Rather than exact sliding ratios, tendencies could be communicated in the conclusions Doing a modelling inversion is hard and certainly was time consuming but I think the details (peff and exact sliding %) currently do not add that much. Maybe the modelling part can be simplified and reduced as the outcomes are due to the large uncertainties rather speculative.
As described in responses to reviewer #1, the sliding model was removed and we estimated uncertainties on the deformation model.

Specific comments Abstract lines 12+13: I am a bit critical about these sliding ratio numbers, the method behind and think there are very high uncertainties attached to these numbers (could well be smaller: : :).

We estimated uncertainties for sliding % and report those in the abstract as well as elsewhere in the paper, in addition to our best estimates. The updated sentence in the abstract reads: ". Simple 2D ice flow modeling using TRI velocities suggests that sliding accounts for 91% and 99% of the July velocity field for the Emmons and Nisqually glaciers with possible ranges of 60 - 97% and 93 - 99.5%, respectively, considering ice thickness and ice softness uncertainties."

p. 4068 line 25: this is a very general statement but the references refer to the very specific glaciers of this study.
Yes, but they also happened to be studies with point sparse measurements, so they are used as examples here.

p. 4069 line 16: rather a remark: excuse me my ignorance but I was initially surprised about this statement of 'among best studied glaciers', as I did not know much about them. After reading the paper I agree that they are well researched but maybe 'beststudied' is another league.
These glaciers have a very long history of continuous and on-going study (led now by the National Park Service), and are very well-studied compared to most glaciers, but we tone this statement down a little since this isn't an important point and we don't want it to distract. It now says "Though Rainier's glaciers are among the best-studied alpine glaciers in the U.S…."

p. 4073, line 2: but before (introduction 1min minimum repeat intervals are mentioned and later for this study 3min are chosen (and as far as I know 1min is minimum given by the gamma-make used here). So why not mention thes actully used intervals od 3 min.
We changed this to: "The interval between acquisitions can be as short as ~1 min."

p. 4074, line 1: but I guess snow compaction was not measured the targeted glacier surface, so my questions is if this snow compaction can really be ignored.
We mean under the instrument, not on the glaciers, as implied by the context of the previous sentence, but we clarified this point anyway.

p. 4075, line 5-6: I do not quite follow this what 'interpolated result' is meant here
This is explained in the previous sentence ("we interpolated apparent displacement values over static control surfaces…"), but we do agree that the sentence wording here is a little confusing so we clarified this in the text.

p. 4075, line 7-8: maybe this stacking needs to be explained a bit further, for non-TRI experts this is maybe not clear.
Stacking is a pretty standard concept in many fields (e.g., seismology), but for additional clarification, we added "To stack, we take all the images for a given time period and compute the mean or median at each pixel, this has the effect of augmenting signal and canceling out noise. The median is less affected by outliers and is our preferred result."

p. 4076, line 22: specify here from when DEM is: ': : :an existing DEM from 2008 to…
We specified the 2007/2008 DEM.

p. 4081 section 4.4 and figure 8: I think here this comparison of velocities could quantified better by just comparing absolute line of sight (LOS) values (project all data in LOS direction). The figure is useful as a visual comparison but maybe a comparison of summary measures (Mean, SDT,: : :) would be useful.
We added a summary table, Table 4, and changed this section to say "In general, the velocity magnitudes are similar, with the overall mean of the Walkup et al. (2013) measurements slightly higher on average but often falling between the 7 July and 2 November GPRI magnitudes, as would be expected of a mean

velocity spanning approximately the same period. The velocity directions are also relatively consistent, with a median difference of 12°."

p. 4082 line 15: interesting this increase in velocity from July to winter at the ice fall and certainkly good to discuss this. But maybe worth saying that it is a 'slight' increase. To be positive, I think even if velocity do not change there this is interesting.
We added the slight qualifier to this sentence.

p. 4082 line 18-20: a note following on the point just above: according the the kinematic wave theory applied for glaciers (Nye 1961, 1963, 1965, also in Vanderveen book Fundamentals of Glacier Dynamics 2nd edition, p301ff)) the along-flow propagation of changes in thickness/flux is related to flow speed and the inverse of slope, which implies changes in ice thickness/speed struggle to propagate over steep ice falls. Although this paper does not deal with thickness change
This is an interesting note, and is certainly consistent with our observations. Since we do not have thickness change data, we would prefer to avoid speculation about flux variations. We will keep this point in mind as we pursue future studies of simultaneous velocity and elevation change data.

p. 4083, line 5: just a note: given the large diurnal variation in air temperature (and potentially atmospheric conditions I am quite surprised that the interferometric results are not affected more by atmosphere. I guess the stacking and corrections take care of that.
We agree.

p. 4084, section 5.3 flow modelling: if the modelling remains a central part of the analysis I would move the brief model description (with a celar and early reference to the details in the appendix) already in the METHODS section.
This change was made.

p. 4084 line 17/18: it is crucial to refer to the Appendix here for model details (at the end of this section is in my mind too late) and I would specify here what ratefactor (A) is used e.g. ': : :using an ice rheology corresponding to temperate ice (see Appendix: : :). This is crucial as firstly the choice of A introduces relatively large uncertainties (which should be communicated) (see also main comments).
We made this change.

p. 4084 line 24: how is 'weak' spatial dependence done? Is it partly a consequence of the length coupling (weighting) of the ice deformation calculation. If such a peff inversion has been done (although I think given the data available this may overdo (see main comments)) I would be interested to see the resulting peff variations with space. Or is it basically spatially constant, then I guess such an inversion dos not add too much anyway.
The sliding model was removed.

p. 4085 line 8: based on the given data (and modelling analysis) I do not quite agree with this conclusion of almost all flow by basal sliding. The uncertainties from rate factor, bed topography (thickness), etc. are pretty high (several fold) (as explained in detail in main comments), so these sliding ratios could well be quite different (in both directions but with a tendency to be rather smaller). Thus, I would not take these sliding % numbers as too narrow. Certainly, the uncertainties in these numbers should be discussed and communicated and maybe to conclusions be softened up a bit (e.- g. according to this modelling analysis, flow is likely to be dominated by basal sliding). Similar for the spatial variations in sliding I would be a bit more vague, the uncertainties in bed topography and type of model used will for some areas likely dominate the signal.
As explained in responses to earlier comments, we now provide a possible range of sliding % based on the uncertainties in ice thickness and ice softness.

p. 4086 line 1: again, the poor fit may well point to the large uncertainties in the modelling approach (parameter, model, datasets,: : :).
We removed the sliding model.

p. 4086 lines 11: I would rather say ': : :are consistent with: : :' or ': : : can likely be attributed with : : :' as apart from velocity chages there are virtually no further data supporting this claim. Most of the discussion on related basal hydrology changes are based on general understanding from elsewhere. Although I welcome an integration into the general/existing understanding I think the discussion and interpretation could maybe rely a bit more and clearer on collected data/evidence. Maybe in this paragraph the inversed peff (if it really is useful) could be linked in as well.
We added the word likely, however, it is hard to come up with other explanations for such a large seasonal change in velocity.

p. 4087 line 23-25: again if Neff is really inverted and shows something, I would like to see it here (and how it varies in space).
The sliding model was removed.

p. 4088 line 11-12: near the tongue the decrease in velocity is simply because the glacier retreated (and at the terminus it should be close to zero!!!).
This is already reflected in the existing text at the end of section 5.5.

p. 4089 line 14-15: again I struggle with these very narrow sliding ration numbers, maybe soften the numbers a bit, take into account uncertainties and use a more vague formulation (tendencies).
We now take into account uncertainties when computing the sliding percentage as explained in response to earlier comments.

p. 4091: lines 12 : : :: an assessment of uncertainties in A on U_deformation would be useful: : :
We now consider that A can be up to twice as high and use this to estimate deformation uncertainties.

Figure Fig. 1: the dark green for the arrows is not an ideal color choice, appears almost as black, maybe change color to something more distinct.
The arrows are the only arrows on the plot so we didn't think changing the color was necessary but removed the word "green".

Fig. 3: caption: change to ': : : slope-parallel TRI velocity for: : :'
Added "derived from TRI"

Fig. 4: the legend/colorbars here are very small that I could hardly read the numbers, actually similar for other figures (9/10).
We increased the size of these items.

Fig. 6: it would be nice to have some idea about uncertainties of these velocity data. I agree that the graph should not be cluttered too much but maybe a rough uncertainty bar somewhere would help, or simply put it in text in caption. Should it for the profile location not refer to Fig 4 instead of Fig. 5 in the caption?
We feel that this would make the plot too cluttered. Uncertainties are clearly shown on Fig. 5 and also summarized for the each study period on Figure 2. We added this note to the caption. Good catch on the incorrect reference figure for the profile line. This was fixed.

Observations of seasonal and diurnal glacier velocities at Mount Rainier, Washington using terrestrial radar interferometry

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Abstract

We present ~~spatially continuous~~surface velocity maps ~~using~~derived from repeat terrestrial radar interferometry (TRI) measurements, and use these time series to examine seasonal and diurnal dynamics of alpine glaciers at Mount Rainier, Washington. We show that the Nisqually and Emmons glaciers have small slope-parallel velocities near the summit (<0.2 m/day), high velocities over their upper and central regions (1.0-1.5 m/day), and stagnant debris-covered regions near the terminus (<0.05 m/day). Velocity uncertainties are as low as ±0.02-0.08 m/day. We document a large seasonal velocity decrease of 0.2-0.7 m/day (-25 to -50%) from July to November for most of the Nisqually glacier, excluding the icefall, suggesting significant seasonal subglacial water storage under most of the glacier. We did not detect diurnal variability above the noise level. ~~Preliminary~~Simple 2D ice flow modeling using TRI velocities suggests that sliding accounts for ~~roughly~~ 91% and 99% of the July velocity field for the Emmons and Nisqually glaciers~~,~~, with possible ranges of 60 - 97% and 93 - 99.5%, respectively, when considering model uncertainty. We validate our observations against recent in situ velocity measurements and examine the long-term evolution of Nisqually glacier dynamics through comparisons with historical velocity data. This study shows that repeat TRI measurements with >10 km range can be used to investigate spatial and temporal variability of alpine glacier dynamics over large areas, including hazardous and inaccessible areas.

1 Introduction

Direct observations of alpine glacier velocity can help improve our understanding of ice dynamics. Alpine glacier surface velocities are typically dominated by basal sliding, which is tightly coupled to subglacial hydrology (Anderson et al., ~~2004; Bartholomous~~2014; Bartholomaus et al. 2008). However, the spatial extent and spatial/temporal resolution of direct velocity measurements are often limited to short campaigns with sparse point measurements in accessible regions (e.g. Hodge, 1974; Driedger and Kennard, 1986). Remote sensing can help overcome many of these limitations. Radar interferometry, a form of active remote sensing, detects mm- to cm- scale displacements between successive images of the same scene and can see through clouds and fog. In the past few decades, satellite interferometric synthetic aperture radar, or InSAR (e.g. Massonnet and Feigl, 1998; Burgmann et al., 2000) has emerged as an invaluable tool for quantifying glacier dynamics (e.g., Joughin et al, 2010). However, limited data availability and revisit times limit the application of InSAR for the study of many short-term processes.

Terrestrial radar interferometry (TRI), also referred to as ground-based radar interferometry, has recently emerged as a powerful technique for observing glacier displacement that is not prone to the same limitations (Caduff et al., 2014). Sets of radar data acquired at intervals as ~~low~~short as ~1 minute~~ intervals~~ from up to several km away allow for observations of velocity changes over short timescales and large spatial extents~~, while stacking~~. Stacking these large numbers of interferogram pairs over longer timescales can significantly reduce noise. Here, we employ this relatively new technique to provide spatially- and temporally-continuous surface velocity observations for several glaciers at Mount Rainier volcano in Washington State (Fig. 1). Though Rainier's glaciers are among the best-studied alpine glaciers ~~on earth~~in the U.S. (Heliker et al., 1984; Nylen, 2004), there are many open questions about diurnal and seasonal dynamics that TRI can help address. Specifically, many aspects of subglacial hydrology and its effects on basal sliding are poorly constrained, especially for inaccessible locations like the Nisqually icefall and ice cliff. ~~In this study we gain~~Our observations provide new insight ~~on the evolution of this hydrological network~~into these processes through analysis of the relative magnitude and spatial patterns of surface velocity over diurnal and seasonal timescales. To our knowledge, no other studies have investigated seasonal changes to glacier dynamics using TRI.

Mount Rainier offers an excellent setting for TRI, with several accessible viewpoints offering a near-continuous view with ideal line-of-sight vectors for multiple glaciers, and well-distributed static bedrock exposures for calibration. The ability to image the velocity field of entire glaciers from one viewpoint with minimal shadowing sets this study area apart. Most previous studies only image part of the glaciers under investigation, usually due to less favorable viewing geometries (e.g. Noferini et al., 2009; Voytenko et al., 2015; Riesen et al., 2011). However, the steep topography and local climatic factors at Mount Rainier result in strong atmospheric variability and turbulence – a major source of noise for radar interferometry techniques (Goldstein, 1995). Atmospheric noise is a particular issue for the long ranges (>10 km) associated with accessible viewpoints at Mount Rainier. ~~We~~To overcome this limitation, we successfully combine, expand on, and evaluate noise reduction techniques such as stacking interferograms (e.g. Voytenko et al. 2015) and ~~using stable rock points to fit and subtract~~deriving atmospheric noise corrections over static control surfaces (bedrock exposures) (e.g. Noferini et al. 2009). We demonstrate ~~the~~that these techniques offer significant uncertainty reductions ~~in velocity uncertainties these techniques yield with~~ using a ~~new~~novel bootstrapping approach ~~that does not require stable rock points (e.g. Voytenko et al., 2015).~~.

In the following sections, we provide background on Mount Rainier's glaciers, and detail our sampling methodology and data processing techniques. We then present TRI results documenting seasonal and diurnal velocity variations for the Nisqually, Wilson, and Emmons Glaciers, and quantify measurement uncertainty. Next we examine the partitioning of observed surface velocities between deformation and basal sliding at different times of year using a simple 2D flow model, and compare our observations to other recent and historical velocity measurements. These comparisons provide ground truth for TRI measurements and new insight into the evolution of the Nisqually glacier since the late 1960s.

2 Study area

With a summit elevation of 4392 m, Mount Rainier (Fig. 1) is the largest stratovolcano in the Cascades and is considered the most dangerous volcano in the United States (Swanson et al., 1992). It also holds the largest concentration of glacial ice in the mainland United States (Driedger and Kennard, 1986~~):~~) - 87 km^2 ~~is~~was covered with perennial snow and ice ~~as of~~in 2008 (Sisson et al., 2011). The steep upper sections of the major glaciers are relatively thin, with typical thicknesses of ~30 to 80 m (Driedger and Kennard, 1986). Thickness increases at lower elevations, with a maximum of ~200 m for the Carbon glacier, although these estimates likely provide an upper bound, as these glaciers have experienced significant thinning in recent decades, losing 14% of their volume between 1970 and 2008 (Sisson et al., 2011). Mass balance stake measurements from 2003-2010 show that the average winter balance for Nisqually was 2.4 m water equivalent (m.w.e.), average summer balance was -3.5 m.w.e., and cumulative net balance was -8.6 m.w.e. from 2003-2011 (Riedel, 2010; Riedel and Larrabee, 2015).

The glaciers of Mt. Rainier have been of interest to geoscientists for over 150 years and have a long record of scientific observation (Heliker, 1984). In this study, we focus on large, accessible, well-documented glaciers in the park: the Nisqually and Wilson glaciers on the southern flank, and Emmons and Upper Winthrop glaciers on the northeastern flank (Fig. 1).

The Nisqually Glacier is visible from several viewpoints near the Paradise Visitor Center, which is accessible year-round. The terminus location has been measured annually since 1918, and three transverse surface elevation profiles have been measured nearly every year since 1931 (Heliker, 1984). Veatch (1969) documented a 24-year history of Nisqually's advances and retreats and other dynamic changes through a meticulous photographic survey from 1941-1965. Hodge (1974) conducted a detailed 2-year field study of the seasonal velocity cycle for the lower Nisqually. He found that velocities varied seasonally by about 50%, with maximum velocities in the spring (June) and minimum in the fall (November). This finding, and the lack of correlation between runoff and sliding speeds, advanced the idea that efficient conduits close as meltwater input decreases in the fall, leading to distributed subglacial storage through the fall, winter and spring. Increased surface melting in spring and summer leads to increased subglacial discharge and the opening of a more efficient network of conduits capable of releasing some of this stored water (Hodge, 1974). More recently, Walkup et al. (2013) tracked the movements of supraglacial rocks with high precision from 2011-2012, yielding velocity vectors for a wide network of points over the lower parts of Nisqually glacier.

The Emmons glacier, visible from the Sunrise Visitors Center, has received less attention than Nisqually, despite the fact that it is the largest glacier by area on the mountain ~~in terms of area~~ (Driedger and Kennard, 1986), mainly because it is not as easily accessible as Nisqually. A large rock fall (~1.1x10^7 m^3) from Little Tahoma in December 1963 covered much of the lower

Emmons glacier with a thick debris layer (Crandell and Fahnestock, 1965). The insulating debris cover likely contributed to the advance and thickening of the Emmons Glacier from 1970-2008, while all other glaciers on Mount Rainier experienced significant thinning (Sisson et al., 2011). Average 2003-2010 winter balance for Emmons was 2.3 m.w.e., average summer balance was -3.2 m.w.e , and cumulative net balance was -7.7 m.w.e. from 2003-2011 (Riedel, 2010; Riedel and Larrabee, 2015).

The National Park Service's long-term monitoring protocols include both the Nisqually and Emmons glaciers and involve regular photographs, annual mass balance measurements, meltwater discharge rates, plus area and volume change estimates every decade (Riedel, 2010; Riedel and Larrabee, 2015).

3 Methods

3.1 Instrument description

~~In~~For this study, we ~~employed the~~used a GAMMA portable radar interferometer (GPRI) (Werner et al., 2008; Werner et al., 2012) - a ground-based, frequency-modulated continuous waveform (FMCW) radar that can capture mm-scale surface displacements. The instrument includes three 2-m antennas mounted on a vertical truss, with one transmit antenna 35 cm above the upper of two receiving antennas, spaced 25 cm apart (Fig. 2). The transmit antenna produces a 35° vertical beam with 0.4° width that azimuthally sweeps across the scene to build a 2D radar image as the truss rotates. The radar operates at a center frequency of 17.2 GHz, with selectable chirp length of 2-8 ms and bandwidth of 25 to 200 MHz. The radar wavelength is ~~0.0176 m~~17.6 mm with range resolution of ~0.75 cm and one-way interferometric change sensitivity of 8.7 mm/cycle of phase providing <1 mm line-of-sight precision. Line-of-sight interferograms are generated by comparing phase differences in successive acquisitions from the same viewpoint. ~~Short repeat acquisition intervals of tens of seconds to minutes ensure high coherence~~The interval between acquisitions can be as short as ~1 min, allowing for high coherence even in rapidly changing scenes.

3.2 Survey Description

We performed four data collection campaigns in 2012 (Table 1). The first campaign occurred on 6-7 July 2012. This timing corresponds to just after the expected peak seasonal glacier velocities at Mount Rainier (Hodge, 1974). Following the success of this study, three subsequent deployments were performed during the late fall and early winter, which should capture near-minimum seasonal velocity (Hodge, 1974). These campaigns were timed to occur before, immediately after, and a few weeks after the first heavy snowfall of the season (2 Nov 2012, 27 Nov 2012 and 10 Dec 2012, respectively).

Three viewpoints were selected for data collection: GLPEEK and ROI, which overlook the Nisqually and Wilson glaciers, and SUNRIZ, which overlooks the Emmons and upper Winthrop glaciers (Fig. 1). ROI and SUNRIZ were directly accessible from park roads, which greatly facilitated instrument deployment, and GLPEEK was accessed on foot. ROI was occupied during all campaigns, while SUNRIZ and GLPEEK were only occupied during the July 2012 campaign because of access limitations. Figures A1-A3 show the field of view from each viewpoint.

Distances from the GPRI to the summit were 6.7, 7.6, and 10.8 km from GLPEEK, ROI and SUNRIZ, respectively. Radar images were continuously collected with a 3-minute interval for all surveys. Total acquisition time at each site was dictated by logistics (weather conditions, personnel), with ~24 hour acquisitions at SUNRIZ and ROI to capture diurnal variability.

The instrument was deployed on packed snow during the 6 July 2012 GLPEEK and 27 Nov and 10 Dec 2012 ROI acquisitions. Over the course of the GLPEEK survey, we noted limited snow compaction and melt beneath the GPRI tripod with total displacement of ~2-4 cm over ~6 hours. However, this instrument motion proved to be negligible for the interferogram interval used (6 min). ~~No~~We did not note significant snow compaction ~~was noted~~under the tripod during the fall/winter surveys.

Weather conditions during the July 2012 surveys were clear with light/variable wind. The 2 Nov 2012 survey involved high-altitude clouds, passing showers and brief interruptions in data collection. Weather conditions were clear with sun for the 27 Nov 2012 campaign, and fog with limited visibility on 10 Dec 2012.

3.3 Data Processing

All radar data were processed with the GAMMA SAR and Interferometry software suite. ~~Multi-look intensity (mli) products were generated from original single-look complex (slc) data by~~

~~averaging 15 samples in the range direction.~~ Interferograms were generated from ~~the mli~~single-look complex SLC products with a time separation of 6 minutes, though sometimes longer if acquisition was interrupted ~~(for~~. For example images~~,~~ see Fig. A4~~)~~.. Interferograms were multi-looked by 15 samples in the range direction to reduce noise. A correlation threshold filter of 0.7 and an adaptive bandpass filter (ADF) with default GAMMA parameters were applied to the interferograms to improve phase unwrapping. Phase unwrapping was initiated in areas with high correlation scores and negligible deformation, such as exposed bedrock or stagnant ice.

3.3.1 Atmospheric noise corrections

Slight changes in the dielectric properties of the atmosphere between the GPRI and target surfaces can lead to uncertainty in the interferometric displacement measurements (Zebker et al. 1997; Werner et al., 2008). Changes in atmospheric humidity, temperature, and pressure can all affect radar propagation velocity~~, which determines the two-way travel time~~ (Goldstein, 1995). These variations are manifested as phase offsets in the received radar signal, which must be isolated from phase offsets related to true surface displacements.

This atmospheric noise proved to be significant for the long range ~~(~~~~~~(i.e., ~22 km two-way horizontal path at SUNRIZ), mountainous terrain ~~(~~~~~~(i.e., ~2.4 km vertical path from SUNRIZ to summit), and turbulent atmosphere involved with this study, with the magnitude of this noise often exceeding that of surface displacement signals. ~~In~~The scale of the atmospheric noise features we observed in the data was typically much wider than the width of the glaciers, so in order to minimize this atmospheric noise in the individual interferograms, we interpolated apparent displacement values over static control surfaces (e.g. exposed bedrock~~) via Delauney~~). To do this, we fit a surface using Delaney triangulation~~. A~~ to a subset (5%) of pixels over exposed bedrock. The subset of pixels was resampled randomly for each unwrapped interferogram and the interpolated result was smoothed to reduce artifacts~~. These~~ and then subtracted from the interferogram. The corrections were applied to all individual interferograms, and the resulting products were stacked to further reduce noise. To stack, we took all the images for a given time period and computed the mean and median at each pixel. This has the effect of augmenting signal and canceling out noise. The median is less affected by outliers and is our preferred result. The median line-of-sight (LOS) velocities from this stack provide a single measurement with a high signal to noise ratio for the entire sampling period.

In addition to computing ~~these campaign~~the median LOS velocities for the entirety of each sampling period, we also computed a running mean of the LOS velocities to characterize any short-term velocity variations in the extended occupation datasets: 7-8 July SUNRIZ (24 hours) and 1-2 November ROI (21 hours). The running mean was computed every 0.3 hours with a 2-hour centered (acausal) window, with standard error used to estimate uncertainty.

Interferograms with significant phase unwrapping errors, low ~~correlations~~correlation, or anomalous noise were excluded from stacking. ~~Only~~We only excluded a few images ~~were deleted~~ for each site ~~for these reasons~~ with the exception of SUNRIZ, which produced many images with anomalous noise and unwrapping errors, possibly due to instrument noise and/or the extended range through significant atmospheric disturbance. For this reason, more than half of the data from SUNRIZ was excluded from the analysis (Table 2). For GLPEEK and ROI, interferograms with occasional localized unwrapping errors were preserved during stacking, as they have little influence on the final stack median. However, localized areas with persistent unwrapping errors in the SUNRIZ data were masked using a threshold standard deviation filter of 0.6 m/day.

We estimated median LOS velocity uncertainties using a bootstrapping approach (Efron, 1979). This involved resampling the set of images used in the stack with replacement 1000 times for each campaign. Then, for each pixel, the 25th and the 975th ordered values were set as the lower and upper bounds of the 95% confidence interval.

3.3.2 Conversion from radar coordinates to map coordinates

~~For campaign intercomparison, we~~We developed a sensor model and tools to terrain-correct the stacked GPRI data (in original azimuth, range coordinates) using an existing 2 m/pixel airborne ~~2008~~ LiDAR digital elevation model (DEM) ~~for Mount Rainier~~ acquired in September 2007/2008 (Robinson et al., 2010). While some elevation change has undoubtedly occurred for glacier surfaces between September 2008 and July 2012, the magnitude of these changes (<20 m) is negligible for orthorectification purposes given the GPRI acquisition geometry. A single control point identified over exposed bedrock in the LiDAR DEM and the multi-look image (mli) radar data was used to constrain absolute azimuth orientation information for each campaign. A ~10 m/pixel (mean of azimuth and range sample size) grid in UTM 10N (EPSG: 32610) was created for each campaign, with extent computed from the GPRI GPS coordinates, min/max range values, and min/max absolute azimuth values. Each 3D pixel in this grid was then populated by extracting the radar sample with corresponding range and azimuth.

3.3.3 Correction to slope-parallel velocities

While the line-of-sight vectors for these surveys are roughly aligned with surface displacement vectors (median incidence angles for glacier surfaces are ~22° for GLPEEK, ~25° for SUNRIZ and ~26° for ROI), glaciological analyses typically require horizontal and vertical velocity components relative to the glacier surface. As each GPRI survey offers only a single look direction, this is not possible. However, we can assume that displacement is dominated by surface-parallel flow, and use ~~an existing~~the 2007/2008 LiDAR DEM to extract surface slopes needed to estimate 3D displacement vectors (e.g., Joughin et al., 1998).

This approach is intended for relatively smooth, continuous surface slopes over length scales >2-3x ice thickness. It is therefore possible that the slope-parallel correction can overestimate velocity for steep, high relief surfaces with significant high-frequency topographic variability (e.g. icefalls). The slope-parallel assumption also begins to break down where the vertical flow velocity component becomes significant. This is expected in the upper accumulation and lower ablation zones, where the submergence and emergence velocities become more significant, respectively, but is less important near the equilibrium-line altitude (ELA) or locations where sliding dominates surface motion. The latter is expected for much of the Nisqually Glacier at least (Hodge, 1974).

We implement a slope-parallel correction by first downsampling the 2007/2008 LiDAR DEM to 20 m/pixel and smoothing with a 15x15 pixel (~300 m), 5-sigma Gaussian filter. The slope-parallel velocity (V_S) is defined as:

$$V_S = V_L / (\hat{S} \cdot \hat{L}) \qquad (1)$$

where $\hat{S} \cdot \hat{L}$ is the dot product between the unit vector pointing directly downslope from each grid cell ($\hat{S}$) and the unit vector pointing from each grid cell to the sensor ($\hat{L}$). Regions where the angle between these two vectors exceeded 80° were masked to avoid dividing by numbers close to zero which could amplify noise.

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3.4 2-D glacier deformation modeling

Surface flow velocity can be partitioned into internal deformation and basal sliding components. We present a simple, 2-D plane-strain ice deformation model for a preliminary assessment of the importance of basal sliding for the glaciers in our study area. The deformation model uses the shallow ice approximation (SIA) – an approximate solution of the Stokes Equations (Greve and Blatter, 2009; Cuffey and Paterson, 2010). The expected surface velocity u_s due to internal deformation from the SIA model is:

$$u_s = \frac{2\,A(s\quad(\alpha)\rho_i g)^n\, H^{n+1}}{n+1} \qquad (2)$$

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where ρ_i represents ice density, g represents gravitational acceleration, α represents local surface slope, H represents local ice thickness, A represents an ice softness parameter and n represents a flow rate exponent. The coordinate system is vertically aligned.

The SIA is not well-suited for narrow mountain glaciers, so we modify it to simulate the effect of non-local conditions, such as lateral sidewall drag and longitudinal stretching. The ice thickness H and surface slope α are smoothed using a weighting function based on Kamb and Echelmeyer (1986). Kamb and Echelmeyer (1986) calculated a longitudinal coupling length l using a 1-D force balance approach, for each point in their domain. They calculated l to be in the range of one-to-three ice thicknesses for valley glaciers. We simplified this by using a single value for l over the domain of model. The longitudinal couple length l is used in a weighting function to smooth α and H. The weighting function has the form:

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

where x and y represent the horizontal coordinates of the weight position, and x' and y' represent the horizontal coordinates of the reference position. Weights are calculated at each point in the model domain, over a square reference window (side length of A_w). H and α are smoothed at the reference position by normalizing weights over the reference window. We choose a coupling length l of ~1.5 ice thickness and an averaging window size of ~3 ice thicknesses, consistent with the usage in Kamb and Echelmeyer (1986). We use a spatially uniform and temporally constant ice softness parameter suitable for ice at the pressure melting point of 2.4×10^{-24} Pa^{-3} s^{-1} (Cuffey and Paterson, 2010, pg. 75). Ice softness can be affected by several factors (e.g., englacial water content and impurities), so we also consider an ice softness parameter up to twice this best estimate in accounting for model uncertainties, as described below. Our best estimates of model input parameters are summarized in Table 3.

Surface slope (Fig. A1B) was estimated from the 2007/2008 LiDAR DEM (Robinson et al. 2010). Surface velocities u_s are the TRI-derived median slope-parallel velocities. Ice thicknesses H (Fig. A1A) were estimated by differencing the 2007/2008 LiDAR DEM surface elevations and the digitized and interpolated bed topography from Driedger and Kennard (1986). The Driedger and Kennard (1986) bed topography contours were derived from ice-penetrating radar point measurements and surface contours from aerial photographs. The published basal contours for Nisqually/Wilson, Emmons, and Winthrop Glaciers were digitized and interpolated to produce a gridded bed surface using the ArcGIS Topo to Raster utility. The gridded bed elevations have root mean squared error (RMSE) of 11 m when compared with the 57 original radar point measurements. A point-to-plane iterative closest point algorithm (implemented in the NASA Ames Stereo Pipeline pc_align utility (Shean et al., 2015)) was used to coregister the 1986 bed topography to the 2007/2008 LiDAR topography over exposed bedrock on valley walls. Mean error over these surfaces was 7.6 m following coregistration, although some of this error can be attributed to actual surface evolution near glacier margins (e.g., hillslope processes) from 1986-2008. In addition to these interpolation and coregistration errors, there were likely small changes in ice thickness during the 4-5 years between the 2007/2008 DEM data collection and the 2012 TRI observations, as mass balance measurements suggest that both the Nisqually and Emmons Glaciers experienced net mass loss during this time period (Riedel and Larrabee, 2015). Propagation of these uncertainties results in estimated ice thickness uncertainties of ~5-25%. In order to account for this large uncertainty, we ran the model with ±25% ice thickness as well as 2x ice softness in order to estimate the possible range of expected deformation velocities.

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

4 Results

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

Stacking alone was very effective; the velocities of the mean and median stacks with and without the atmospheric noise correction were very similar. The main benefit of the extra step of using stable rock points to subtract an estimate of the atmospheric noise was to significantly reduce the uncertainties and to reduce the noise where velocities are slow. The uncertainties before and after atmospheric correction are compared on Table 2. The median width of the 95% confidence interval for each corrected, stacked pixel is plotted in Fig. 3B and Fig. 5. Note near-zero values over exposed bedrock surfaces used to derive atmospheric noise correction. We were able to reduce uncertainties (half the median confidence interval width) to about ±0.02 to ±0.08 m/day over glacier surfaces for some campaigns, with uncertainty dependent on the total number of stacked images, weather conditions, and target range (Table 2). For example, the 6 July 2012 ROI survey had a final confidence interval width of 0.11 m/day (~±0.06 m/day) while the 10 Dec 2012 ROI survey had a final confidence interval width of 0.15 m/day (~±0.08 m/day) despite a 50% increase in stack count. This is likely due to increased local atmospheric variability, as low-altitude clouds obscured the surface during 10 Dec 2012 survey. The 2 Nov 2012 ROI survey had the highest stack count (359) with the lowest uncertainty values of ±0.02 m/day (Table 2).

4.1 July 2012 Surface Velocities

The 6-7 July 2012 observations show slope-parallel velocities that range from ~0.0-1.5 m/day for both the Nisqually and Emmons glaciers (Fig. 3A, 4, 6). Both display high velocities over their upper and central regions that taper into essentially stagnant (<0.05 m/day) debris-covered regions near the terminus. In general, slope-parallel velocities near the summit are small (<0.2 m/day).

On the Nisqually Glacier, a series of local velocity maxima (>1.0 m/day) are associated with increased surface slopes between local surface highs. Local velocity maxima are also observed for the fast-flowing Nisqually icefall (western branch of Upper Nisqually, see Fig. 3) and above the Nisqually ice cliff (eastern branch). A relatively smooth velocity gradient from slow- to fast-moving ice is present upstream of the icefall, while the velocities above the ice cliff display a steep velocity gradient (Fig. 3).

The main (south) branch of the Emmons glacier displays generally increasing velocity from the summit to lower elevations. A large high velocity region (>0.7-1.1 m/day) is present over central Emmons, downstream of the confluence of upper branches. These elevated velocities decrease at lower elevations, where ice thickness increases and surface slopes decrease (Fig. ~~B1~~A5). A central ~~faster channel~~“core” of exposed ice ~~surrounded by slow or stagnant~~displays slightly elevated velocities relative to surrounding debris-covered ice ~~decreases to below our detection limit as it approaches~~within ~1-1.5 km of the terminus.

Velocities exceed 1 m/day over the “central” branch of the Upper Emmons Glacier, where flow is restricted between two parallel bedrock ridges, with local maxima similar to Nisqually. Velocities at higher elevations within the “central” branch appear slower (<0.1-0.5 m/day), separated from the fast downstream velocities by a small area that was excluded due to phase unwrapping errors. Photographs show that this area appears heavily fractured with many large blocks indicative of rapid, discontinuous flow (Fig. A3).

Smaller, relatively thin glaciers, such as the Fryingpan, Upper Kautz, and Inter Glacier (labeled on Fig. 1), also display nonzero surface velocities of <0.1-0.2 m/day, but with limited spatial variability.

4.2 Seasonal variability

The repeat observations from the ROI viewpoint provide time series that capture seasonal velocity variability for the Nisqually, Wilson and Upper Kautz Glaciers. We observe significant velocity changes during the summer to winter transition and more subtle changes within the winter period. These changes are shown in map view on Fig. 4 and in profile view with corresponding slope and ice thickness on Fig. 6.

These data show a velocity decrease of 0.2-0.7 m/day (-25 to -50%) from July to November 2012 for most of the Nisqually Glacier. This includes central and lower Nisqually and the ice above the ice cliff. The greatest velocity decreases are observed near the crest and lee of surface rises (downstream of data gaps from radar shadows, Fig. 4), where some of the highest velocities were observed in July. In contrast, the area immediately downstream of the ice cliff and the area surrounding the icefall both display an apparent velocity increase for the same time period (Fig. 4, 6). While the increase is less than the 95% confidence interval for most areas, we can confidently state that the icefall and area below the ice cliff do not display the significant decrease in velocity observed elsewhere.

The majority of the Wilson Glacier displays a similar ~0.3-0.7 m/day (-40 to -60%) velocity decrease from July to November. Interestingly, the steep transition where the Wilson merges with the Nisqually displays an apparent velocity increase of ~0.1 m/day during this time period (Fig. 4). These data also reveal subtle velocity increases in the debris-covered ice near the Nisqually terminus and the Upper Kautz glacier (Fig. 4), though these increases are statistically insignificant.

The repeat winter observations of Nisqually show relatively constant velocities with some notable variability. Analysis of the 2 Nov. to 10 Dec. observations reveals a statistically significant -0.1 m/day (-50%) velocity decrease ~1 km upstream of the terminus (centered on ~0.7 km in Fig. 6A profile), a +0.1 to +0.2 m/day (+20 to +30%) increase over central Nisqually centered on ~3.5 km in the Fig. 6D profile, and an apparent +0.2 m/day (+130%) increase over the Upper Wilson. In the latter case, the 10 Dec. velocities are actually higher than those observed in July. The slowdown over lower Nisqually appears robust, but other trends have amplitudes that are mostly below the 95% confidence interval for the 27 Nov. and 10 Dec. observational campaigns (Fig. 4).

4.3 Diurnal variability

We collected ~21 and ~24 hour time series for the Emmons and Nisqually/Wilson Glaciers (Table 1) in July and November, respectively~~.~~, and look at changes throughout the day. Although uncertainties are large, we present the time series on Fig. 7.

In general, velocities for these regions remain relatively constant during their respective sampling periods. The Emmons time series shows an apparent decrease in velocity over the central, fast-flowing regions (B, C, D in Fig. 7A) from ~18:00 to 21:00 local time, and an apparent increase between ~07:00 to 09:00 local time (Fig. 7A). The Nisqually time series shows an apparent decrease from ~06:00 to 11:00 local time for the icefall and ice cliff, and an apparent decrease for several areas of the glaciers followed by an increase (Fig. 7B). However, uncertainties are large and none of these are statistically significant.

4.4 Comparison with independent velocity measurements

We now ~~validate~~compare our TRI results ~~against~~with independent velocity measurements for ~~the same~~an overlapping time period. Walkup et al. (2013) performed repeat total station surveys to document the location of sparse supraglacial cobbles and boulders on the lower Nisqually glacier from 2011-2012. While measurement errors ~~such as~~(e.g., cobble rolling~~ and ~~/sliding) for these observations are difficult to document, the large sample size and relatively long measurement intervals allow for accurate surface velocity estimates.

Figure 8 shows average velocity vectors measured by Walkup et al. (2013) for the period between 19 July and 11 October 2012, with corresponding surface-parallel velocity vectors from the 7 July and 2 November TRI surveys. This comparison is summarized on Table 4. In general, the velocity magnitudes ~~appear~~are similar, with the overall mean of the Walkup et al. (2013) measurements slightly higher on average, but often falling between the 7 July and 2 November GPRI ~~magnitude~~magnitudes, as would be expected of a mean velocity spanning approximately the same period. The velocity directions are also relatively consistent, with a median angular difference of 12°. The greatest ~~deviation~~deviations are observed near the ice margins and over small-scale local topography (e.g. ice-cored moraine near western margin), where surface-parallel flow assumptions break down. In general, the two techniques provide similar results and offer complementary data validation. However, since the Walkup et al. (2013) measurements were limited to accessible areas, they cannot be used to validate TRI observations for heavily crevassed areas, icefalls, and other hazardous dynamic areas generally higher on the mountain.

4.5 2-D flow modeling

Figure 9 shows modeled deformation, sliding velocity residual (observations - deformation model), and sliding percent (sliding velocity residual as percentage of total velocity) with best estimate model parameters for Nisqually glacier in July and November. Figure 10 shows corresponding output for Emmons. The SIA deformation models suggest that most areas of both glaciers are moving almost entirely by sliding. The modeled glacier deformation alone is unable to account for the observed surface velocity during any of the observation periods. Only a median of 1% of the velocity field over the Nisqually glacier area can be explained by internal deformation in July, and only 2% in November. If we consider ±25% ice thickness and up to 2x the ice softness, the possible range of the median deformation contribution is still small, 0.5 – 7% in July and 0.5 – 8 % in November. If we consider only ±25% ice thickness and do not change the ice softness, the range narrows to 0.5 – 4% in both cases. Using stake measurements, Hodge (1974) estimated deformation contributed ~5-20% of the velocity for the upper third of the ablation area of the Nisqually glacier. He did not study any areas above the equilibrium line, so to compare directly to Hodge's (1974) numbers, we take the median deformation percentage over approximately the upper third of the ablation area and find a best estimate of 1% (range 0.3 – 5%) for July and 2% (range 0.5 – 7%) for November. These numbers suggest that sliding is even more dominant than Hodge (1974) estimated in this area, though it is difficult to say if the differences are real (i.e. sliding was higher in 2012 than it was five decades ago) or just due to differences in methods and assumptions.

The model results for Emmons suggest that deformation is more important for the Emmons glacier than for Nisqually. A median of 9% of the July velocity field of Emmons can be explained by deformation, with a possible range of 3 – 40% when considering ±25% ice thickness and up to 2x the ice softness. If we consider only ±25% ice thickness, the range narrows to 3 – 20%.

There are a few regions where the observed surface velocity can be explained entirely or nearly entirely by internal deformation. These include the area within ~1-2 km of the Nisqually and Emmons Glacier terminus, where ice is relatively thick and observed velocities are small.

5. Discussion

The continuous coverage of the TRI provides information about the spatial distribution of surface velocities and strain rates. Several local velocity maxima are apparent along the centerline of the Nisqually glacier and the central branch of the Emmons glacier. These velocity maxima are associated with surface crevasses and increased surface slopes, with peak velocities typically observed just upstream of peak slope values (Fig. 6). They are likely related to accelerated flow downstream from local bedrock highs,

However, the local velocity maxima at ~2.1 km in Fig. 6 corresponds to a region of decreased surface slopes and increased ice thickness. This location also displayed significant seasonal velocity change, ~~suggesting that~~which could be related to variations in local subglacial hydrology ~~could be a controlling factor here~~(e.g. reservoir drainage) during this time period.

5.1 Icefall and ice cliff dynamics

Terrestrial radar interferometry offers new observations over dynamic, inaccessible areas that have received limited attention in previous studies (e.g., icefalls, ice cliffs). For example, the velocities above the Nisqually ice cliff display an abrupt transition from slow- to fast-moving ice (Fig. 4). The high strain rates associated with this transition are suggestive of crevasse opening and “detached slab” behavior rather than continuous flow, which is reflected in the heavily crevassed surface at this location.

Our results show that the Nisqually icefall and the icefall at the convergence of the Wilson and Nisqually glaciers ~~appear to~~show a slight increase in velocity from July to the winter months. This suggests that the icefalls may not be susceptible to the same processes that caused the seasonal velocity decrease over much of the rest of the glacier. This may indicate that there is a lack of local continuity through icefalls, which appears to prevent or dampen propagation of downstream seasonal velocity decreases. It could also indicate that the icefall is relatively well~~-~~-drained year-round, and is not significantly affected by seasonal changes in subglacial hydrology. ~~Another~~A potential explanation for the observed minor increase in velocity could be early winter snow accumulation on blocks within the icefall.

Interestingly, in contrast to the icefall, the hanging glacier above the Nisqually ice cliff displayed a significant velocity decrease from July to November, despite similar steep surface slopes and crevasse density. This could potentially be related to the lack of backstress from downstream ice and an increased sensitivity to minor fluctuations in subglacial hydrology. Hanging glaciers are also thought to be the source of some of the repeating glacial earthquakes that are triggered by snow loading (Allstadt and Malone, 2014), which highlights their sensitivity to minor perturbations.

5.2 Lack of significant diurnal variability

We expected to see significant variability over the 24-hour July time series for Emmons, as atmospheric temperatures varied from 16°C to 27°C at Paradise Visitors Center (~1600 m.a.s.l), and skies remained cloud-free during data collection. We hypothesized that the resulting increase in meltwater input from late morning through late afternoon might produce an observable increase in sliding velocity. While the results potentially show a slight velocity decrease at higher elevations overnight, and a slight velocity increase in the morning (Fig. 7A, A-D), these changes are not statistically significant, nor coincident with times expected to have highest melt input. The lack of a significant diurnal speedup suggests that the subglacial conduits are relatively mature by July, and are capable of accommodating the diurnal variations in meltwater flux without affecting basal sliding rates.

We did not expect to see significant diurnal changes in the 21-hour November time series for Nisqually (Fig. 7A), as atmospheric temperatures ranged between 2°C and 6°C at Paradise Visitors Center (~1600 m a.s.l.) and skies were partly-cloudy to overcast during data collection, so surface meltwater input should have been minimal. Our results show only a minor velocity decrease higher on the glacier in the morning hours but it is not statistically significant and does not occur at times when we would expect increased meltwater.

Though some of the subtle changes in the extended time series may reflect actual diurnal velocity variability, we cannot interpret these with confidence. This suggests that the magnitude of diurnal variability, if it exists, during these time periods is minor when compared to the observed seasonal changes. It also implies that other stacks derived from a subset of the day can be considered representative of the daily mean, and can be compared for seasonal analysis.

5.3 ~~2D Flow Modeling~~

~~Surface ice flow can be partitioned into internal deformation and basal sliding components. We present simple, 2-D plan-view ice flow and sliding models constrained by the TRI slope-parallel velocity observations assess the importance of basal sliding for the glaciers in our study area and to connect changes in the observed flow field to the temporal and spatial evolution of subglacial~~

~~hydrology. Sophisticated ice flow models (e.g. Gagliardini et al., 2013; Le Meur et al., 2004; Zwinger et al., 2007) can capture complete 3-D stress and velocity fields, but we do not consider these options due to residual uncertainty in ice thickness (~5-25%) and bedrock topography, as described in Appendix B. With improved bedrock topography and ice thickness data, it might be possible to use these more sophisticated models in the future to invert observations of surface velocity for basal shear stress. This would offer a much more accurate picture of the spatial and temporal variability of glacier sliding and thus illuminate the changes in subglacial hydrology. However, this is beyond the scope of this study so here we implement simple ice flow models to provide preliminary estimates of deformation and sliding contributions and leave more sophisticated modeling to future work.~~

~~The deformation model uses the shallow ice approximation (SIA) (Cuffey and Paterson, 2010). The SIA is not well-suited for narrow mountain glaciers, so we modify it to simulate the effect of non-local conditions, such as lateral sidewall drag and longitudinal stretching using a weighting function based on Kamb and Echelmeyer (1986). The sliding model uses a Budd type sliding parameterization (Bindschadler, 1983) where the minimization technique solves for two parameters *k* and N_{eff}. *k* is a temporally constant and spatially uniform coefficient that represents a coupling of the glacier to its bed, which could be affected by bedrock geology, sediment layer thickness, size, and distribution. N_{eff} is the effective pressure, and is allowed to vary temporally with weak spatial dependence. We adjust *k* and N_{eff} to produce a sliding velocity model that most closely matches the residual velocity (observed TRI velocity minus the modeled deformation velocity). A detailed description of the modeling methods is presented in Appendix B, model input parameters are summarized on Table 3.~~

~~Figure 9 shows modeled deformation, sliding velocity residual (observations – deformation model), percentage sliding (sliding velocity residual as percentage of total velocity), modeled sliding velocity, and sliding velocity model mismatch for Nisqually glacier in July and November. Figure 10 shows corresponding output for Emmons. Table 4 summarizes the best fitting values of *k* and N_{eff} for each survey period.~~

~~The SIA deformation models suggest that most areas of both glaciers are moving almost entirely by sliding. The modeled glacier deformation alone is unable to account for the observed surface velocity during any of the observation periods. Only a median of 1% area of the velocity field~~

~~over the Nisqually glacier area can be explained by internal deformation in July, and only 2% in November. These numbers are similar to those calculated by Hodge (1974), who estimated ~5-20% sliding for the upper third of the ablation area and 1-3% at the equilibrium line, though the numbers are not directly comparable because Hodge (1974) did not include any area above the equilibrium line. The results for Emmons suggest that that sliding is slightly less important for the Emmons glacier, where a median of 9% of the velocity field can be explained by deformation.~~

~~However, there are a few regions where the observed surface velocity can be explained entirely or nearly entirely by internal deformation such as the relatively flat downstream portions of the Emmons glacier and the terminus of the Nisqually glacier where ice is thick, surface slopes are small, and observed velocities can be explained largely by deformation.~~

~~The sliding models, on the other hand, offer a poor fit for the observations. The sliding model residual (Fig. 9E, 10E) provides a measure for goodness of fit, with 28% and 19% of the glacier area fitting the observations within ±0.1 m/day for winter and summer models, respectively. If the residual threshold is increased to ±0.2 m/day, the model accounts for 46% and 60% of the summer and winter models, respectively.~~

~~Despite doing a poor job of fitting the data, the simple sliding models provide some useful insights. The pattern of the misfit is informative: in general, the sliding models over-predict sliding velocities near glacier margins and under-predict sliding velocities toward glacier centerlines. More importantly, the misfit suggests that there is significant spatial variation in the sliding behavior of these alpine glaciers that cannot be captured by sliding models with homogeneous bed properties and subglacial water pressure. Collecting better bed topography and ice thickness data and implementing more sophisticated models would yield better results but is beyond the scope of the current study.~~

~~5.~~4 Seasonal velocity changes

The observed seasonal velocity changes from July to November can likely be attributed to changes in glacier sliding, which in turn are driven by evolving englacial and subglacial hydrology (Fountain and Walder, 1998). During the spring-summer months, runoff from precipitation (i.e. rain) and surface snow/ice melt enters surface crevasses, moulins, and/or conduits near the glacier margins. This water travels through a series of englacial fractures, reservoirs and conduits, and eventually ends up in a subglacial network of channels and reservoirs between the ice and bed. Storage time and discharge rates within the subglacial system are variable, with water finally exiting the system through one or more proglacial streams at the terminus. This dynamic system is continuously evolving due to variable input, storage capacity, and output. In early July, ongoing snowmelt should produce high meltwater discharge that travels through a relatively efficient network of mature conduits. As discharge decreases later in the summer, these subglacial conduits/reservoirs close due to ice creep without high flow to keep them open through melting due to heat from viscous dissipation. By November, there should be little or no surface meltwater input and we would expect to see a minimum in basal sliding velocity (Hodge, 1974). This is consistent with the observed velocity decrease in Fig. 4. However, the deformation modeling results (Fig. 9) show that a significant sliding component is still present for most of the Nisqually glacier in November and December, when minimum surface velocities are expected.

The spatial patterns of the velocity change observed between July and November can be used to infer the extent of basal sliding. This may provide some insight into subglacial water storage, since the deformation component of surface velocity should remain nearly the same year-round. Fig. 4 indicates that almost the entire Nisqually glacier slows down significantly between July and November, suggesting that storage is occurring under most of the glacier below the icefall and ice cliff. Significant velocity decreases are observed near local surface rises (Fig. 4), where some of the highest velocities were observed in July. This suggests that there are likely subglacial cavities downstream of these areas with high basal water pressures that can support enhanced sliding during the summer.

Hodge (1974) interpreted a delay in both the maximum summer velocity and minimum winter velocity between the terminus and ELA as a propagating “seasonal wave” traveling ~55 m/day. While our sampling is limited, the continued November 2 to November 27 slowdown over the lower Nisqually near the terminus (Fig. 4F) could represent a delayed response to the significant slowdown over central Nisqually. This might be expected, as surface velocities near the terminus are dominated by internal deformation and should respond more slowly than areas dominated by basal sliding.

~~The seasonal velocity changes are reflected in the sliding models, with best-fit sliding parameters (Table 4) indicative of seasonal changes in subglacial hydrology. The Nisqually Glacier experienced more sliding in July 2012 relative to November-December 2012, which corresponds to an increase in effective pressure N_{eff}. This change in N_{eff} is likely due to the lack of meltwater input into the subglacial hydrological system during the fall-winter.~~

5.~~5~~4 Comparison with historical velocity measurements

As described earlier, Hodge (1972, 1974) measured surface velocity for a network of centerline stakes on the lower Nisqually from 1968-1970. He documented a significant seasonal cycle with minimum velocities in November and maximum velocities in June.

To put our velocity data in historical context, we digitized Hodge's (1972) July and November 1969 surface velocity data at 19 stake locations along a profile of the lower half of the Nisqually glacier. We then sampled the 2012 TRI slope-parallel velocities at these locations (Fig. 11). Remarkably, in spite of significant terminus retreat of up to ~360 m and surface elevation changes of approximately -20 m (Sisson et al., 2011), the November 1969 and November 2012 surface velocities are almost identical at stakes 12-20, suggesting that bed properties and local geometry ~~rather~~have greater influence over sliding velocity than ice thickness ~~is a controlling factor~~or relative distance from the terminus. In contrast, the July 2012 velocities at stakes 12-20 are 8-33% faster than the July 1969 velocities. The ice is mostly sliding at these locations, so the change could be related to a difference in the timing of the peak summer velocities, or potentially enhanced sliding in 2012. The nearly identical surface velocities in November 1969 and 2012 suggests that the discrepancy between Hodge's sliding percentage estimates and our estimates (section 4.5) is likely related to different methodology and assumptions rather than actual changes in sliding since 1969.

The most notable difference between the profiles is observed closer to the terminus at stakes 7-12. At these locations, the July and November 2012 velocities are both <0.05 m/day, whereas July and November 1969 velocities are ~0.2 and ~0.1 m/day, respectively, with significant seasonal variability. This suggests that the ice near the present-day terminus is essentially stagnant and no longer strongly ~~affected~~influenced by changes in subglacial ~~water~~hydrology.

6. Conclusions

In this study, we used repeat TRI measurements to document spatially continuous velocities ~~of~~for the Emmons and Nisqually glaciers at Mount Rainier, WA. We produced surface velocity maps that reveal speeds of >1.0-1.5 m/day over the upper and central regions of these glaciers, <0.2 m/day near the summit, and <0.05 m/day over the stagnant ice near their termini. Novel data processing techniques reduced uncertainties to ±0.02-0.08 m/day, and the corrected, surface-parallel TRI velocities for Nisqually display similar magnitude and direction with a set of sparse interannual velocity measurements (Walkup et al., 2013).

Repeat surveys show that Nisqually glacier surface velocities display significant seasonal variability. Most of the glacier experienced a ~25-50% velocity decrease (up to -0.7 m/day) between July and November. ~~Interestingly, steep icefalls display no velocity change or even a slight velocity increase over the same time period.~~ These seasonal variations are most likely related to changes in basal sliding and subglacial water storage. Interestingly, the steep icefall displays no velocity change or even a slight velocity increase over the same time period. We documented no statistically significant diurnal velocity variations in ~24-hour datasets for Nisqually and Emmons, suggesting that subglacial networks efficiently handled diurnal meltwater input.

~~Using a simple 2D ice flow model, we estimate that sliding is responsible for approximately 91-99% of the observed velocity for Emmons and Nisqually glaciers, similar to sliding percentage estimates for Nisqually glacier in 1969 (Hodge, 1974).~~ Comparisons with 1969 velocity measurements over the Lower Nisqually (Hodge, 1972; 1974) reveal similar ~~2012 and 1969~~ November velocities in both 2012 and 1969, and faster ~~2012~~ July velocities in 2012.

Using a simple 2D ice flow model, we estimate that basal sliding is responsible for most of the observed surface velocity signal except in a few areas, mainly near the termini. The model suggests that about 99% of the July velocity field for the Nisqually glacier is due to sliding. Even when we account for the large uncertainties in ice thickness and ice softness, the possible range of sliding percentage is still narrow: 93 – 99.5% Deformation is more important for the Emmons glacier, where we estimate 91% of the observed motion is due to sliding, with a much wider possible range of 60 – 97% when accounting for uncertainties.

In summary, TRI presents a powerful new tool for the study of alpine glacier dynamics ~~remotely~~. With just a few hours of fieldwork for each survey, we were able to document the dynamics of several glaciers at Mount Rainier in unprecedented extent and detail from up to 10 km away. TRI is ~~an ideal technique~~particularly well suited for examining diurnal and seasonal glacier dynamics, ~~particularly~~especially for areas that are difficult to access directly (e.g., icefalls), like many parts of the glaciers at Mount Rainier. Repeat surveys provide precise surface displacement measurements with unprecedented spatial and temporal resolution, offering new insight into complex processes involving subglacial hydrology and basal sliding. Future studies involving coordinated, multi-day TRI occupations during critical seasonal transition periods could undoubtedly provide new insight into these and other important aspects of alpine glaciology.

Figure Captions

Figure 1. Glaciers at Mount Rainier and locations of viewpoints used for ground based radar interferometry. Instrument view angle ranges are indicated by ~~green~~ arrows extending away from each viewpoint location. Boxes A-C show zoom areas for later figures. Inset map shows regional location of Mount Rainier. Glacier outlines in this and subsequent figures are from Robinson et al. (2010).

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

Figure 3. A) Median slope-parallel velocity ~~for timeseries~~derived from TRI for GLPEEK and SUNRIZ viewpoints taken on July 6-7, 2011. B) Width of 95% confidence interval (high minus low limits for slope parallel flow field) of slope parallel velocities for July 6-7, 2011 computed by bootstrapping after performing atmospheric noise corrections and stacking. Area shown is indicated by Box A on Fig. 1.

Figure 4. A-D) Median slope-parallel velocities for Nisqually and Wilson glaciers for four different time periods taken from ROI viewpoint. Dashed lines on top left panel show locations of profiles taken to create Fig. 6, markers indicate distance in km. E-G) Percent change in median slope-parallel velocity for the Nisqually and Wilson glaciers between time periods. Blue indicates a velocity decrease and red indicates a velocity increase relative to the earlier time

period, gray polygons indicate areas where velocity change is significant with 95% confidence. Area shown is indicated by Box B on Fig. 1.

Figure 5. Width of 95% confidence interval (high minus low limits for slope parallel ~~flow field)~~velocity) over Nisqually glacier computed by bootstrapping. Shown for four sampling periods from the ROI viewpoint. Note that the color bar is scaled differently than Fig. 3B.

Figure 6. A and C: Slope parallel velocity profiles along the two branches of Nisqually glacier (profile lines shown in map view on Fig. ~~5~~4A) for all sample time periods and viewpoints. B and D: Surface slope and ice thickness along each profile line. Surface slope is smoothed identically to that used for slope parallel corrections (see text), ice thicknesses are estimated from digitized basal contours from Driedger and Kennard (1986) and surface elevations from the 2007/2008 LiDAR (Robinson et al., 2010). Refer to Figure 5 and Table 2 for uncertainty estimates.

Figure 7. LOS ~~velocities computed using a sliding mean with a 2-hour centered window~~velocity time series for areas ~~of the glaciers indicated~~outlined on ~~the median flow field~~ maps ~~at~~to the right. ~~Error margins, shown as transparent polygons~~Shaded region around each line ~~represent~~represents ± one standard error for a 2-hour running mean. a) 24-hour timeseries at SUNRIZ on July 7-8, 2012, gray box indicates the period with poor data quality (see text for details). b) 22-hour timeseries at ROI on Nov 1-2, 2012.

Figure 8. Comparison of average azimuth and velocities measured by Walkup et al. (2013) between 19 July 2012 and 11 October 2012 (black) compared to TRI slope-parallel velocities derived from this study at the same locations for two time periods that bracket the time period measured by Walkup et al. (2013). ~~Area shown is indicated by~~See Table 4 for comparison statistics and Box C on Fig. 1 for context.

Figure 9. ~~Modeling~~Model results for summer (6 July 2012) and a late fall (2 Nov 2012) time period for Nisqually and Wilson glaciers. A, ~~F) Deformation model~~D) Modeled surface velocity for internal deformation, B, ~~G~~E) Sliding residual (observed slope parallel ~~velocities from observations~~velocity minus the modeled deformation ~~model~~velocity), C, ~~H~~F) Estimate of the sliding percentage ~~sliding~~ (sliding residual divided by total slope~~-~~-parallel ~~velocities), D, I) Sliding model, E, J) Difference between sliding residual and sliding model.~~velocity).

Figure 10. Same as Fig. 9 but for Emmons ~~glacier~~Glacier.

Figure 11. ~~Surface~~July and November 1969 surface velocities measured~~ in 1969~~ by Hodge (1974, digitized from Hodge, 1972) at 19 stake locations along lower Nisqually profile (circles), compared with ~~the~~sampled 2012 slope~~-~~-parallel velocities ~~derived from this study~~ for ~~the same times of year at the same~~corresponding locations/seasons (triangles). Stake locations are labeled and indicated with dotted lines and are shown in map view at right~~, area shown is~~ (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.

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

Figure A4. ~~Samples of pair~~Pair of multi-look ~~images (mli~~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 ~~resulting~~multi-looked interferogram generated from the SLC images (right~~) taken from ROI viewpoint.~~).

Figure ~~B1~~A5. A) Filtered ice thickness and B) filtered slope used as ~~modeling~~model inputs.

Appendix A

Appendix A contains supplementary figures.

~~Appendix B~~

~~The model uses the shallow ice approximation (SIA) – an approximate solution of the Navier-Stokes Equations (Cuffey and Paterson, 2010). The surface velocity u_s resulting from internal deformation given from the SIA is~~

$$\cancel{u_s = \frac{2A(s(\alpha)\rho_i g)^n H^{n+1}}{n+1} \qquad (B1)}$$

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~~where ρ_i represents ice density, g represents gravitational acceleration, α represents local surface slope, H represents local ice thickness, A represents an ice softness parameter and n represents a flow rate exponent.~~

~~However, the SIA only takes into account local conditions to determine the deformation component of the surface velocity. To simulate the effect of non-local conditions, such as lateral sidewall drag and longitudinal stretching, the ice thickness H and surface slope α are smoothed using a weighting function based on Kamb and Echelmeyer (1986). Kamb and Echelmeyer (1986) calculated a longitudinal coupling length l using a 1-D force balance approach, for each point in their domain. They calculated l to be in the range of one-to-three ice thicknesses for valley glaciers. We simplified this by using a single value for l over the domain of model. The longitudinal couple length l is used in a weighting function to smooth α and H. The weighting function has the form:~~

$$W(x,y) = e^{-\frac{\sqrt{(x-x')^2+(y-y')^2}}{l}} \qquad \text{(B2)}$$

~~where x and y represent the horizontal coordinates of the weight position, and x' and y' represent the horizontal coordinates of the reference position. Weights are calculated at each point in the model domain, over a square reference window (side length of A_w). H and α are smoothed at the reference position by normalizing weights over the reference window. We choose a coupling length l of ~1.5 ice thickness and an averaging window size of ~3 ice thicknesses, consistent with the usage in Kamb and Echelmeyer (1986). Model input parameters are summarized in Table 3.~~

~~In order to model the basal sliding component of motion, we use a Budd type sliding parameterization (Bindschadler, 1983). The sliding parameterization has the form~~

$$u_s = k\frac{\tau_d^m}{N_e} \qquad \text{(B3)}$$

~~where k represents a coefficient that approximates the coupling strength of the glacier to the underlying bed, τ_d represents the gravitational driving stress $\tau_d = s\,(\alpha)\rho_i g$, m represents an adjustable flow rate exponent, N_{eff} represents the effective pressure at the base of the glacier and ρ_i represents ice density. Effective pressure is the difference of water from overburden ice pressure $N_e = p_i - p_w$ where ice pressure $p_i = \rho_i g$ water pressure $p_w = \rho_w g\ _w$, H_{wt} represents the height of the water table above the bed and ρ_w represents water density.~~

~~We use a spatially uniform and temporally constant ice softness parameter suitable for ice at the pressure melting point of 2.4×10^{-24} Pa^{-3} s^{-1} (Cuffey and Paterson, 2010, pg. 75). Surface slope (Fig. B1B) was estimated from the 2008 LiDAR DEM (Robinson et al. 2010). Surface velocities u_s are the median slope-parallel velocities. Ice thicknesses H (Fig. B1A) were estimated by differencing the 2008 LiDAR DEM surface elevations and the digitized and interpolated bed topography from Driedger and Kennard (1986). The Driedger and Kennard (1986) bed topography contours were derived from ice-penetrating radar point measurements and surface contours from aerial photographs. The published basal contours for Nisqually/Wilson, Emmons, and Winthrop Glaciers were digitized and interpolated to produce a gridded bed surface using the ArcGIS Topo to Raster utility. The gridded bed elevations have root mean squared error (RMSE) of 11 m when compared with the 57 original radar point measurements. A point-to-plane iterative closest point algorithm (implemented in the NASA Ames Stereo Pipeline pc_align utility (Shean et al., 2015) was used to coregister the 1986 bed topography to the 2008 LiDAR topography over exposed bedrock on valley walls. Mean error over these surfaces was 7.6 m following coregistration, although some of this error can be attributed to actual surface evolution near glacier margins (e.g., hillslope processes) from 1986-2008. The accumulation of these uncertainties results in highly uncertain ice thickness estimates.~~

~~The sliding model minimization technique solves for two parameters, a coupling coefficient k, and effective pressure N_{eff}. We chose to force k to be both temporally constant and spatially uniform, but we allow N_{eff} to be temporally variable with a weak spatial dependence. k represents a coupling of the glacier to its bed, which could be affected by bedrock geology, sediment layer thickness, size, and distribution. We chose not to explicitly include any of those processes because they are poorly constrained, they are not expected to vary much on timescales of days to months, nor is their effect on the sliding relationship well understood. N_{eff} can change rapidly over the course of months and possibly days as subglacial water storage is strongly affected by seasonal melt cycles, we therefore choose to allow it vary temporally. N_{eff} is weakly spatially dependent in our model because we don't allow N_{eff} be greater than local ice pressure, if that were the case it would imply an unphysical, negative water pressure. We prohibit N_{eff} from being greater than ice pressure at any particular location.~~

~~We adjust k and N_{eff} to produce a sliding velocity model that most closely matches the residual velocity (observed velocity minus the modeled deformation velocity). To do this, we calculate a~~

~~model mismatch, the difference between the model sliding velocity and the residual velocity, at each point, choosing parameters that minimize the summed squares of the mismatch over the domain.~~

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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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°

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)

Table 3 Constants used in modeling analysis

Name	Symbol	Value	Units
Ice softness parameter	A	2.4×10^{-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	ρ_i	900	$kg\ m^{-3}$
Density of water	ρ_w	1000	$kg\ m^{-3}$

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