

Response to the Interactive comment on

“Recent dynamic changes on Fleming Glacier after the disintegration of Wordie Ice Shelf, Antarctic Peninsula”

by Peter Friedl et al.

Anonymous Referee #1

Received and published: 13 July 2017

We would like to thank the reviewer again for the constructive and helpful comments on our manuscript.

All comments have been considered and a list of responses and changes in the manuscript is given below. Responses are written in bold face type and changes in the manuscript are written in blue.

Friedl et al. present a study on glacier retreat and changes in ice flow for Antarctic Peninsula outlet glaciers after disintegration of Wordie Ice Shelf. The work is based on analysis of remote sensing data from various sources. It extends the period of observations on frontal retreat, flow acceleration and glacier thinning that was reported by Rignot et al. (2005) and Wendt et al. (2010) up to the year 2009. Of particular interest is the production of a close time series of surface velocities, up to the year 2016, including the filling of gaps from previous years. The analysis of surface elevation change, comparing the change 2011 to 2014 to that of previous years, and the time series of frontal retreat are also very relevant for describing glacier behaviour. The review on previous work and the presentation and interpretation of the observations are presented in coherent manner at large. However, there are various individual points that are not well explained or questionable. The presented data sets are very useful for characterizing the glacier behaviour during the last two decades, but the discussion and conclusions focus on the description of the observed phenomena and do not provide any substantial new insights into the processes leading to the observed changes.

We have largely restructured and reformulated the Discussion and the Conclusion sections. In particular we have included a discussion of recently published data on atmospheric driven CDW upwelling events and elevation change in Wordie Bay. By linking these findings to our results on high resolution surface velocity change, ice thinning and grounding line retreat, we are able to provide new insights in the timing of the processes and reasons for the observed changes in Wordie Bay.

Specific Comments:

P1, L16: The conclusion on “pronounced basal melt at the grounding line” is not based on any direct observations for these glaciers.

We understand the reviewer’s concerns. However, Depoorter et al. (2013) and Rignot et al. (2013) observed high basal melt rates for the remaining parts of the ice shelf, similar to those found for ice shelves in the Amundsen Sea sector (see answer P2, L13, L14). Furthermore the coincident timing of exceptional warm water intrusions in 2008/2009 and 2010/2011 into Marguerite Bay found by Walker and Gardner (2017) and the two acceleration phases found by us indicates that ocean warming and the observed changes in glacier dynamics are strongly correlated. Anomalous low sea ice extent in Marguerite Bay observed in 2008 and 2010 suggests that the warming events affected the upper water column, but Walker and Gardner (2017) found that the years 2009-2011 had the highest temperatures also at depths of 400 m. Hence, it is very likely that submarine ice melting was increased during phases of strong CDW upwelling events and that this has triggered unpinning and grounding line retreat. We rewrote the abstract in such a way that it points out that our conclusion on basal melt at the grounding line is not based on direct observations made in this study but on reasonable assumptions upon existing evidence. We also discuss this in more detail in the restructured Discussion and Conclusion parts.

P1, L19: The length of the centreline (for which this values is valid) should be specified

We thank the reviewer for this suggestion. We have changed the sentence into:

The resulting loss in buttressing is able to explain a remarkable median speedup of $\sim 1.3 \text{ m d}^{-1}$ ($\sim 30 \%$) between 2007 and 2011 observed along a centreline extending between the grounding line in 1996 and $\sim 16 \text{ km}$ upstream.

P1, L21: Fig. 5 shows in the downstream part of the profiles for 2011–2014 thinning rates that are smaller than for 2004 to 2008. 60% to 70% higher rates are only evident for a subsection of the profile shown in Fig. 5.

Despite of lower ice thinning rates measured towards the ice front in 2011–2014, our data show an overall increase of median ice thinning rates of $\sim 1.1\text{--}1.3 \text{ m a}^{-1}$ or $\sim 70 \%$ between the periods from 2004 to 2008 and from 2011 to 2014. Median elevation change rates were calculated for entire profiles, including subsections with lower and increased ice thinning rates. Hence the numbers refer to full profiles not just to the upstream parts. In some upstream parts ice thinning rates increased by even more than 100%. We have changed the corresponding section in chapter 5.2 in such a way that we hope it makes this now clearer to the reader:

Despite of lower ice thinning rates measured towards the ice front in 2011–2014, our data show an overall median increase of ice thinning rates along the profiles of $\sim 1.1\text{--}1.3 \text{ m a}^{-1}$ or $\sim 70 \%$ between the periods from 2004 to 2008 and from 2011 to 2014. However, in some areas 10–15 km upstream, ice thinning rates even doubled in the latter period.

P1, L26: Hardly possible that the glaciers draining into Wordie Bay have “a huge potential for an increase in sea level rise”.

Our choice of words was probably too dramatic. We have changed the sentence into:

Hence, this endangers upstream ice masses, which can significantly increase the contribution of Fleming Glacier to sea level rise in the future.

P2, L4: 4.21 Gt/a refers to Larsen-A and Prince-Gustav-Channel glaciers, 2011–2013.

We apologize for this inaccuracy. We have changed the sentence into:

Rott et al. (2014) estimated the total dynamic ice mass loss for the glaciers along the Nordenskjöld Coast and the Sjögren-Boydell glaciers after ice shelf disintegration to be $4.21 \pm 0.37 \text{ Gt a}^{-1}$ between 2011–2013.

P2, L13, L14: Rignot et al. (2014), Suppl. Material, show for Wordie Ice Shelf clear dominance of calving losses compared to ablation, rather than “basal melt exceeding the ablation induced by calving”.

We agree with the reviewer. Indeed Fig. 1 in Rignot et al. (2013) indicates that mass loss from Wordie Ice Shelf is much bigger from calving than from basal melt. However, mass losses of George VI, Wilkins, Bach and Stange ice shelves are shown to be clearly dominated by basal melt. Tab. 1 in Rignot et al. (2013) reveals that mass loss at Wordie Ice Shelf is calculated to be $7.6 \pm 3 \text{ Gt/a}$ for calving and $6.5 \pm 3 \text{ Gt/a}$ for basal melt, which means that $\sim 54 \%$ of mass loss can be attributed to calving and $\sim 46 \%$ to basal melt. However, the calculated basal melt rate of $23.6 \pm 10 \text{ m/a}$ is the second largest value obtained for Antarctica in this study. Interestingly a similar study of Antarctic wide basal melt rates by Depoorter et al. (2013) attributes 82 % of the mass loss of Wordie Ice Shelf to basal melt and only 18 % to calving flux (Fig. 1, Depoorter et al., 2013). The values correspond to a calving flux of $2 \pm 3 \text{ Gt/a}$ and a mass loss of $10 \pm 4 \text{ Gt/a}$ (or $14.79 \pm 5.26 \text{ m/a}$ respectively) due to basal melt (Table 1 in Suppl. Material, Depoorter et al., 2013). The high basal melt rates at Wordie Ice Shelf are similar to those found for ice shelves in the Amundsen Sea sector (Table 1 in Suppl. Material and Fig. 2, Depoorter et al., 2013). The differences between the two studies seem to be mainly caused by the different datasets used for the computations (e.g. Rignot et al., 2013 used ice thickness from Bedmap-2 whereas Depoorter et al., 2013 calculated ice thickness from satellite borne surface elevation data). Despite of the differences, both studies show that Wordie Ice Shelf undergoes pronounced basal melt.

Consequently we have changed the section into:

For the south-western Antarctic Peninsula Rignot et al. (2013) demonstrated that basal melt of George VI, Wilkins, Bach and Stange Ice Shelves exceeded the ablation induced by calving. For Wordie Ice Shelf high basal melt rates of $23.6 \pm 10 \text{ m a}^{-1}$ and $14.79 \pm 5.26 \text{ m a}^{-1}$ have been reported by Rignot et al. (2013) and Depoorter et al. (2013) respectively. However, the presented melt ratios (i.e. the ratio between basal melt and the sum of calving flux and basal melt) differ between 46 % (Rignot et al., 2013) and 82 % (Depoorter et al., 2013).

P2, L15: Please explain the “small coastal atmospheric and oceanic processes”.

Padman et al. (2012) found that changes in basal melt at Wilkins Ice Shelf are primarily determined by small-scale coastal atmospheric and oceanic processes that can alter the depth of the cold winter water layer, rather than by large-scale atmospheric forcing of benthic inflows of warm CDW along troughs cutting across the continental shelf. Such small-scale processes include complex interactions between coastal wind stress, surface radiation balance, sea ice and freshwater fluxes (including land runoff, sea-ice production and melt, and feedbacks between the ice shelf and regional hydrography and circulation). For example, increased freshwater fluxes from runoff and ice-shelf basal melt consolidated by downwelling forced by enhanced wind stress can lead to a depression of the cold winter water layer and hence to a reduction of basal melt. For the sake of clarity we have changed the corresponding sentence into:

Wilkins Ice Shelf experienced amplified basal thinning controlled by small-scale coastal atmospheric and oceanic processes that assist ventilation of the sub-ice-shelf cavity by upper-ocean water masses (e.g. variations in wind stress or reduced freshwater fluxes from runoff and ice-shelf basal melt) until ~8 years before break-up events took place in 2008 and 2009 (Braun and Humbert, 2009; Padman et al., 2012).

P4, L4: Date and source for the Bedmap2 DEM section over the study area should be mentioned (may have some impact for geocoding and analysis of surface elevation change).

The Bedmap2 DEM is a combination of several DEMs covering all or part of Antarctica (Fig. 5 and Tab. 3, Fretwell et al., 2013). The multiple surface elevation datasets were gridded together into a seamless DEM of Antarctica. On the Antarctic Peninsula north of 70° S the Bedmap2 DEM is entirely based on the improved ASTER GDEM by Cook et al. (2012), providing a vertical accuracy of $\pm 26 \text{ m}$. The ASTER GDEM is compiled from stacked photogrammetric DEMs from ASTER scenes acquired between 2000 and 2009 that are unspecified in the final product. On the Antarctic Peninsula south of 70° S the Bedmap2 DEM consists of data from the SPIRIT DEM, compiled from SPOT images, acquired in 2007-2008 (vertical accuracy is $\pm 6 \text{ m}$) and from the NSIDC DEM (DiMarzio et al., 2007), derived from ICESat data acquired in 2003–2005 (vertical accuracy varies from $\pm 0.4 \text{ m}$ to $\pm 20 \text{ m}$).

We used the Bedmap2 DEM as a source of topographic information for orthorectification of the velocity fields and for the derivation of local incidence angles required for the conversion from slant to ground range displacement. Here consistency of the elevation data is more important than the date. Since the extent of our Sentinel-1 scenes exceeded 70° S (i.e. the extent of the improved ASTER GDEM by Cook et al., 2012), the coverage of the Bedmap2 DEM allowed us to use a single consistent topographic reference for all velocity products. This is important when comparing velocities obtained from the different sensors.

We would like to clarify that we did not use the Bedmap2 DEM for the calculation of elevation changes between the Bedmap2 DEM and the TanDEM-X DEMs. Of course in this case the dates of the data would have been important. Instead, we used the differential phase between the simulated phase of the Bedmap2-DEM and the topographic phase of TanDEM-X to support phase unwrapping. However in the revised version of the manuscript we now use a subset of the TanDEM-X Global DEM for this procedure, as described in answer P5, L34.

Notwithstanding the above, we of course agree with the reviewer that providing more information on the Bedmap2-DEM is an asset. We have therefore changed the corresponding section into:

The Bedmap2 digital elevation model (DEM) of Antarctica (Fretwell et al., 2013), resampled to 100 m resolution, was taken as a topographic reference for orthorectification of the surface velocity fields and for the derivation of local incidence angles required for the conversion from slant to ground range displacement. Over the Antarctic Peninsula the Bedmap2 DEM provides a seamless

compilation of data from the improved ASTER GDEM (from ASTER stereo images acquired between 2000 and 2009) (Cook et al., 2012), the SPIRIT DEM (from SPOT stereo images acquired in 2007 and 2008) and the NSIDC DEM (from ICESat data acquired between 2003 and 2005) (DiMarzio et al., 2007).

P4, L24: May mention here that the OIB and Huss bedrock data are compared in Fig. S5.

We have added a reference to Fig. S5 at the end of the corresponding sentence.

P5, L15: point clouds of differential elevation measurements are shown in Fig. 5. (not Fig. 4).

In Fig. 4 the locations of the measuring points are depicted which form the point clouds shown in Fig. 5. For more clarity we have changed the corresponding sentence into:

The locations of the resulting points of differential elevation measurements are shown in Fig. 4.

P5, L34: It is not meaningful using a low resolution DEM from a different epoch for performing DEM differencing with high resolution TanDEM-X data, if an up-to-date high resolution DEM from TanDEM-X is available.

As already mentioned in answer P4, L4 we originally subtracted a simulated topographic phase from the Bedmap2 DEM from the TanDEM-X phase prior to the unwrapping of the TanDEM-X phase and re-added the subtracted height-information afterwards. The fact that then only a differential phase has to be unwrapped, facilitates the phase unwrapping and makes it more robust. So the only purpose of subtracting the phase was facilitating the phase unwrapping rather than calculating elevation changes between Bedmap2 and TanDEM-X. Nevertheless, in the revised manuscript we now use a subset of the TanDEM-X global DEM with a spatial resolution of 12 m (Rizzoli et al., 2017) for the procedure. This is because we have also changed the vertical referencing in such a way that we now use the TanDEM-X global DEM to vertically adjust the TanDEM-X DEMs (see answer P6, L19 to L28).

We consequently have changed the corresponding sentence into:

A subset of the TanDEM-X global DEM, covering the two TSX/TDX-DEMs, was chosen to be the reference DEM.

We also have added a sentence to chapter 3 (Data):

A simulated phase from a subset of the TanDEM-X global DEM with a spatial resolution of 12 m (Rizzoli et al., 2017) was used to facilitate phase unwrapping during the generation of the two TSX/TDX DEMs. The TanDEM-X global DEM was also used as a reference for the absolute height adjustment of the TSX/TDX DEMs.

P6, L19 to L28: The differences between ATM and (uncorrected) TanDEM-X rates of elevation change (up to 2m/a) need further explanations and checks. 2m/a corresponds to 6 m difference for the 3 year time span. The area for the profile is located in the percolation zone. Typical values for TanDEM-X penetration bias in the percolation zone are about 4 m (see e.g. Wessel et al., ISPRS Annals. VL-III-7, doi:10.5194/isprsannals-III-7-9-2016). If the morphology of the snow and firn medium is the same on both dates, the penetration bias cancels out for DEM differencing. A difference in dh/dt of 6 m versus optical data can only be explained if the snow morphology is completely different on both dates (e.g. melting surface snow without penetration vs. completely frozen snow volume). This should show up clearly in the backscatter signatures.

We thank the reviewer for raising this important issue. We completely agree with the reviewer that a penetration bias of 6 m is pretty large for the percolation zone. We found that our approach of vertically referencing the TSX/TDX DEMs by calculating mean offsets over sea ice was not optimal. We therefore changed the vertical registration procedure in such a way that we now use a subset of the TanDEM-X global DEM at 12 m resolution as an absolute height reference. Before differencing, both TSX/TDX digital elevation models were vertically adjusted to the TanDEM-X global DEM according to their median offsets measured over stable ground. This improved the vertical registration and resulted in a more realistic penetration depth bias.

The maximum difference between ATM and uncorrected TDX/TSX elevation change rates was 1.25 m/a, corresponding to a 3.75 m difference for the 3 year time span. However, differences in elevation change were only measured in the lower areas of the glacier tongue, where surface melt occurs. In the upper areas (above 600 m altitude) the difference between ATM and

TSX/TDX elevation change rates was close to 0 m. Here the medium was likely completely frozen on both dates of acquisition, so that the penetration bias cancelled out. A comparison of backscatter values showed that in areas below 600 m altitude the backscatter of the 2014 acquisition was lower than the backscatter of the 2011 acquisition. In the upper areas above 600 m, however, the backscatter values were similar.

We have updated the TSX/TDX elevation change rates in the text and all figures that contain TSX/TDX elevation change data (i.e. Fig. 4, Fig. 5 b, Fig. S4 a-c and Fig. S6.1-4).

We have also added a Figure S4 d showing the differences in backscatter between both acquisition dates.

Furthermore we have updated Fig. S1 for the new areas over stable ground used for vertical adjustment and error estimation.

Finally we have largely changed the corresponding section in the text:

Before differencing, the TSX/TDX-DEMs must be vertically referenced. For this purpose the median vertical offset between the DEMs and the TanDEM-X global DEM was measured over stable areas (i.e. tops of nunataks and rock outcrops, which were not affected by image distortions) at altitudes between 150 m and 1000 m (Fig. S1), before both DEMs were adjusted accordingly.

After subtracting the vertically registered DEMs, the elevation differences were converted into yearly elevation change rates. We assessed the accuracy of the vertical registration over another set of stable areas at altitudes between 150 m and 1300 m (Fig. S1). The absolute median value of the extracted change rates was 0.37 m a^{-1} which primarily accounts for errors related to the vertical registration.

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The comparison between elevation change rates obtained from the 2011–2014 OIB ATM flights and the 2011–2014 TSX/TDX data after the vertical registration of the DEMs showed a maximum overestimation of ice thinning of 1.25 m a^{-1} for the TSX/TDX measurements (Fig. S4 a, b). However, the general trend of the elevation change rates fits well to those calculated from the LiDAR data and significant differences in elevation change were only measured in the lower areas of the glacier tongue. In the upper areas (above 600 m altitude) the difference between ATM and TSX/TDX elevation change rates was close to 0 m. Here the snow volume was likely completely frozen on both dates of acquisition, so that the penetration bias cancelled out. A backscatter comparison showed lower values in 2014 than in 2011 in areas below 600 m altitude, whereas the backscatter in the upper areas above 600 m altitude was similar for both dates (Fig. S4 d).

P7, L23: “determent” is probably the wrong word here.

The word has been deleted during reformulation of the section according to the suggestions of reviewer #2.

P8, L9: It seems there was some slowdown after 1994, and gradual acceleration started in 2003. The selected graphic representation is not favourable for capturing such features.

We thank the reviewer for this helpful comment. We have added a curve showing the (smoothed) median velocities measured along the centreline profile to Fig. 2. The median absolute deviation (NMAD) of the median velocities between 1994 and 2007 is just 0.06 m d^{-1} . Hence, velocities were pretty stable during this time. A comparison of velocities acquired in October 2007 and October 1995 shows almost no difference in median surface velocities. A clear sudden strong acceleration is first visible in 2008 and a strong gradual acceleration is apparent between 2010 and 2011. However, we have changed the corresponding section into:

Figure 2 shows that glacier velocities were rather stable between 1994 and 2007. The normalised median absolute deviation (NMAD) of the median velocities during this time was 0.06 m d^{-1} . A comparison of the velocities on 1995-10-27 and 2007-10-23 (Fig. 3a) along the centreline profile reveals that the median velocity difference between 1995 and 2007 was just 0.04 m d^{-1} .

P9, L7: “Figure 5 shows that prior to the speedup (2004–2008) Fleming Glacier was already affected by pronounced surface lowering.” No elevation change data prior to 2004–2008 are shown in Fig. 5.

We apologize for the misleading phrasing. In Fig. 5 elevation change rates are shown that were calculated from data acquired in 2004 and 2008. The time period 2004–2008 does not refer to the speedup but to elevation change. We have changed the section into:

Figure 5 shows that prior to the speedup in 2008, Fleming Glacier has already been affected by pronounced surface lowering. A clear trend of increasing ice thinning rates towards the glacier front is visible for 2004–2008.

P9, L28: “a vast part of the formerly grounded glacier tongue”. “vast part” is a subjective impression; be specific.

We thank the reviewer for this helpful advice. We have changed the corresponding section into:
However, the same calculations for data acquired in 2011 and 2014 (Fig. 6, Track 3–5) as well as patterns of low and positive elevation change rates in the TSX/TDX 2011–2014 dh/dt map (Fig. 4, S7) show that after the final stage of glacier acceleration in 2011 an area of about 56 km² (referring to the front in 2014) of the formerly grounded glacier tongue of the Airy Rotz Seller Fleming system had been floating

P10, L14: “A possible location of the grounding line after the initial ungrounding in 2008. . .”. Confusing statement. Was the area floating before 2008 and then became grounded again?

We apologize for the misleading phrasing. Until 2008 the frontal part of the glacier tongue was likely pinned to a subglacial ridge (probably a sill) located at the 1996 GL position. However, an ice cavity between the pinning point and a smaller hump ~ 2.5–3 km upstream of the pinning point has probably already existed. In 2008 the glacier tongue likely detached from its pinning point at the 1996 GL position and from then on the glacier was suddenly only grounded at the hump ~2.5–3 km upstream. The grounding line was stabilized there until between 2010 and 2011 it further gradually retreated to its present position ~ 7–9 km upstream. We have changed this section accordingly during restructuring of the Discussion part. We have further added a view graph to the supplement (Fig. S 8) showing our interpretation of the ungrounding process, in order to make it better understandable to the reader.

P10, L25: The Rothera station data (near Wordie I.S.) do not show a cooling trend for recent years. Oliva et al. (Science of the Total Environment 580, 210–223, 2017) report higher mean annual air temperature for 2006 to 2015 than in the previous decades.

We thank the reviewer for pointing us to the publication of Oliva et al. (2017). In contrast to other studies (e.g. Turner et al., 2016) which found an overall cooling trend for the entire Antarctic Peninsula since the late 1990s based on stacked temperature records, this study provides a regional analysis of temperature trends at different stations across the Antarctic Peninsula and the South Shetland Islands (SSI). Indeed Oliva et al. (2017) report higher mean annual air temperatures (MAAT) for the period 2006–2015 at Rothera and San Martin (which is closer to Wordie Ice Shelf than Rothera) than in previous decades. While the decadal mean temperatures at all stations on the SSI and the N-NE Antarctic Peninsula were 0.2–0.9 °C lower in the period 2006–2015 than in the period 1996–2005, at Rothera and San Martin decadal mean temperatures were 0.1 and 0.2 °C higher, respectively (Fig. 1 and Tab. 5, Oliva et al., 2017). However, when looking at decadal temperature trends, Oliva et al. (2017) reveal that although temperatures at Rothera and San Martin do not show a continuous cooling trend as recorded at other stations since the late 1990s, warming rates at both stations have decreased markedly since the decade 1996–2005 and a cooling trend was present between 2006 and 2015 (Tab. 5 and Figs. 2 and 4, Oliva et al., 2017). We consequently reworded the corresponding section into:

Furthermore, a trend of cooling air temperatures is reported for the Antarctic Peninsula since the end of the 1990s (e.g. Turner et al., 2016). Although decadal mean surface temperatures in the period 2006–2015 were 0.2 °C higher than in 1996–2005 at San Martin station (~120 km north of Wordie Ice Shelf), warming rates have decreased markedly since the decade 1996–2005 and show a cooling trend in 2006–2015 (Oliva et al., 2017). This may have reduced surface melt during recent years.

P10, L31, L32: Here it is stated that “flow acceleration usually affects both the floating and the grounded part of the glacier, but is largest near the grounding zone”, and also “This is consistent with our observation of a highest relative speedup by ~32–35 % between 7 and 11 km upstream”. If the highest speedup is 7 to 11 km upstream, the second statement is not consistent with the first one.

The sentence has been removed during rewriting of the section.

P11, L4: Any direct observations supporting the statement that basal melt is particularly effective in the grounding zone?

Same answer as for P1, L16

P11, L6: In which respect is the bedrock topography “unfavourable”?

The trough underneath Fleming Glacier has a retrograde slope. Such a bed topography is known to be an unstable configuration which fosters rapid grounding line retreat (e.g. Rignot et al., 2014). We have added a note on this during rewriting of the Discussion section.

P11, L7, L8: Fig. 5 shows thinning rates only for grounded ice. This does not provide any clear link to basal melt rates of the floating part in the grounding Zone

We agree. However, it has been shown by several other studies that Wordie Bay is subject to intruding warm CDW which causes substantial subglacial melt. Hence, it is very likely that basal melt also occurred prior to 2008 at the pinning point (at the 1996 grounding line). The sentence has been deleted during rewriting of the Discussion section and the central statement has been changed accordingly.

P12, L12, L13: “Our data suggest that enhanced basal melt at the grounding line due to increased shoaling of warm CDW most likely played a major role for the recent changes at Fleming Glacier.” This conclusion is not based on the measurements presented in this study, but rather on measurements in other locations and reported in other publications.

We agree with the reviewer. We have changed this during restructuring of the Discussion section. Apart from that, same answer as for P1, L16

P12, L21: Please explain the expected “fatal effects on the stability and sea level contribution”.

Modelled bed topography shows another deep subglacial trough with a retrograde sloped bed geometry 3-4 km upstream of the current grounding line. If the grounding line retreats to this trough, this may trigger a positive feedback loop of rapid grounding line retreat, flow acceleration, dynamic thinning, increased calving, and mass loss (marine ice sheet instability). However, we have reformulated and weakened the statement during restructuring of the Conclusion part

P19, L5: Fig. 1 caption. Please check the dates for Sentinel-1 data. One year time span for velocity retrieval?

We apologize for the mistake. The data used for calculating the velocity were acquired on 28-08-2015 and 09-09-2015. We have changed this.

P24, L5: Fig. 6 caption. “Fulfillment of floating condition”. Does this refer to grounded or floating ice?

We apologize for the misleading phrasing. Orange dots show grounded ice along PIB flight lines as derived from hydrostatic equilibrium for the time before acceleration. We have changed the caption into:

Orange dots: Grounded ice before acceleration as derived from PIB LiDAR and ice thickness data. Dates of PIB flights: a) 2002-11-26, b) 2004-11-18.

P24, L7: “Fulfillment of the hydrostatic equilibrium condition”. Same question as for P24, L5.

We apologize for the misleading phrasing. Blue dots show freely floating ice and red dots show grounded ice along OIB flight lines as derived from hydrostatic equilibrium for the time after acceleration. We have changed the caption into:

Blue and red dots: Freely floating and grounded ice after acceleration as derived from OIB laser altimeter and ice thickness data. Dates of OIB flights: c) 2011-11-17, d) 2014-11-16, e) 2014-11-10.

L25, Table 1: SAR, column 6 shows single dates (not “Time Interval”) Supplement

We thank the reviewer for pointing us to the misleading labelling of the column. Indeed the dates in the column refer to time intervals of SAR data used in the study. We have changed the labelling of the column to make this clearer to the reader.

Table S2: Pixel size should be accurately specified in along track (or LOS) and across track direction. These can be quite different even for a single sensor (e.g.Sentinel-1).

We agree with the reviewer that pixel sizes can largely differ between the across track and the along track direction. Hence, to make a conservative estimate of velocity errors, we always chose the coarser resolution value to be the image resolution Δx used in Formula 1 (S 2). In order to make this clearer to the reader we follow the recommendation of the reviewer by adding two columns to the table which show the resolution in ground range and azimuth. We also have changed the caption of Table S2 accordingly:

Ground range resolution Δx_{GR} , azimuth resolution Δx_{AZ} and image resolution Δx used in Formula 1 (S2) for calculating velocity errors. For most of the sensors Δx_{GR} is coarser than Δx_{AZ} , except for Sentinel-1a that has a coarser resolution in azimuth direction and TSX/TDX which have fairly equal resolutions in both directions. In order to make a conservative estimate of velocity errors, always the coarser resolution value was chosen to be Δx

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Response to the Interactive comment on

“Recent dynamic changes on Fleming Glacier after the disintegration of Wordie Ice Shelf, Antarctic Peninsula”

by Peter Friedl et al.

Anonymous Referee #2

Received and published: 25 October 2017

We would like to thank the reviewer for constructive and helpful comments on our manuscript. All comments have been considered and a list of responses and changes in the manuscript is given below. Responses are written in bold face type and changes in the manuscript are written in blue.

Friedl et al. present a dense time series of surface velocities and elevation change rates of the Fleming Glacier system in the past two decades from multi-source remote sensing data. This study identified two significant acceleration phases occurred in 2008 and 2010-2011 after the Wordie Ice Shelf has nearly disappeared in front of the Fleming Glacier. The determination of floating area based on the hydrostatic height anomalies, bedrock data, elevation change rates, and acceleration phase is very helpful to explain the rapid dynamic changes after 2008. However, some points in this study are not well explained and need further rewording. Some minor wording and grammar errors also need to be fixed.

General Comments:

1) It is not clear about the epoch of the 2008 grounding line position. Please clarify it, particularly regarding locations in Fig. 6 and Fig. 6S.

We have added more information on how we decided the grounding line locations in 2008 to the methods section (Sect. 4.3) and especially to the discussions on Fig. 6S. We are aware that the exact 2008 grounding line positions are vague and we have tried to make this clearer in the text. However, since the glacier front had retreated behind the grounding line 1996 in 2008 for the first time, the 2008 grounding line must have been situated upstream of the 1996 position.

2) The method used to determine the grounding line position in 2008, 2011 and 2015 from the velocity data, surface-thinning rate, and the bedrock data is not clear enough in the method (Sect. 4.3, P7, L19-27) and result section (Sect. 5.3), especially regarding the determination of the grounding line position in 2008. From Fig. 6, it is appeared that the 2015 grounding line is mainly based on the ridges of modelled bedrock combined with the hydrostatic height anomalies along OIB flight lines. However, profiles in Fig. S5 show that how much the Bedmap2 bedrock and Huss and Farinotti's modelled bedrock can differ from the PIB and OIB measurements. Then how reliable are the hills used to decide the grounding line location? In the analysis of Fig. 6S, it is better to explain how the different evidence (velocity profiles, elevation change profiles, hydraulic height anomalies, and modelled bedrock) is being combined to determine the estimated 2008 and 2015 grounding line. If they are not consistent with each other, please explain the final selection.

We have largely rewritten Sect. 4.3 Sect. 5.3 and the discussions to Fig. S6 a-d in order to better explain how we finally decided the grounding line positions. We have also addressed possible uncertainties in the bedrock data.

3) This study (P11, L27-29) concludes that the basal melting driven by ocean-warming is the dominant trigger of the grounding line retreat and glacier acceleration. However this lacks direct evidence, namely the basal or frontal melting rates underneath the ice shelf. Given the little ice shelf left in 2008, it is not very convincing to say that the basal melting underneath the ice shelf dominates the ungrounding process.

We understand the reviewer's concerns. However, Depoorter et al. (2013) and Rignot et al. (2013) calculated high basal melt rates of 14.79 ± 5.26 m/a and 23.6 ± 10 m/a respectively for the remaining parts of the ice shelf. These melt rates are similar to those found for ice shelves in the Amundsen Sea sector (Table 1 in Suppl. Material and Fig. 2, Depoorter et al., 2013).). Hence, it is likely that basal melt also occurs at the grounding line. Furthermore the coincident timing of exceptional warm water intrusions in 2008/2009 and 2010/2011 into Marguerite Bay found by Walker and Gardner (2017) and the two acceleration phases found by us, indicates that ocean warming and the observed changes in glacier dynamics are strongly correlated. Anomalously low sea ice extent in Marguerite Bay, observed in 2008 and 2010, indicates that the warming events affected the upper water column, but Walker and Gardner (2017) found that the years 2009-2011 had the highest temperatures also at depths of > 400 m. Hence, it is very likely that submarine ice melting at the grounding line was increased during phases of strong CDW upwelling events and that this has triggered unpinning in 2008 and gradual grounding line retreat between 2010 and 2011. We rewrote the Discussion, Conclusion and Abstract sections in such a way that it points out that our conclusion on basal melt at the grounding line is not based on direct observations made in this study but on reasonable assumptions upon existing evidence.

Specific Comments:

P1, L8: change to “. . . regions most affected by climate change”
The sentence has been changed accordingly.

P1, L14: The finding about glacier ungrounded in 2008 between January and March has not been confirmed by observation or measurements. Please reedit this sentence.

The glacier front had retreated behind the 1996 grounding line location for the first time in 2008. Hence, the grounding line must have been located upstream of its 1996 position by this time. Before 2008 no marked changes in surface velocity have been observed, which would have indicated grounding line retreat. Although the precise position of the grounding line after the acceleration in 2008 remains vague, at least it is clear that in 2008 the grounding line was located upstream of its 1996 position. It is likely that confirmed intrusions of warm CDW to Marguerite Bay in 2008 have triggered unpinning of the glacier tongue from a pinning point located at the 1996 grounding line position. However, the sentence has been reformulated during rewriting of the abstract.

P1, L24: “Currently, the tongue of Fleming Glacier is grounded in a zone of bedrock elevation of -400 m” This statement is not mentioned in the main text. Please state this in the main text and make the elevation color bar in Fig. 6 clearer for reading the elevation values. It will be helpful to add elevation contours.

We thank the reviewer for this helpful suggestion. We have added elevation contours and a more detailed color bar to Fig. 6. We have also added a note on the elevation level of the recent grounding line (~-400 to -500 m) to the main text (section 5.3):

For most regions of the glacier tongue we estimate the current grounding line to coincide with these ridges at an elevation between ~-400 and -500 m.

P1, L26: You can't say “this endangers” “a huge potential”. Add a suitable verb before “a huge potential”. “huge” is not appropriate here.

We have changed the sentence into:

Hence, this endangers upstream ice masses, which can significantly increase the contribution of Fleming Glacier to sea level rise in the future.

P2, L4: Add “be” before the number “4.21”.

We have changed the sentence accordingly:

P3, L3: “former tributary glaciers” is confusing since the glaciers still exist, Suggest instead “. . . originally fed by several major tributary glaciers (Fig.1). Among these, Fleming Glacier is . . .”.
We have changed this accordingly.

P3, L10: “Here the ice shelf was temporarily grounded and stabilized . . .” is confusing. Does “Here” mean at a time point or at the location of the pinning points? Is there evidence of grounding of the previously floating Wordie Ice Shelf? Do you mean it was stabilized temporarily? Do you mean “the next rapid break up event” occurred in 1989? If yes, please specify it.

We apologize for the ambiguous phrasing. “Here” refers to the location of the pinning points (e.g. ice rises/rumples). Based on the analysis of satellite imagery e.g. Vaughan (1993) concluded that the ice front retreat of Wordie Ice Shelf was punctuated by periods of stasis during which the ice front rested on these pinning points. When embedded in the ice shelf they helped to stabilize the ice shelf by compressing it upstream and providing restraint at the grounding line. However, during the following rapid break up events (not just the 1989 event) the ice rises appeared to behave as indenting wedges, contributing to weakening the ice shelf and accelerating break up. We have reformulated the section into:

Analyses of satellite imagery suggest that the ice shelf was temporarily grounded and stabilized at these pinning points until one of the next rapid break-up events took place (Doake and Vaughan, 1991; Reynolds, 1988; Vaughan, 1993; Vaughan and Doake, 1996). However, during phases of ice front retreat, instead of protecting the ice shelf against decay, ice rises that were embedded in the ice shelf appeared to behave as indenting wedges, contributing to weakening the ice shelf and accelerating break up (Doake and Vaughan, 1991; Vaughan, 1993).

P3, L18-19: Please add the ice front position in 1997 and 2000 in Fig. 1. It will help reader to understand your statement here.

We have changed Fig. 1 accordingly.

P4, L21-25: The information about the modelled bedrock topography from Huss and Farinotti 2014 is not enough. Given its role, indication of origins and uncertainties should be included in the paper.

We agree with the reviewer. We have added more detailed information on the dataset to the corresponding section:

The dataset was generated by subtracting modelled ice thickness from the improved ASTER GDEM by Cook et al. (2012). Ice thickness was derived by constraining a simple model based on the shallow ice approximation for ice dynamics with observational data of ice thickness (OIB) and surface velocity (Rignot et al., 2011). Where available the modelled ice thickness was corrected with OIB ice thickness data, leading to more precise ice thickness values in such areas. On average, the local uncertainty in ice thickness of the dataset is $\pm 95\text{m}$ but values can reach $\pm 500\text{m}$ for deep troughs without nearby OIB measurements. Since OIB coverage is fairly good across Fleming Glacier, uncertainties in modelled ice thickness are relatively low in this area. However, a comparison of bedrock elevations from Huss and Farinotti with bottom elevations calculated from OIB ATM and CoRDS measurements shows that although the modelled bedrock reflects the general subglacial topography well, the absolute difference in bottom elevation can be even more than 100 m (Fig. S5). One possible reason for this is a difference between ATM heights and the refined ASTER GDEM, which transfers to bedrock elevation.

P7, L18: Modify “Sect. S2” to “Sect. S3”.

We have changed this.

P7, L22: “mitigates” is hardly correct for 90

We have substituted “mitigates” by “decreases”.

P7, L22-23: It is not clear how you decide the grounding line position from the surface velocities. Please clarify. If you define the grounding line based on where the maximum velocity increase occurred, please specify it with relevant references.

We have added some more sentences with references on how and why information on surface velocity can be used to identify the grounding line:

Moreover, the grounding line marks the transition between two fundamentally different flow regimes of grounded and freely floating ice. Whereas the flow dynamics of grounded ice are dominated by vertical shear and controlled by basal drag, flow of floating ice is drag free and dominated by longitudinal stretching and lateral shear (Schoof, 2007). The difference in flow dynamics of grounded and floating ice can result in pronounced changes in surface velocity close to the grounding line (Stearns, 2007; Stearns, 2011). Moreover, Rignot et al. (2002) demonstrated that if ungrounding occurs, the resulting flow acceleration usually affects both the floating and the grounded part of the glacier, but is largest near the grounding zone. Thus, velocity profiles can serve as additional information for locating the grounding line (Stearns, 2007; Stearns, 2011).

P7, L23: replace “determent” - do you mean “determinant”?

The word has been deleted during rewriting of the section.

P8, L18: Please comment on the higher velocity changes in the 2011 glacier front in Fig. 3 and why it is ignored.

We thank the reviewer for this helpful question. In 2011 the glacier front had an all-time minimum position. Consequently in 2007 and 2015 the ice front was located seaward of the 2011 front. Ice velocities are highest at the glacier front. Hence, at the location of the 2011 front inherently high front velocities were subtracted from inherently lower velocities of floating ice (2011-2015) or vice versa (2007-2011). This results in peaks of velocity change at the 2011 front position. Thus the strong velocity changes at the 2011 front do not represent “real” dynamic change and were ignored. We have added a note on this in the text. Furthermore, we have modified Fig. 3 a, b in such a way that we now do not show velocity change for 2007-2011 and 2011-2015 seaward of the 2011 front anymore. Otherwise here velocity change would result from comparing velocity of the glacier tongue with velocities of sea ice or ice mélange.

Fig. 3a, b show absolute and relative velocity changes along the centreline profile for the periods 1995–2007, 2007–2011 and 2011–2015. In 2011 the location of the glacier front reached its most inland position. Consequently velocities measured on sea ice or ice mélange in 2011 (seaward of the 2011 front) were excluded from our analyses. Large changes in surface velocity close to the 2011 front in the periods 2007–2011 and 2011–2015 do not represent real dynamic change, but result from comparing the inherently higher frontal velocities in 2011 with lower velocities of the floating glacier tongue in 2007 and 2015.

P8, L23-24: Suggest updating Table2 by adding the velocity in 2015 from Walker and Gardner, 2017 at three sites.

We have updated the table and have added a note on the higher velocities in 2015 at the three sites to the text:

However, surface velocities derived from Landsat 8 feature tracking suggest that in 2015 velocities at the three measuring sites had increased by ~20% in comparison to 2008 (Walker and Gardner, 2017; Zhao et al., 2017) (Tab. 2).

P8, L27: As I could tell, the highest ice thinning rates for Fleming Glacier from Fig. 4 (orange color) is higher than 6 m a^{-1} .

The reviewer is right. We apologize for the imprecise formulation. We have changed the sentence into:

On Fleming Glacier the highest ice thinning rates with peak values of more than $\sim 6 \text{ m a}^{-1}$ were recorded in a zone extending from ~ 8 to $\sim 14 \text{ km}$ upstream.

P9, L2: “For the location of the data see Fig. 4” ! “The location of the data is shown in Fig. 4”.

We have changed this.

P9, L13: Missing space after the full stop.

We have added a space.

P9, L19-20: It would be useful to compare your thinning rate with that from Zhao et al., 2017 and Walker and Gardner, 2017.

We have added a comparison of our results and the thinning rates from Walker and Gardner, 2017 to the Discussion section: However, we do not compare ice thinning rates with those of Zhao et al., 2017, since the time intervals they used for computing elevation change differ from ours.

P10, L1-3: The “We extracted” part of the statement should explicitly refer to Fig. 6S, and the location of the four profiles should refer to Fig. 6.

We have changed the section into:

We extracted data of surface velocities, TDX/TSX 2011–2014 elevation change rates, bedrock topography along four profiles on Airy and Fleming Glacier in order to estimate the recent and the grounding line location in 2008 after the first acceleration phase (Fig. S6 a-d). The location of the profiles as well as the deduced grounding line positions are shown in Fig. 6.

P10, L3-4: For “The plots”, you mean Fig. 6 or Fig. S6? It would be clearer to say “Those profile plots (Fig S6 1-4) suggest . . .”. Which glacier or profile are you talking about for the 2008 grounding line position?

We mean Fig. S6 1-4. We have changed this in the text accordingly.

P10, L5-6: “possible” and “likely” are same to me. It’s not clear how the grounding line positions in 2008 are decided on Profile 2, 3, and 4.

The sentence has been removed during rewriting of the results section (5.3). We have changed the term “possible” to “tentative” making clear that the 2008 grounding line position is only a rough estimate. We have added more information on how the 2008 grounding line positions were deduced from the profiles to Sect. 4.3, 5.3 and Fig. S6 a-d.

P10, L7-15: Is detailed discussion of a supplementary figure in the main text proper? Consider including Fig. S6-3.

We agree with the reviewer. We have removed the section during rewriting of the results section (5.3). We hence have kept the Figure S6 c in the supplemental material.

P10, L18-20: About “no significant acceleration since 1996 . . .”, if you compare the velocity at three sites with the 2015 velocity from Walker and Gardner, 2017, you will find some speed up from 1996 to 2015. Alternatively you need to qualify with “by 2013”.

We have changed the section into:

The remarkable median speedup of $\sim 1.3 \text{ m d}^{-1}$ which we recorded between 2007 and 2011 is in good agreement with an acceleration of $\sim 400\text{--}500 \text{ m a}^{-1}$ reported by Walker and Gardner (2017) and Zhao et al. (2017) for the period 2008–2015. However, a comparison of our velocities in 2013 with their velocities in 2015 at the three measuring sites of Doake (1975) $\sim 50 \text{ km}$ upstream suggests that the recent speedup had not propagated up to these locations prior to 2015.

P10, L27-28: Besides the surface melt, basal melting water generated from the basal friction heating could be another trigger of enhanced basal sliding. It’s hard to rule out the possibility of enhanced basal sliding unless you have further evidence.

We agree that liquid water generated from frictional heat can enhance basal sliding. Basal friction heat can increase if the glacier accelerates or if it gains mass. Hence, increased frictional heat is more likely a result of the acceleration, rather than the trigger of it. However, we do not want to rule out that in areas where the glacier is currently grounded, as a consequence of initial acceleration, basal sliding is enhanced due to melt water from increased frictional heat.

We have changed the section into:

All in all, enhanced basal sliding due to percolating meltwater is likely not the explanation for the observed increase in flow velocities. However, we do not rule out that as a consequence of the acceleration, basal sliding had increased in grounded areas by liquid water generated from greater basal frictional heat.

P10, L30-31: If this is your method to decide the grounding line position from the velocity data, please move this sentence to Sect. 4.3.

We have moved and reformulated the sentence (see answer P7, L22-23)

P11, L12: Please cite references for the Twaites Glacier and the Pine Island Bay.

We have added a reference to Rignot et al. (2014).

P24, L4: Modify “Linens” to “Lines”.

We have changed this.

Supplement Material

It is better to use consistent zero distance mark for the various profiles in both the main text (Fig. 2, 3, 5, 6) and the supplementary material (Fig. S6). The use of different starting points (the 1996 grounding line position or the 2007 ice front position) for describing the distance is confusing. If it is too difficult to shift the origin in the supplementary figures, the GL96 grounding line needs to be clearly marked.

We thank the reviewer for the helpful suggestion. We have changed the profile extent to the 1996 grounding line in Fig. S6.

P12: Modify the legend of a) and b) from “Ice Surface/Bottom elevation (OIB)” to “Ice Surface/Bottom elevation (PIB)”.

We have changed the legend accordingly.

P15, L3: Modify “profiles 1-5” to “profiles 1-4”. Also consider labeling figure parts as S6.a-d for consistency with other sections. Anyway add labels to the profile figures.

We have changed the labelling to S6.a-d and have modified the references in the text accordingly.

P15, L4: Can’t find OIB ice surface/bottom in those figures.

We have deleted the corresponding sentence.

P15, L5: Modify “grounding line” to “GL”.

We have changed this accordingly.

P15, L6-7: Can’t find Green, Red and Blue lines in those figures.

The sentence has been deleted.

P16, L7-8: It’s hard to tell the 2008 grounding line position upstream the 2011 front without direct evidence. Please explain it.

We agree with the reviewer that it is difficult to tell the exact 2008 grounding line position from the modelled bedrock data only. The only thing which is clear is that the grounding line must have been upstream of its 1996 position in 2008, since in 2008 the glacier front had retreated behind the 1996 grounding line location. The position which we state is only a best guess based on the data we have in hand. We hope that we have made this clearer in the rewritten text to Fig. S6 b and in the results section. We have also pointed to the limitations of the bedrock data in the text and have labelled the 2008 position in Fig. S6 b with a question mark.

P16, L10-11: Please add GL11 in Fig. S6.2.

We have added the GL11 to Fig. S6 b.

P17, L3: Move “in 2008” after “the grounding line”.

We have changed this accordingly.

P17, L5-6: Along Profile2, it’s hard to tell this hill in Fig. 6. Considering the uncertainty of bedrock (yellow and brown line in Fig. S6.3), is it possible that this smaller hill is not physically real? On P18, L5-6 you suggest that “the topography may be distorted in the modelled bedrock data”.

Same answer as for P16, L7-8

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Response to the Interactive comment on

“Recent dynamic changes on Fleming Glacier after the disintegration of Wordie Ice Shelf, Antarctic Peninsula”

by Peter Friedl et al.

Anonymous Referee #3

Received and published: 16 November 2017

We would like to thank the reviewer his helpful comments on our manuscript.

All comments have been considered and a list of responses and changes in the manuscript is given below. Responses are written in bold face type and changes in the manuscript are written in *blue*.

The topic of this paper, Fleming Glacier, is extremely interesting given the significant changes reported by previous publications, so I was very keen to read these new results. The authors present new ice velocity, elevation change and grounding line position data acquired from a range of airborne and satellite based instruments. The ice velocity results are nice, and I believe constitute the most complete time series of velocity measurements over Fleming glacier which provides new insight into the timing of ice speedup on this sector.

However, after reading the manuscript there are a number of major flaws with the methods employed to derive surface elevation change and to measure the grounding line retreat. As it stands, the problem with the techniques make it highly likely that both the magnitude of the elevation change signal, and the grounding line retreat, may not be correctly reported in this paper. For example, cross calibrating the DEM elevations over sea ice, which varies annually and can range in thickness from 0-5m, is extremely unsatisfactory. Moreover, even if this correction is accepted, the authors estimate that known X-band penetration bias can account for ~50% of the dh/dt signal across the basin. Even if the estimate of elevation change is accepted, the associated error measurements do not reflect spatial variability in the data quality, and are unrealistically small given the spread of the raw data. For example, it is stated that the CAMS-ATM dh/dt data has an error of 0.2m/yr in regions where the point measurements at the same location range from -0.5 to -7 m/yr. These errors must therefore be revised. Regarding the estimate of grounding line retreat, the technique is unproven and is un-validated in this paper. Even if the authors demonstrate that the technique can be trusted, as it stands the results presented in this paper are contradictory because regions measured to be grounded are located within the new floating ice shelf area.

On top of these technical issues, I have a few more minor concerns about the use of scientific terminology throughout the paper, and the overly simplistic nature of the analysis and discussion sections. The specifics of these and other concerns are documented in detail below. My criticisms of the methods and results presented in the paper are major, and will be time consuming and require a significant effort to properly address. The implications of these concerns is that I believe the magnitude and spatial pattern of the elevation change and grounding line retreat data may be incorrectly reported in this paper. This is significant as I do not have confidence in two of the three core datasets presented, therefore, it is my recommendation that this paper is not suitable for publication in its current form.

Specific Edits

P1 L21 – Edit ‘far upstream of the glacier’, using this as a location is ambiguous, better use a fixed reference location, like the calving front in ‘x’ year.

The sentence has been deleted during rewriting of the abstract.

P1 L26 – Edit paper to quantify ‘much larger ice masses’.

It is difficult to exactly predict how much ice will get lost in the future due to the response of the glacier to further grounding line retreat from observational data. However, if the grounding line retreats up to the deep retrograde trough ~3-4 km upstream of its current position, it is likely that the glacier gets into an unstable configuration where rapid grounding line retreat and more mass loss than today can be expected (see answer P11 L11). The quantification of the expected mass loss, however, needs the involvement of ice sheet modelling, which is out of the scope of this paper. Nevertheless, we have changed the sentence into:

Hence, this endangers upstream ice masses, which can significantly increase the contribution of Fleming Glacier to sea level rise in the future.

P2 L4 – Edit missing to ‘be’.

We have changed this accordingly.

P2 L7 – Edit ice ‘shelf’ tributaries.

We have changed this accordingly.

P2 L30 – Edit ‘explain’ to ‘investigate’. I’d argue at this point the authors haven’t demonstrated they can explain the observed signal.

We have changed this accordingly.

P3 L23 – Poor wording, edit out ‘got’ both times.

We have changed the sentence into:

During the following years (2010–2015) the fronts of the glaciers in Wordie Bay remained quite stable, except at the Prospect system where the once interconnected floating ice tongues of the three glaciers disconnected and some floating ice was lost.

P4 L2 – Edit ‘dynamic’ and check use throughout paper. Previous publications (e.g. Rignot et al 2005) have demonstrated that Fleming exhibited dynamic imbalance during their study period, but as the authors state this paper spans a longer time period, and at this point it hasn’t been proven that Fleming is dynamically imbalanced for the full duration of their study period.

We agree with the reviewer, that at this point it is not clear, whether the dynamic changes we observe are consequences of the disintegration of the ice shelf or a result of other processes. However, at least we investigate the dynamic changes of the glacier in the time period after the disintegration of the ice shelf. We hence have changed the sentence into:

We used a broad remote sensing data set in order to investigate the changes in ice dynamics at Fleming Glacier between 1994 and 2016 after the disintegration of Wordie Ice Shelf.

We have also checked the use of the word “dynamic” throughout the paper and have changed it where necessary.

P5 L6 – More informative to state what % of each velocity image is removed during the filtering process as this should be an indicator of the quality of the tracking output. The filtering should remove 100% of the unreasonable results, otherwise its not a very good filter!

We apologize for this inaccuracy. The filter removes more than 99 % of erroneous vectors and removes very few false negatives (Burgess et al., 2012). We have changed this in the text accordingly. The filter has been successfully applied in several other peer reviewed studies (e.g. Rankl et al., 2014; Rankl et al., 2016; Seehaus et al., 2015; Seehaus et al., 2016). However, the proportion of velocity vectors removed from each velocity image by the filter is not only depended on the quality of the tracking but also on the coverage of the scene. For example, more erroneous tracking results are likely to be found in areas where no marked features can be used for tracking (e.g. in the interior of the peninsula). If a scene covers more of such areas, the amount of removed velocity vectors will be higher than in other scenes which only cover the glacier tongue. Hence we have calculated the proportion of removed velocity vectors for an area over Fleming Glacier, which is covered by all scenes and which is relevant for our velocity measurements. The area for which the calculations have been done has been added to Fig. S1.

Additionally Tab. S3 has been updated with a column containing the proportion of removed measurements.

P5 L16 – 0.2 m is the accuracy of the original point measurements; the authors are using the dataset after re-gridding it so state the accuracy of this dataset instead or as well as the accuracy of the raw data. The accuracy is also different for different sensors, so the authors should provide statistics for each dataset.

We thank the reviewer for this important comment. We have now cited additional references to state the accuracies of the different sensors used in this study:

We derived ice thinning rates on Fleming Glacier for 2004–2008 and 2011–2014 by comparing ellipsoid heights of the PIB (ATM, 2004), the Centro de Estudios Científicos Airborne Mapping System (CAMS, 2008) and the OIB (ATM, 2011, 2014) airborne LiDAR datasets. The vertical accuracy of the ATM elevation data is estimated to be better than 0.1 m (Krabill et al., 2002; Martin et al., 2012). For the CAMS data, vertical accuracy is 0.2 m (Wendt et al. 2010).

P6 L2/3 – This is correction extremely unsatisfactory. Sea ice is a complex parameter, and is certainly not a stable/constant reference surface for precise cross calibration of elevation measurements. In the Antarctic, sea ice can range from 0 to 5 m thickness with very large spatial and temporal variability, snow depth on sea ice is not routinely measured but can account for half the thickness retrieval, and ocean height varies with tides, atmospheric pressure etc. When deriving the correction, the authors have not attempted to account for interannual variability in sea ice thickness so this must be addressed before any confidence can be had in the elevation change measurements. This is critical because the range of thickness variability is the same order of magnitude as the dh/dt signal calculated from the DEM differencing. The authors must revise the manuscript to characterise the temporal variability of the sea ice over which they are cross calibrating the DEM's, and to rule out any influence from this factor on the end elevation change. If this effect can be proven to be negligible, the authors should also state the size of the correction, and which DEM was adjusted, in the manuscript. Having said all this, I suggest the authors don't cross calibrate the DEM's over sea ice at all as it hugely reduces the confidence that I believe we can have in these results.

We thank the reviewer for raising this important issue. According to the previous suggestions of reviewer #1 we have changed the vertical registration procedure in such a way that we now use a subset of the TanDEM-X global DEM at 12 m resolution as an absolute height reference. Before differencing, both TSX/TDX digital elevation models were vertically adjusted to the TanDEM-X global DEM according to their median offsets measured over stable ground. This improved the vertical registration and resulted in a more realistic penetration depth bias.

The maximum difference between ATM and uncorrected TDX/TSX elevation change rates was 1.25 m/a, corresponding to an absolute 3.75 m difference for the 3 year time span. However, differences in elevation change were only measured in the lower areas of the glacier tongue, where surface melt can occur. In the upper areas (above 600 m altitude) the difference between ATM and TSX/TDX elevation change rates was close to 0 m. Here the medium was likely completely frozen on both dates of TDX acquisition, so that the penetration bias cancelled out. A comparison of backscatter values showed that in areas below 600 m altitude the backscatter of the 2014 acquisition was lower than the backscatter of the 2011 acquisition. In the upper areas above 600 m, however, the backscatter values were similar.

We have updated the TSX/TDX elevation change rates in the text and all figures that contain TSX/TDX elevation change data (i.e. Fig. 4, Fig. 5 b, Fig. S4 a-c and Fig. S6.1-4).

We have also added a Figure S4 d) showing the differences in backscatter between both acquisition dates.

Furthermore we have updated Fig. S1 with the new areas over stable ground used for vertical adjustment and error estimation.

Finally we have largely changed the corresponding section in the text:

Before differencing, the TSX/TDX-DEMs must be vertically referenced. For this purpose the median vertical offset between the DEMs and the TanDEM-X global DEM was measured over stable areas (i.e. tops of nunataks and rock outcrops, which were not affected by image distortions) at altitudes between 150 m and 1000 m (Fig. S1), before both DEMs were adjusted accordingly.

After subtracting the vertically registered DEMs, the elevation differences were converted into yearly elevation change rates. We assessed the accuracy of the vertical registration over another set of stable areas at altitudes between 150 m and 1300 m (Fig. S1). The absolute median value of the extracted change rates was 0.37 m a^{-1} which primarily accounts for errors related to the vertical registration.

...

The comparison between elevation change rates obtained from the 2011–2014 OIB ATM flights and the 2011–2014 TSX/TDX data after the vertical registration of the DEMs showed a maximum overestimation of ice thinning of 1.25 m a^{-1} for the TSX/TDX measurements (Fig. S4 a, b). However, the general trend of the elevation change rates fits well to those calculated from the LiDAR data and significant differences in elevation change were only measured in the lower areas of the glacier tongue. In the upper areas (above 600 m altitude) the difference between ATM and TSX/TDX elevation change rates was close to 0 m. Here the snow volume was likely completely frozen on both dates of acquisition, so that the penetration bias cancelled out. A backscatter comparison showed lower values in 2014 than in 2011 in areas below 600 m altitude, whereas the backscatter in the upper areas above 600 m altitude was similar for both dates (Fig. S4 d).

P6 L20 – The authors calculation that up to 2 m/a of the thinning rate can be attributed to TSX DEM penetration bias. This is a huge error which accounts for ~50% of the dhdt signal present across the majority of the basin. The ICESat and ATM tracks that this error was calculated from have extremely limited coverage, and don't pass through the region with the highest thinning rates, therefore its possible that this number might even be an underestimate. For example, other studies have shown that the penetration bias in DEM's derived from TSX/TDM data over snow covered terrain can be as large as 4m (e.g. Dehecq et al 2016), which is the same magnitude as the dh/dt signal presented in figure 4. The large size of the known errors relative to the size of the signal, combined with the limited data that has been used to characterise the error makes it very difficult to have confidence in the thinning rates presented here. Other auxiliary datasets such as atmospheric temperature data, or SAR backscatter images might also be used to characterise the onset and spatial pattern of melt in the study area. The authors description of how they have accounted for this source of error is cursory given its size relative to the dh/dt signal in the study area. As surface melt and therefore penetration is known to have large spatial variability, I recommend that the authors revise their approach to account for this spatial variation across the basin, as a polynomial fit derived from a single track of airborne data will definitely not capture the magnitude or pattern of this effect across the study area.

Same answer as for P6 L2/3. Furthermore we understand the reviewer's concerns on our approach of applying a penetration depth correction derived from a single track of altimeter data to the entire study area. However, we have to deal with the data we have in hand. The tongue of the Airy-Rotz-Seller-Fleming system (which is our main study area) has only a width of ~20 km. We therefore think that it is acceptable to assume that melt conditions and the penetration depth bias is similar for this area and just altitude depending.

P6 L27 – Assuming an error of 0.2m/a just because its one order of magnitude higher than the direct inter-comparison with the ATM data the dh/dt was calibrated against isn't satisfactory. Errors are spatially variable, so the authors should revise their approach.

Due to the relatively small size of the study area the spatial variability of penetration bias in our study area should be little too (see previous answer). Hence, we think that assuming an error of 0.2 m/a due to penetration bias variability is already a conservative estimate.

P6 L33 – Why use a 35 m buffer if the ICESat footprint is known to be 70 m? I recommend the authors use the same footprint size as the aim is to do the most direct comparison possible.

We apologize for the misleading phrasing. We used a buffer with a radius of 35 m, which is equal to the 70 m diameter of the ICESat footprint. We have changed the sentence into:

To take into account the 70 m footprint of the GLAS instrument, we applied a buffer with a radius of 35 m and calculated the median from the extracted values at each point.

P7 L15 – Cite a reference for the source of the firn density correction variable.

We have added a reference to Griggs and Bamber (2011).

P7 L25 – As far as I'm aware, this method of detecting grounding line position has not been proven in peer reviewed literature. Although the logic behind it is reasonable, (i.e. if the ice is in hydrostatic equilibrium it must be floating), factors such as the spatial resolution and error on each input dataset will severely limit the sensitivity of the technique for detecting grounding line position, let alone change in grounding line position. To be convinced that the technique works, I recommend grounding line retreat from this method is evaluated against known retreat rates, in the Amundsen sea for example. If suitable data isn't available to validate this technique in another area, then alternatively a proven technique can be employed to evaluate the hydrostatic technique in this study area. For example, ERS-2 SAR data with a 3-day temporal baseline was acquired in this area in 2011, so if coherent, this should be used to produce a grounding line estimate from the proven quadruple difference interferometry technique (Rignot et al, 1998). At a minimum the authors must state the error on their estimate of grounding line position from hydrostatic height anomaly, and it follows that if the uncertainty on the measurement is greater than the change in position assumed, then the method is not viable.

We understand the reviewer's concerns. Of course DInSAR would have been the preferred method for grounding line detection. We tried to form interferograms from ERS-1 (1994) and ERS-2 (2011) data with a temporal baseline of 3 days as well as from Sentinel-1 data (2016) with a temporal baseline of 6 days. However, due to the high speed of the glacier none of the datasets maintained enough coherence over the relevant areas. Besides strong temporal decorrelation, the ERS-2 data of 2011 suffers from large and unstable Doppler centroid frequencies due to the failure of the gyroscope in 2001, which additionally hampers coherence (Miranda et al., 2003). Hence, we moved to the alternative approach of deriving the recent grounding line positions by calculating hydrostatic equilibrium from ice elevation and ice thickness data. This is a frequently used method in peer reviewed literature (e.g. Enderlin and Howat, 2013; Fricker, 2002; Münchow et al., 2014; Tinto and Bell, 2011). Typically the point of first stable hydrostatic equilibrium is located a few hundred meters seaward of the interferometric grounding line (Rignot et al., 2011). An exception are a few ice streams which have an ice plain (i.e. a region upstream of the grounding line with low surface slopes that is only lightly grounded and where the ice is very close to hydrostatic equilibrium). Here, hydrostatic analysis may place the grounding line some kilometers upstream of the real grounding line (Corr, H. F. J. et al., 2001; Crabtree and Doake, 1982). However, in our datasets we do not see any signs for the presence of an ice plain on Fleming Glacier.

In order to demonstrate the applicability of the hydrostatic height anomaly method, we have conducted the same calculations as on Fleming Glacier for the glaciers in the Cabinet Inlet region (Larsen C Ice Shelf) (Fig. 1, 2). Here coherence is also maintained in Sentinel-1 6-day temporal baseline interferograms and no long term grounding line migration has been observed since the last ERS measurement in 1996. This enabled us to compare our hydrostatic calculations with interferometric grounding lines from different epochs. Fig. 1 shows that the hydrostatic calculations are generally in good agreement with the DInSAR measurements. The differences between the 1996 and the 2016 DinSAR grounding line are mainly due to short term tidal grounding line migration. The mean difference between the first hydrostatic equilibrium and the closest DInSAR-grounding line is ~ 300 m.

We have also added a sentence on the estimated uncertainty of grounding line positions derived from hydrostatic height anomalies to S3:

For all possible sensor-depending errors of $H_i=750$ m the Monte Carlo simulation yielded a standard deviation of < 12 m for Δe . Thus we assumed the total uncertainty of Δe to be ~ 12 m. Consequently, we assigned locations on our OIB and PIB profiles to be freely floating ice, if the calculated values of Δe lay within this range. In the vicinity of the grounding zone the TDX global DEM has a minimum gradient of 1° , and therefore we estimate the uncertainty in the horizontal position of the transition from grounded to freely floating ice to be ~ 700 m.

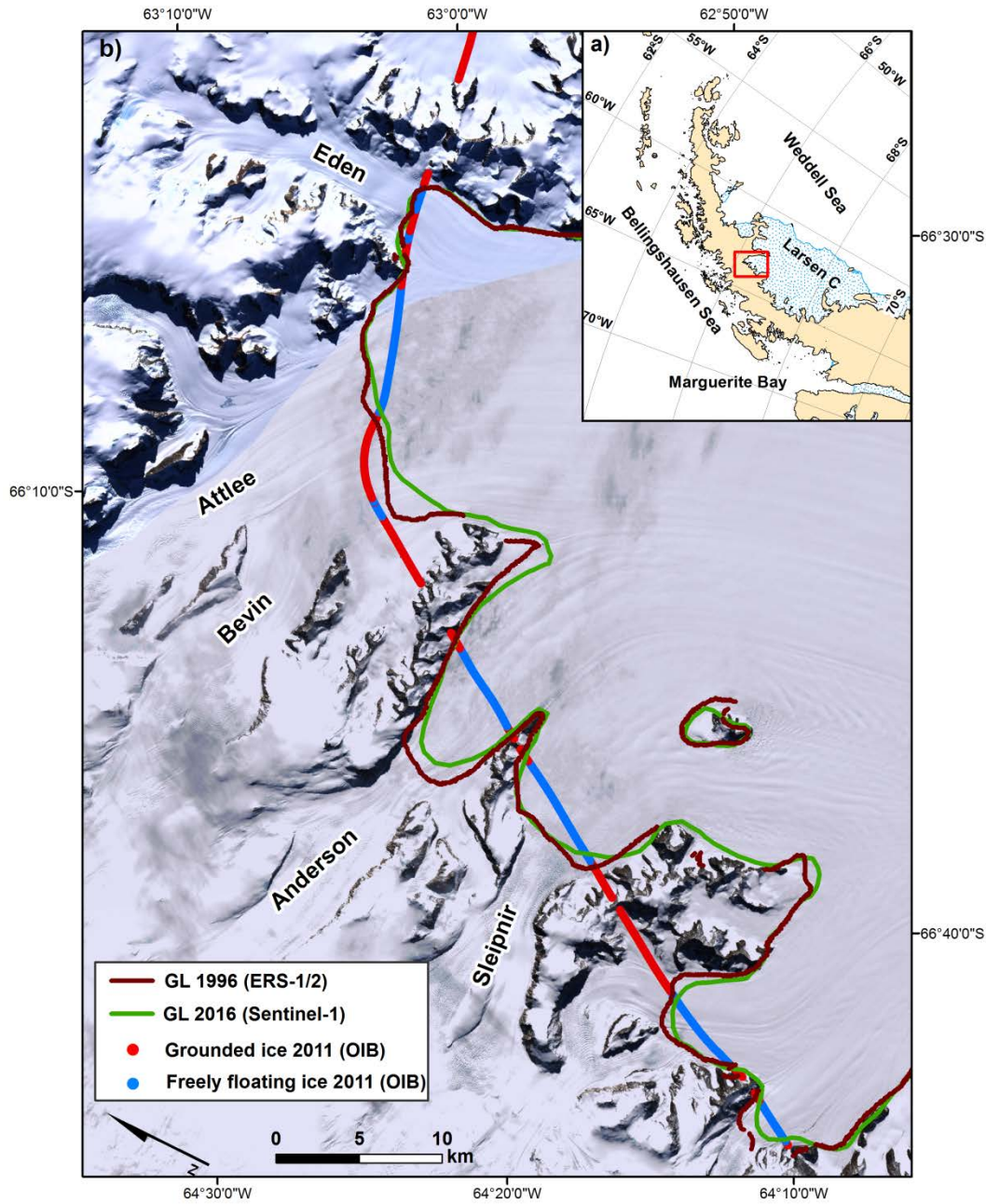


Figure 1: (a) Location of the study area at the Antarctic Peninsula. Map base: SCAR Antarctic Digital Database, version 6.0. (b) Comparison of floating and grounded ice from hydrostatic height anomalies with grounding lines derived from DInSAR. Brown line: grounding line in 1996 from ERS-1/2 (Rignot et al., 2011). Green line: grounding line in 2016 from Sentinel-1 a/b acquisitions on 2016-09-27, 2016-10-03 and 2016-10-09. Blue and red dots: Freely floating and grounded ice after acceleration as derived from OIB laser altimeter and ice thickness data on 2011-11-14. Background: USGS Landsat Image Mosaic of Antarctica (LIMA).

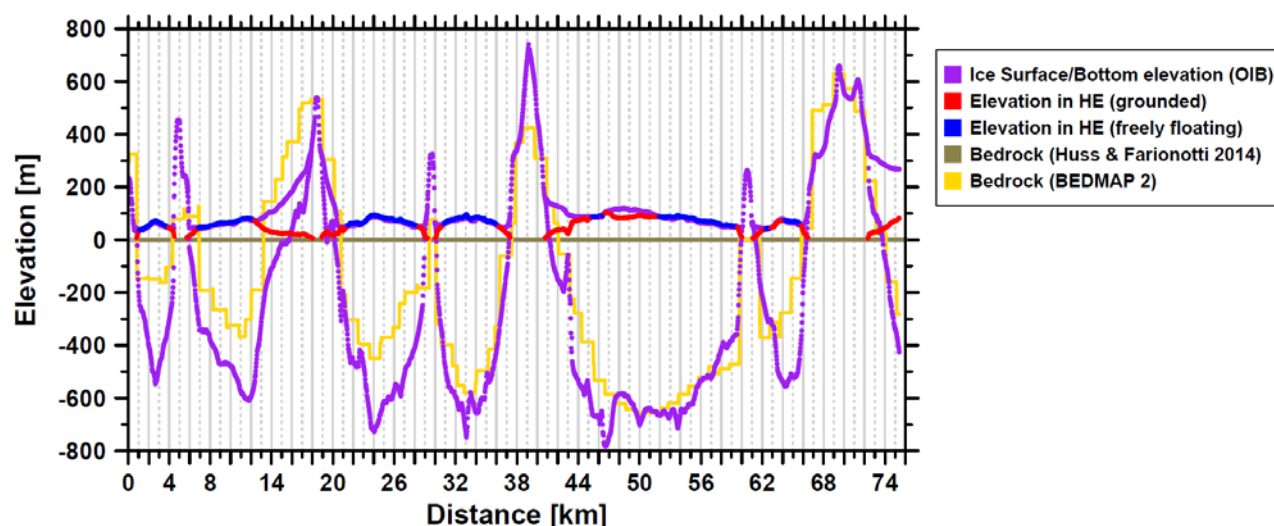


Figure 2: Fulfillment of the hydrostatic equilibrium assumption from hydrostatic height anomaly calculations along the OIB profile from 2011-11-14. The profile extends from southwest to northeast on the map (Fig. 1). Purple dots: PIB/OIB ice surface/bottom elevations. Ice surface elevation is taken from PIB/OIB ATM measurements and ice bottom elevation is calculated by subtracting OIB/PIB ice thickness from ice surface elevation. Yellow line: Bedrock elevation from Bedmap 2 (Fretwell et al., 2013). Red and blue dots: calculated ice surface elevation in hydrostatic equilibrium e_{he} and information on hydrostatic equilibrium (blue: freely floating ice, red: grounded ice)

P8 L3 – The Figure 2 z-scope is not easy to interpret. I suggest the authors re-plot this information as a standard x/y line plot of the time series of flow line ice speeds, with an inset showing change in calving front position.

We understand the reviewer's concerns. However, in our view a standard x/y line plot which contains all of our 175 measurements would be very confusing. Furthermore the huge amount of lines would make a distinguishable color coding impossible. Thus we think that our plot is a clear way to show a chronologic time series of all our measurements. The plot also reveals the timing and the propagation of pronounced acceleration very well. However, for additional information we have added a graph of smoothed median center line velocities to the plot.

P8 L14 – Poor sentence wording. Edit.

We have changed the sentence into:

Since 2008 the glacier tongue never has exceeded the 1996 grounding line anymore

P8 L28 – 8 to 14 km upstream of which location. Edit sentence to be more precise.

All distances in the text are relative to the 1996 grounding line. We have stated this at the beginning of section 5: "Distances in the subsequent text are given in reference to the grounding line of 1996." We hope that the reviewer is fine if we do not change the text here.

P8 L30 – Edit text to quantify 'lower parts'.

We have changed the section into:

A tendency to lower negative or even positive elevation change rates could be observed on the lower parts of the joint Fleming and Seller glacier tongue between 0 and up to ~9 km upstream. However the pattern was not as clear as on Airy Glacier, where a distinct area of low ice thinning rates could be detected between 0 and ~4 km upstream.

P8 L31 – Edit manuscript to remove all 'could be detected' wording. You are stating what results you have observed, so it 'has been detected', not the less affirmative 'could be'.

We thank the reviewer for pointing this out. We have changed it throughout the text.

P9 L4 – The scatter on figure 5a is very large and must be addressed given that it is significantly greater than the previously stated errors. I recommend that a) distance markers are annotated onto the ATM and ICESat track locations on Figure 4 so we can see how this corresponds to

the x axis distance scale in Figure 5. My interpretation of figure 5a is that the elevation change measurements are unusable between 0 and 20 km of the grounding line, which looks like its about up to the 'g' on the Fleming annotation on figure 4. This is the key area of interest, so vastly limits the usefulness of these datasets. B) state the method used to calculate the lines of best fit, e.g. moving average, polynomial fit? How has the clearly erroneous data been removed? C)

The large scatter of the ATM and the CAMS data between 0 and 20 km of the grounding line is due to the undulated and heavily crevassed surface, where a purely horizontal displacement of crevasses can cause apparent positive and negative elevation differences. Many crevasses are up to 50 m wide and 300 m long. Due to the high spatial resolution and precision of the airborne laser altimeter data, these characteristics of the surface are well reflected in the signal. If looking at the ICESat elevation change of 2004–2008, which by the way shows a very similar trend like the 2004–2008 ATM-CAMS data, indeed the scatter is smaller, but this is just because of the much lower spatial resolution of the data. Hence, more scatter does not necessarily mean that the data is more “erroneous”. In general, the ATM and CAMS data have been shown to be very useful to calculate elevation change over Fleming Glacier (Gardner et al., 2017; Wendt et al., 2010; Zhao et al., 2017).

To better account for the large scatter, we now follow the previously published approach by Wendt et al. (2010) by applying a median filter to the data, before calculating the cubic regression functions. However, this did not change the results substantially. We have added a note on this to the text:

The large scattering of the data is due to the highly crevassed surface of the glacier tongue, where a purely horizontal displacement of crevasses can cause apparent positive and negative elevation differences. Therefore, a median filter was applied to the data before adjustment of a cubic function.

We have also updated Figure 5a and the numbers accordingly. Furthermore, we have followed the suggestion of the reviewer by adding distance markers to Figure 4.

P9 L10 – Based on figure 5a, stating that the CAMS-ATM show elevation change of 4.1 ± 0.2 m/a is not credible. The raw data shown in figure 5a shows that at this location the elevation change ranges between -0.5 to 7 m/yr, so the error of 0.2m/yr is effectively meaningless. Please revise the error estimate here, and throughout the rest of the results paragraph.

In order to minimize the influence of scatter, we now apply a median filter before calculating the cubic regression functions of elevation change. Furthermore, for all median ice thinning rates we now calculate the normalised median absolute deviation (NMAD) as a measure of the statistical dispersion (“error”) of the input data. We have added a note on this to the results paragraph (5.2):

For all median elevation change rates presented below, we calculated the NMAD in order to account for the statistical dispersion of the input data.

P9 L12 – The fact that Fleming is thinning between 04-08, doesn't prove that the catchment hasn't reached an equilibrium since shelf collapse in 1989. The two effects may be entirely uncorrelated, so although its possible, without a continuous dataset I don't think it can be proven one way or the other. I recommend the authors revise this wording. Changing 'shows' to 'might suggest' would be more factually correct.

We agree. We have changed the sentence according to the reviewer's suggestion.

P9 L22 - Although I don't like the method, it's clear how the hydrostatic equilibrium has been calculated along the airborne tracks. Can the authors clarify what method they have used to draw the grounding line connecting the dots in Figure 6? For example, according to their own data, a section of grounded ice on track 'c' is included in the now 'floating' area. I recommend the authors revise the line as their data shows it isn't correct.

The reviewer is right. Originally we assumed that since the data used for calculating the hydrostatic equilibrium on track c was acquired in 2011, the area now may be floating. However, since there is no further evidence for this assumption, we have revised the line according to the reviewer's suggestion and have updated Fig. 6 and Fig. S7 accordingly.

Following the suggestions of reviewer #2, we have also added some more information to the methods section on how the dots have been connected:

Wherever possible, we gave preference to information on hydrostatic equilibrium for the final decision of the recent grounding line location. For selected profiles across the glacier, recent and previous (2008) grounding line positions were estimated by combining evidence from elevation change, bedrock topography and surface velocity. In the remaining areas, the recent grounding line was deduced from combining information on elevation change rate patterns in the TDX/TSX 2011–2014 dh/dt map with information on bedrock topography.

P9 L30 – Based on the above comment, the number stated for the area of the floating shelf will also need to be recalculated.

We have changed the number for the area of floating ice throughout the text to ~56 km².

P9 L33 – Quantify ‘several km’.

We have changed the sentence into:

The ridges reach up to ~9 km upstream of the 1996 grounding line.

P10 L25 – Although Turner et al 2016 shows that the long term air temperatures are decreasing the situation may be more complicated than that, and Sundal et al (2011) showed that a simple linear relationship between melt water vs lubrication is not correct, as melt induced speed-up can be offset by drainage efficiencies. I’d revise this text to avoid oversimplifying these relationships.

We agree with the reviewer that in Greenland, where strong surface melt in summer leads to enhanced basal sliding due to rapid migration of meltwater to the ice-bedrock interface, the acceleration effect may be offset by efficient drainage. However, the meltwater production on Fleming Glacier is generally not large enough to percolate to the bed (Rignot et al., 2005). No seasonal variation in surface velocities, which would indicate enhanced glacial sliding during summer, is observed. Furthermore, surface melt is likely further reduced due to decreasing air temperatures. Hence, since there is no enhanced basal sliding due to percolating surface meltwater on Fleming Glacier, the mechanisms as described by Sundal et al. (2011) play no role. Nevertheless, we have added a sentence on basal sliding and variations in drainage efficiency:

In Greenland seasonal velocity fluctuations have been linked to both enhanced basal sliding due to the penetration of surface melt water to the ice-bedrock interface and inter-annual differences in drainage efficiency (Moon et al., 2014; Sundal et al., 2011; Zwally et al., 2002).

P11 L4 – Remove sentence about basal melt. This hasn’t been measured in this study so is just a generic assumption, and no reference provided to previous study evidencing statement.

We have largely rewritten the discussion, conclusion and abstract sections. In the revised manuscript we link our observations to high basal melt rates measured by Depoorter et al. (2013) and Rignot et al. (2013) and to exceptional phases of warm CDW intrusions in 2008/2009 and 2010/2011 found by Walker and Gardner (2017). Anyway, the sentence has been deleted during reformulation of this section.

P11 L8 – Same statement as above re basal melt inference.

Same answer as for P11 L4

P11 L11 – Really poor sentence wording. Edit to be more diligent with regards to terminology. Stability is a specific process, i.e. ‘unstoppable’ retreat that will continue to propagate even if environmental conditions returned to their original state. Glacier imbalance and grounding line retreat can occur stably. I haven’t seen evidence presented in this paper of about the likely future instability of Fleming, so tighten up language.

We apologize for the poor wording. However, it is well known that if grounding line retreat occurs on ice resting on a retrograde bed, this can trigger a positive feedback loop of flow acceleration, dynamic thinning, increased calving, mass loss and further grounding line retreat

(“marine ice sheet instability” or “tidewater instability”) (e.g. DeConto and Pollard, 2016; Favier et al., 2014; Rignot et al., 2014; Schoof, 2012). The subglacial topography of Fleming Glacier reveals a trough 3-4 km upstream of the current grounding line, which has such a retrograde bed. This suggests that once the grounding line has exceeded the edge of the trough, instability is possible. Especially since further oceanic forcing is likely to continue in the future (see answer P12L18), this is a likely future scenario for Fleming Glacier. We have changed the sentence into:

Hence, if grounding line retreat exceeds the edge of this trough, destabilisation like on Thwaites Glacier and in the Pine Island Bay region (Favier et al., 2014; Rignot et al., 2014) is possible, which would involve further rapid grounding line retreat and amplified mass loss in the future.

P11 L12 – edit Thwaites

We have changed this accordingly.

P11 L18 – Again I feel the analysis here is overly simplistic. Wouters et al were the first to present the rapid thinning rates and mass loss from the Western Palmer Land region, but subsequent publications (Hogg et al 2017) have shown that only ~30% of this should be attributed to ocean induced dynamic imbalance. Revise text to reflect known complexity. Discussion general – the authors have stated results from other regions of WAIS/AP, however this really isn't tied very coherently into how this impacts on the results they have presented on Fleming.

We understand the concerns of the reviewer. We have added a sentence on the publication of Hogg et al. (2017):

For the glaciers on Western Palmer Land Hogg et al. (2017) showed that ~ 35% of the ice loss after 2009 can be attributed to dynamic thinning triggered by ocean driven melt.

Additionally, we have substantially rewritten the discussion section, following the suggestions of reviewer #1 and #2. The timing of acceleration found by us is coincident with the timing of recently published exceptional warm CDW intrusions in 2008/2009 and 2010/2011 to Wordie Bay due to phases of strong westerly winds during strong La Niña/+SAM (positive Southern Annular Mode) events. Hence, we could substantiate the link between ocean forcing and our observed dynamic changes.

P12 L10 – as previously stated I do not think the authors have proven dynamic instability on Fleming. Imbalance maybe, but instability, no. Equally, attributing the signal to ocean induced dynamic forcing without properly evaluating any oceanographic or atmospheric data is poor. These interlinked processes are very complex, and really hard to disentangle. Although its entirely plausible that ocean forcing is responsible, I do not think the analysis presented in this paper has proven it.

See answer P11L18

P12 L18 – ‘ocean forcing is likely to continue’. Do the authors present any evidence to support this statement, or is it just a guess?

We thank the reviewer for this important remark. Oceanic forcing is likely to continue, since the SAM is expected to be shifted further poleward, which would foster conditions like in 2008/2009 and 2010/2011. We have changed the sentence into:

Pronounced oceanic forcing is likely to continue, since the SAM is forecasted to be shifted further poleward, which will foster conditions like those during the strong La Niña/+SAM events in 2008/2009 and 2010/2011 (Abram et al., 2014; Fogt et al., 2011; Walker and Gardner, 2017). Thus, further retreat of the grounding line and more dynamic thinning are expected on Fleming Glacier.

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Recent dynamic changes on Fleming Glacier after the disintegration of Wordie Ice Shelf, Antarctic Peninsula

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Abstract. The Antarctic Peninsula is one of the world's regions most affected ~~regions~~ by climate change. Several ice shelves retreated, thinned or completely disintegrated during recent decades, leading to acceleration and increased calving of their tributary glaciers. Wordie Ice Shelf, located in Marguerite Bay at the south-western side of the Antarctic Peninsula, completely disintegrated in a series of events between the 1960s early 1970s and the late 1990s. We investigate the long-term dynamics-response (1994–2016) of Fleming Glacier after the disintegration of Wordie Ice Shelf by analysing various multi-sensor remote sensing datasets. We present a dense time series of Synthetic Aperture Radar (SAR) surface velocities that reveals a rapid acceleration of Fleming Glacier in 2008 and a phase of further gradual acceleration and upstream propagation of high velocities in 2010–2011. The timing in acceleration correlates with strong upwelling events of warm Circumpolar Deep Water (CDW) into Wordie Bay, most likely leading to increased submarine melt. This, together with continuous dynamic thinning and a deep subglacial trough with a retrograde bed slope close to the terminus probably has induced unpinning of the glacier tongue in 2008 and gradual grounding line retreat between 2010 and 2011. Our data suggest that the glacier's grounding line had retreated by ~6–9 km between 1996 and 2011, which caused ~56 km² of the glacier tongue to go afloat. The resulting reduction in buttressing explains a remarkable median speedup of ~1.3 m d⁻¹ (~30 %) between 2007 and 2011, which we observed along a centreline extending between the grounding line in 1996 and ~16 km upstream. ~~Our analysis reveals that after two decades of accelerated glacier flow and dynamic thinning the glacier tongue partially ungrounded between January and March 2008. From 2010 to 2011 a further phase of gradual grounding line recession was observed. In total, the retreat of the grounding line between 2008 and 2014 amounted to ~6–9 km and caused ~68 km² of the glacier tongue to go afloat. We attribute this to continuous dynamic thinning and pronounced basal melt at the grounding line, probably by a south-western Antarctic Peninsula wide oceanic warming. The bedrock topography revealed that a deep subglacial trough facilitated the grounding line retreat. In response to the ungrounding of the glacier tongue we observed an upstream propagation of high surface velocities. This resulted in a median speedup along the glacier's centreline of ~1.4 m d⁻¹ (~29 %) between 2007 and 2011. The propagation of acceleration has not yet affected regions far upstream (~50 km) of the glacier.~~ Current median ice thinning rates (2011–2014) along profiles in areas below 1000 m altitude range between ~2.6 to 3.24 m a⁻¹ and are ~60–70 % higher than between 2004 and 2008. Our study shows

that Fleming Glacier is far away from approaching a new equilibrium and that the glacier dynamics are not primarily controlled by the loss of the ice shelf anymore. Currently, the tongue of Fleming Glacier is grounded in a zone of bedrock elevation between ~-400 and -500 m. However, about 3–4 km upstream modelled bedrock topography indicates a retrograde bed which transitions into a deep trough of up to ~1000 m at ~10 km upstream. Hence, this endangers upstream much larger ice masses, in the future which can significantly increase the contribution of Fleming Glacier to sea level rise in the future, and a huge potential for an increase in sea level rise contribution.

1 Introduction

Recent studies have shown that the Antarctic Peninsula ice masses are strong contributors to sea level rise. In a consolidated effort Shepherd et al. (2012) estimated the contribution between 2005 and 2010 to $36 \pm 10 \text{ Gt a}^{-1}$ corresponding to $0.1 \pm 0.03 \text{ mm a}^{-1}$ SLE (sea level equivalent). This is considerably higher than their reported ice mass loss for the period from 1992 to 2000 of $8 \pm 17 \text{ Gt a}^{-1}$ (Shepherd et al., 2012). Huss and Farinotti (2014) computed from their ice thickness reconstruction of the northern and central Antarctic Peninsula a maximum potential sea level rise contribution of $69 \pm 5 \text{ mm}$.

Rott et al. (2014) estimated the total dynamic ice mass loss for the glaciers along the Nordenskjöld Coast and the Sjögren-Boydell glaciers for the Larsen A and B region after ice shelf disintegration to be $4.21 \pm 0.37 \text{ Gt a}^{-1}$ between 2011–2013.

Seehaus et al. (2015; 2016) revealed similar values for tributary glaciers of the former Larsen-A and Prince-Gustav-Channel ice shelves. On the western Antarctic Peninsula south of $\sim 70^\circ$ increased ice discharge and considerable thinning rates have been reported for various ice shelf five tributaries (Wouters et al., 2015).

The main cause for the current increased ice discharge on the Antarctic Peninsula is the dynamic response of tributary glaciers to the disintegration and basal thinning of several ice shelves (e.g. Angelis and Skvarca, 2003; Pritchard et al., 2012; Rignot, 2006; Wouters et al., 2015; Wuite et al., 2015). With the reduction or loss of the buttressing effect of the ice shelves (Fürst et al., 2016; Mercer, 1978) due to thinning or disintegration, the tributary glaciers accelerate and show imbalance (Rignot et al., 2005; Rott et al., 2014; Scambos et al., 2004).

For the south-western Antarctic Peninsula Rignot et al. (2013) demonstrated that basal melt of George VI, Wilkins, Bach and Stange Ice Shelves exceeded the ablation induced by calving. For Wordie Ice Shelf high basal melt rates of $23.6 \pm 10 \text{ m a}^{-1}$ and $14.79 \pm 5.26 \text{ m a}^{-1}$ have been reported by Rignot et al. (2013) and Depoorter et al. (2013) respectively. However, the presented melt ratios (i.e. the ratio between basal melt and the sum of calving flux and basal melt) differ between 46 % (Rignot et al., 2013) and 82 % (Depoorter et al., 2013).

For the south-western Antarctic Peninsula Rignot et al. (2013) demonstrated that basal melt of ice shelves exceeded the ablation induced by calving. Wilkins Ice Shelf experienced amplified basal thinning controlled by due to small-scale coastal atmospheric and oceanic processes that assist ventilation of the sub-ice-shelf cavity by upper-ocean water masses (e.g. variations in wind stress or reduced freshwater fluxes from runoff and ice-shelf basal melt) until ~8 years before break-up events took place in 2008 and 2009 (Braun and Humbert, 2009; Padman et al., 2012). Subsequent changes in ice dynamics

and stresses leading to break-up have been observed (Rankl et al., 2016). On George VI Ice Shelf, surface lowering is linked to enhanced basal melt caused by an increased circulation of warmed Circumpolar Deep Water (CDW) (Holt et al., 2013) and a 13 % increase in ice flow was observed between 1992 and 2015 for its tributary glaciers (Hogg et al., 2017).

However, how the dynamic response after ice shelf loss progresses and how long this process lasts, is frequently unknown.

Seehaus et al. (2016; 2017) showed that significant temporal differences in the adaptation of glacier dynamics in response to ice shelf decay can occur and that those can only be resolved, if dense time series of satellite-based measurements are available. Wendt et al. (2010) also concluded for Wordie Ice Shelf that its former tributaries were still far from reaching a new equilibrium after ~~the~~ retreat and collapse of the ice shelf starting in the 1960's. However, given the limited data used in previous studies, Wendt et al. (2010) pointed out that a much closer monitoring is required to verify this. A recent comparison of stacked surface velocities of Fleming Glacier derived from InSAR in 2008 with velocities obtained from Landsat 8 feature tracking in 2014 and 2015 revealed that the glacier had sped up by $\sim 400\text{--}500\text{ m a}^{-1}$, which is the largest acceleration in ice flow recorded across all of Antarctica (Walker and Gardner, 2017; Zhao et al., 2017). However, the question why the magnitude of change was much higher than recorded elsewhere at the western Antarctic Peninsula over a similar time period remained unanswered so far.

In this study we investigate the glacier dynamics of Fleming Glacier after the disintegration of Wordie Ice Shelf on the south-western Antarctic Peninsula. Our study ties in with previous works in the region, but covers a much longer time period at a much higher temporal resolution. We provide a dense time series of ice velocity measurements from Synthetic Aperture Radar (SAR) satellite data for the time period 1994–2016 for Fleming Glacier. In order to investigate ~~explain~~ the observed changes in ice dynamics, we conducted an in-depth analysis of other geophysical and geodetic remote sensing data such as airborne Light Detection and Ranging (LiDAR) and satellite-borne laser altimetry, radio echo sounding for ice thickness, bistatic and monostatic SAR data as well as optical satellite images. We derive frontal retreat, surface velocity changes, ice elevation changes, grounding line positions and estimate the area of freely floating ice from hydrostatic equilibrium.

2 Study site

The former Wordie Ice Shelf was located in Marguerite Bay on the south-western Antarctic Peninsula. The ice shelf was originally fed by several major input units (Fig. 1). Among ~~these all former tributary glaciers~~, Fleming Glacier is the biggest. It has a current length of approx. 80 km and is up to 10 km wide at its tongue. With a speed of more than 8 m d^{-1} close to its calving front (Fig. 1), Fleming Glacier is also the fastest flowing glacier in Wordie Bay. Fleming Glacier merges with Seller and Airy Glacier $\sim 8\text{ km}$ upstream of their joint calving front. Together with Rotz Glacier, which merges with Seller Glacier $\sim 28\text{ km}$ upstream of the front, all four glaciers form the Airy-Rotz-Seller-Fleming glacier system, spanning a total catchment area of about 7000 km^2 (Cook et al., 2014).

Starting in the 1960s, Wordie Ice Shelf ran through a stepwise disintegration process (Fig. 1), which was controlled by pinning points (i.e. ice rises/rumples). Analyses of satellite imagery suggest that the ~~Here the~~ ice shelf was temporarily

grounded and stabilized at these pinning points until one of the next rapid ~~break-up~~break-up events took place (Doake and Vaughan, 1991; Reynolds, 1988; Vaughan, 1993; Vaughan and Doake, 1996). However, during phases of ice front retreat, instead of protecting the ice shelf against decay, ice rises that were embedded in the ice shelf appeared to behave as indenting wedges, contributing to weakening the ice shelf and accelerating break-up (Doake and Vaughan, 1991; Vaughan, 1993). -It is likely that the collapse of Wordie Ice Shelf was triggered by a combination of amplified ablation due to rising air temperatures (Doake and Vaughan, 1991), enhanced tidal action as a consequence of relaxed sea-ice conditions in Marguerite Bay (Reynolds, 1988) and increased basal melt rates on ice shelves ~~yesfs~~ in the Bellingshausen Sea due to rising ocean temperatures (Depoorter et al., 2013; Holland et al., 2010; Pritchard et al., 2012; Rignot et al., 2013).

After the last big disintegration event in 1989, the ice shelf was split into a northern and a southern part (Doake and Vaughan, 1991). In the early 1990s most of the remaining floating ice in Wordie Bay consisted only of the protruding, unconfined tongues of the disconnected tributary glaciers. These tongues disappeared between 1998 and 1999, so that in 1999 Fleming Glacier was already calving near its 1996 grounding line (Rignot et al., 2005). Wendt et al. (2010) found the remaining area of floating ice to be only 96 km² in 2009. At this time there was virtually no contiguous ice shelf left and only the glaciers of the Prospect unit and Harriot's unnamed neighbouring glacier still possessed floating glacier tongues (Wendt et al., 2010). ~~During the following~~ years (2010–2015) the fronts of the glaciers in Wordie Bay remained quite stable, except at the Prospect system where the once interconnected floating ice tongues of the three glaciers ~~got~~ disconnected and some floating ice ~~was got~~ lost. This resulted in an total area of 84 km² of ice shelf in Wordie Bay in 2015, if taking the grounding line of 1996 as a baseline (Rignot et al., 2005; Rignot et al., 2011a) and ignoring any grounding line migration.

While in the early 1990's an acceleration was not yet observed from the visual inspection of optical satellite imagery (Doake and Vaughan, 1991; Vaughan, 1993), ~~-~~Rignot et al. (2005) found substantial dynamic thinning and an increase of surface velocities by 40–50 % in 1996 against 3 point measurements by Doake (1975) in 1974 on Fleming Glacier (Fig. 1). The higher velocities as well as further thinning were also confirmed through Global Navigation Satellite System (GNSS) measurements in 2008 and airborne LiDAR surveys in 2004 and 2008 respectively (Wendt et al., 2010). Then, an Antarctic-wide unprecedented acceleration of ~ 400–500 m a⁻¹ was recorded between 2008 and 2014/2015 -(Walker and Gardner, 2017; Zhao et al., 2017).

3 Data

We used a broad remote sensing data set in order to investigate the ~~dynamic changes response in ice dynamics at of~~ Fleming Glacier between 1994 and 2016 after the to the retreat and disintegration of Wordie Ice Shelf. Tab. 1 gives an overview of the specifications and the time coverages of the sensors used. -The Bedmap2 digital elevation model (DEM) of Antarctica (Fretwell et al., 2013), resampled to 100 m resolution, was taken as a topographic reference for orthorectification of the surface velocity fields and for the derivation of local incidence angles required for the conversion from slant to ground range displacement. Over the Antarctic Peninsula the Bedmap2 DEM provides a seamless compilation of data from the improved

ASTER GDEM (from ASTER stereo images acquired between 2000 and 2009) (Cook et al., 2012), the SPIRIT DEM (from SPOT stereo images acquired in 2007 and 2008) and the NSIDC DEM (from ICESat data acquired between 2003 and 2005) (DiMarzio et al., 2007). The Bedmap2 digital elevation model (DEM) (Fretwell et al., 2013), resampled to 100 m resolution, was taken as a reference for geocoding and orthorectification of all SAR data products.

5 Calibrated and multi-looked SAR intensity images, Landsat 7 imagery and an existing dataset of ice shelf outlines (Ferrigno, 2008) were taken as a reference for the delineation of the ice shelf/glacier front (Tab. S1).

Calculations of elevation change rates were based on airborne LiDAR measurements, satellite-borne laser altimeter measurements and two DEMs derived from SAR interferometry (Tab. 1). The two DEMs covering the Airy-Rotz-Seller-Fleming glacier system were calculated from bistatic TerraSAR-X/TanDEM-X (TSX/TDX) Coregistered Single look Slant range Complex (CoSSC) strip map (SM) acquisitions on 2011-11-21 and monostatic TSX/TDX CoSSC SM acquisitions on 10 2014-11-03 (Fritz et al., 2012; Krieger et al., 2013). The TSX/TDX data were selected as close as possible to the dates of two NASA Operation IceBridge (OIB) Airborne Topographic Mapper (ATM) flights in 2011 and 2014, in order to be able to correct for radar penetration depth biases.

A simulated phase from a subset of the TanDEM-X global DEM with a spatial resolution of 12 m (Rizzoli et al., 2017) was used to facilitate phase unwrapping during the generation of the two TSX/TDX DEMs. The TanDEM-X global DEM was also used as a reference for the absolute height adjustment of the TSX/TDX DEMs.

For the determination of the floating area on the tongue of Fleming Glacier, we used information on ice thickness and surface elevation from several Pre-IceBridge (PIB) and OIB flight lines across the Airy-Rotz-Seller-Fleming glacier system between 2002-11-26 and 2014-11-16 (Tab. 1). Depending on the date of acquisition, ice thickness data was recorded by 20 different versions of the Coherent Radar Depth Sounder (CoRDS) (Tab. 1).

Our estimation of the recent grounding line of Fleming Glacier was based on a combination of the information on hydrostatic equilibrium with bedrock topography data, profiles of surface velocities and elevation change rate patterns inferred from the 2011–2014 TSX/TDX data. Information on bedrock topography was taken from the modelled bedrock grid of Huss and Farinotti (2014), which represents the most detailed dataset on bedrock topography available for the Antarctic Peninsula. The dataset was generated by subtracting modelled ice thickness from the improved ASTER GDEM by Cook et al. (2012). Ice thickness was derived by constraining a simple model based on the shallow ice approximation for ice dynamics with observational data of ice thickness (OIB) and surface velocity (Rignot et al., 2011b). Where available the modelled ice thickness was corrected with -OIB ice thickness data, leading to more precise ice thickness values in such areas. On average, the local uncertainty in ice thickness of the dataset is $\pm 95\text{m}$, but values can reach $\pm 500\text{m}$ for deep troughs without nearby OIB measurements. Since OIB coverage is fairly good across Fleming Glacier, uncertainties in modelled ice thickness are relatively low in this area. However, a comparison of bedrock elevations from Huss and Farinotti (2014) with bottom elevations calculated from OIB ATM and CoRDS measurements shows that although the modelled bedrock reflects the general subglacial topography well, the absolute difference in bottom elevation can be even more than 100 m (Fig. S5). One possible reason for this is a difference between ATM heights and the refined ASTER GDEM, which transfers to

~~bedrock elevation. However, although the Huss and Farinotti dataset reflects the general subglacial topography well, it has to be considered that its absolute bottom elevation values often differ by even more than 100 m from the OIB *CoRDS measurements.~~

4 Methods

5 4.1 Surface velocities

For each sensor consecutive pairs of coregistered single look complex SAR images were processed using an intensity offset tracking algorithm (Strozzi et al., 2002). A moving window was used to calculate surface displacements in azimuth and slant range direction between two SAR intensity images by localizing the peaks of an intensity-cross correlation function. The technique requires the definition of a tracking patch size and a step size (i.e. the distances in range and azimuth between the centres of two consecutive moving windows, Tab. 1). The parameters were chosen according to the sensor specifications, the temporal baseline between the acquisitions and the expected displacement.

During the tracking procedure the implementation of a cross correlation threshold of 0.05 assured the removal of low quality offset estimates. Post-processing of the velocity fields comprises an additional filtering (Burgess et al., 2012) based on the comparison of the orientation and magnitude of the displacement vectors relative to their surrounding vectors. This algorithm discards over ~~99~~ 90 % of unreasonable tracking results. The filtered displacement fields were then transferred from slant range geometry into ground range geometry, geocoded and orthorectified. The procedure to determine the error of the velocity measurements is described in the Supplemental material, Sect. S24. The errors for each velocity field and the proportion of velocity vectors removed by the filter are listed in Tab. S3.

4.2 Elevation change

We derived ice thinning rates on Fleming Glacier for 2004–2008 and 2011–2014 by ~~comparing subtracting~~ ellipsoid heights of the PIB (ATM, 2004), the Centro de Estudios Científicos Airborne Mapping System (CAMS, 2008) and the OIB (ATM, 2011, 2014) airborne LiDAR datasets. The vertical accuracy of the ATM elevation data is estimated to be better than 0.1 m (Krabill et al., 2002; Martin et al., 2012). For the CAMS data, vertical accuracy is 0.2 m (Wendt et al. 2010). Before subtraction, overlapping data of the originally closely spaced measurements were condensed to a common set of median surface elevations with an equal spacing of 50 m in along and across track direction. The locations of the resulting points ~~clouds~~ of differential elevation measurements are shown in Fig. 4. ~~As described in Wendt et al. (2010), we estimated the vertical accuracy of the measurements to be ± 0.2 m for all flights.~~

Additionally to the airborne LiDAR measurements ice elevation change rates for the period 2004–2008 were calculated from ellipsoid heights measured by the Ice, Cloud and Land Elevation Satellite (ICESat). Saturation of the 1064 nm -Geoscience Laser Altimeter System (GLAS) detector can occur over ice, leading to a distorted echo waveform (Schutz et al., 2005). Hence we applied a saturation elevation correction provided on the GLA12 product prior to subtracting both tracks and

excluded elevation measurements with flagged invalid saturation correction values from our analyses. Since both repeat tracks are not overlapping but separated by ~ 150 m in across track direction, we linearly interpolated the elevation data of the 2004 track onto the latitude values of the 2008 data prior to subtraction, ~~similar to~~ as described in Fricker and Padman (2006). In order to keep the error induced by interpolation low, elevation values were only allowed to be interpolated between two footprint centre locations with an along track spacing of ~ 170 m. This assured that existing gaps in the real data were preserved. Shuman et al. (2006) report a relative accuracy of ± 0.25 m for ICESat elevations measured on surface slopes between 1.5 and 2.0° . The mean surface slopes along the two ICESat elevation profiles were 1.9° . Hence, taking into account further possible inaccuracies of ± 0.15 m due to interpolation, we estimated the accuracy of the ICESat ice ~~elevations thinning rates~~ to be ± 0.4 m.

A map of elevation change rates between 2011 and 2014 was calculated by differencing two TSX/TDX-DEMs. Both DEMs have a spatial resolution of 10 m. To generate the DEMs we applied a differential interferometric approach, which facilitates phase unwrapping by incorporating the topographic information of a reference DEM (Vijay and Braun, 2016). A subset of the TanDEM-X global DEM, covering the two TSX/TDX-DEMs, was chosen to be the reference DEM. ~~In order to stay consistent with the other SAR measurements, the resampled Bedmap2 DEM was chosen to be the reference DEM.~~

Before differencing, the TSX/TDX-DEMs must be vertically referenced. For this purpose the median vertical offset between the DEMs and the TanDEM-X global DEM was measured over stable areas (i.e. tops of nunataks and rock outcrops, which were not affected by image distortions) at altitudes between 150 m and 1000 m (Fig. S1), before both DEMs were adjusted accordingly. ~~Before differencing, the TSX/TDX DEMs must be vertically referenced to each other. For this purpose the median vertical offset between the DEMs was measured on preferably ice berg free areas of sea ice (Fig. S1), before one of the DEMs was adjusted accordingly. Neglecting differences in sea ice thickness, we corrected the amount of vertical adjustment for differences in tide level between the dates of the SAR acquisitions by using the TPX08 tide model (Egbert and Erofeeva, 2002, updated). This model showed best performance for the prediction of tides in Antarctic seas (Stammer et al., 2014).~~ After subtracting the vertically registered DEMs, the elevation differences were converted into yearly elevation change rates. We assessed the accuracy of the vertical registration over another set of stable areas ~~(i.e. tops of nunataks and rock outcrops, which were not affected by image distortions)~~ at altitudes between 150400 m and 1300 m (Fig. S1). The absolute median value of the extracted change rates was 0.370.43 m a⁻¹ ~~and which primarily~~ accounts for errors related to the vertical registration. ~~and the tidal correction.~~

However, since radar signals can penetrate several meters into snow and ice, depending on the radar frequency and the dielectricity of the medium (Mätzler, 1987; Rignot et al., 2001), an additional bias is induced on glaciated areas when differencing interferometric DEMs from different times and/or frequencies (Berthier et al., 2016; Seehaus et al., 2015; Vijay and Braun, 2016). Since the TSX/TDX data was acquired only 4–7 days apart from the ATM data, differences in elevation change rates between the two datasets can be primarily attributed to differences in penetration depth at the TSX/TDX acquisitions in 2011 and 2014 and remaining vertical registration errors. In order to compare the TSX/TDX data with the ATM data, we extracted the TSX/TDX elevation change rates at the locations of the differential OIB ATM measurements

using a buffer with a radius of 25 m ~~buffer~~ and calculated the median for each point. Hypsometric reference values were taken from the resampled Bedmap2 DEM which we converted to ellipsoidal heights using the included geoid correction layer. The comparison between elevation change rates obtained from the 2011–2014 OIB ATM flights and the 2011–2014 TSX/TDX data after the vertical registration of the DEMs showed a maximum overestimation of ice thinning of 1.25 m a⁻¹ for the TSX/TDX measurements (Fig. S4 a, b). However, the general trend of the elevation change rates fits well to those calculated from the LiDAR data and significant differences in elevation change were only measured in the lower areas of the glacier tongue. In the upper areas (above 600 m altitude) the difference between ATM and TSX/TDX elevation change rates was close to 0 m. Here the snow volume was likely completely frozen on both dates of acquisition, so that the penetration bias cancelled out. A backscatter comparison showed lower values in 2014 than in 2011 in areas below 600 m altitude, whereas the backscatter in the upper areas above 600 m altitude was similar for both dates (Fig. S4 d). The comparison between elevation change rates obtained from the 2011–2014 OIB ATM flights and the 2011–2014 TSX/TDX data after the vertical registration of the DEMs showed a systematic overestimation of mean ice thinning of up to 2 m a⁻¹ for the TSX/TDX measurements (Fig. S4 a). However, the general trend of the elevation change rates fits well to those calculated from the LiDAR data. We corrected the TSX/TDX data with a local polynomial model based on the elevation change rate differences between the ATM and the TSX/TDX data (Fig. S4 b). We applied this correction to all glaciated areas below 1000 m and clipped the TSX/TDX elevation change rate map accordingly. The RMSE between the cubic fits of the ATM elevation change rates and the extracted values from the corrected TSX/TDX map was 0.02 m (Fig. S4 c). However, by taking into account unknown errors due to the extrapolation of the correction factors to the entire glacier area, we assumed a remaining error of $\pm 0.2 \text{ m a}^{-1}$ related to penetration depth differences. Together with the error of vertical registration ~~and tidal correction~~, this resulted in a total error of $\pm 0.5763 \text{ m a}^{-1}$ for the TSX/TDX ice thinning rates.

For our analyses of elevation change we compared ice thinning rates from the PIB and the CAMS CECS ~~airborne~~ laser altimeter data (2004–2008) with rates obtained from the OIB data (2011–2014) as well as elevation change rates from the TSX/TDX data (2011–2014) with those derived from ICESat in 2004 and 2008. For the comparison between the TSX/TDX and the ICESat data, ice thinning rates were extracted from the TSX/TDX map at the GLAS centre locations of the 2008-10-04 track. To take into account the 70 m footprint of the GLAS instrument, we applied a buffer with a radius of 35 m and calculated the median from the extracted values at each point.

4.3 Floating area (hydrostatic height anomalies) and estimation of recent grounding line

In order to determine the floating area on the tongue of the Airy-Rotz-Seller-Fleming glacier system at different points in time, we derived hydrostatic height anomalies Δe from the PIB and OIB elevation and ice thickness measurements between 2002 and 2014. For every measuring point of ice thickness, Δe was calculated similar to Fricker (2002) by subtracting a theoretical freeboard height in hydrostatic equilibrium e_{he} from a measured orthometric ATM ice surface elevation e :

$$\Delta e = e - e_{he} \quad (1)$$

Regions on the glacier tongue where $\Delta e \approx 0$ were considered as freely floating. Before deriving hydrostatic height anomalies, we merged simultaneously acquired ice thickness and ATM data by calculating the median elevation within a buffer of 50 m at each ice thickness measurement. As the ATM heights were originally measured relative to the WGS84 ellipsoid, we converted the ellipsoidal ATM values to orthometric heights prior to the buoyancy calculations. For the conversion we used kriged geoid values calculated for a mean tide system with the EIGEN-6C4 global gravity field model (Förste et al., 2014). We calculated e_{he} by applying a modified formula after Griggs and Bamber (2011):

$$e_{he} = (H_i + \delta) - \frac{H_i \rho_i}{\rho_w} \quad (2)$$

where H_i is the measured PIB or OIB ice thickness, i.e. the ice thickness derived under the assumption that all ice is homogeneous and firm free, ρ_i is the ice density of pure ice, ρ_w is the density of sea water and δ is the firm density correction factor, i.e. the difference between the actual thickness of the firm layer above the glacier ice and the thickness that the firm would have if it were at the density of pure ice (Griggs and Bamber, 2011). Details on the uncertainties of all variables used for the calculations as well as the assessment of error propagation are provided in the Supplemental material, Sect. S32.

The buoyancy calculations provided information on the limit of hydrostatic equilibrium between 2002 and 2014 at several locations on the glacier system. As demonstrated in Seehaus et al. (2015), clear patterns of low or positive ice thinning rates in dh/dt maps reveal areas of floating ice, since buoyancy can cause originally grounded ice to bounce and/or ~~mitigates~~ decreases the effect of ice thinning on ice surface elevation to ~10 %.

Moreover, the grounding line marks the transition between two fundamentally different flow regimes of grounded and freely floating ice. Whereas the flow dynamics of grounded ice are dominated by vertical shear and controlled by basal drag, flow of floating ice is drag free and dominated by longitudinal stretching and lateral shear (Schoof, 2007). The difference in flow dynamics of grounded and floating ice can result in pronounced changes in surface velocity close to the grounding line (Stearns, 2007; Stearns, 2011). Furthermore, Rignot et al. (2002) demonstrated that if ungrounding occurs, the resulting flow acceleration usually affects both the floating and the grounded part of the glacier, but is largest near the grounding zone. as basal friction is a major deterrent of surface velocities, Thus, velocity profiles can serve as additional information for locating the grounding line.-(Stearns, 2007; Stearns, 2011).

Bedrock elevation data can reveal subglacial topographic features which act as pinning points for the glacier. Hence, our estimations of recent and previous grounding line positions ~~were~~ based on information on hydrostatic equilibrium from the hydrostatic height anomaly calculations as well as maps and profiles of TDX/TSX 2011–2014 elevation change rates, modelled bedrock topography (Huss and Farinotti, 2014) and surface velocities. Wherever possible, we gave preference to information on hydrostatic equilibrium for the final decision of the recent grounding line location. For selected profiles across the glacier, recent and previous (2008) grounding line positions were estimated by combining evidence from elevation change, bedrock topography and surface velocity. In the remaining areas, the recent grounding line was deduced from combining information on elevation change rate patterns in the TDX/TSX 2011–2014 dh/dt map with information on bedrock topography.

5 Results

5.1 Surface velocities

Figure 2 shows the multi sensor time series (1994–2016) of SAR intensity tracking derived velocities along a centreline profile on Fleming Glacier (Fig. 1). The profile extends from the grounding line location in 1996 to 16 km upstream. The relative distance of the glacier front to the 1996 grounding line is shown on the left side of the plot. Distances in the subsequent text are given in reference to the grounding line of 1996. Positive values relate to positions on the glacier upstream of (behind) the former grounding line, while negative values refer to locations seawards of the 1996 grounding line.

After 1999 the glacier front remained comparatively ~~steady~~ close to the grounding line location in 1996 for almost 10 years.

~~Figure 2 shows that glacier velocities were rather stable between 1994 and 2007. The normalised median absolute deviation (NMAD) of the median velocities during this time was 0.06 m d^{-1} . A comparison of the velocities on 1995-10-27 and 2007-10-23 (Fig. 3a) along the centreline profile reveals that the median velocity difference between 1995 and 2007 was just 0.04 m d^{-1} . No obvious sign of velocity change is visible on our time series between the first measurements in 1994 and 2007. A comparison of the velocities on 1995-10-27 and 2007-10-23 (Fig. 3a) along the centreline profile reveals that the median velocity difference between 1995 and 2007 was just 0.04 m d^{-1} .~~

Between January and April 2008 a rapid acceleration of Fleming Glacier was noticeable along the centreline, which propagated ~8 km inland. Simultaneously, the front of Fleming Glacier retreated behind the 1996 grounding line for the first time. ~~Although, since 2008 the glacier tongue never exceeded the 1996 grounding line anymore, the front remained quite stable.~~ The velocity pattern persisted until March 2010, when a second phase of acceleration began. Our velocity time series shows that the acceleration gradually propagated further inland within one year, until it reached its final extension ~12.4 km upstream in early 2011. ~~Fig. 3a, b show absolute and relative velocity changes along the centreline profile for the periods 1995–2007, 2007–2011 and 2011–2015. In 2011 the location of the glacier front reached its most inland position. Consequently velocities measured on sea ice or ice mélange in 2011 (seaward of the 2011 front) were excluded from our analyses. Large changes in surface velocity close to the 2011 front in the periods 2007–2011 and 2011–2015 do not represent real dynamic change, but result from comparing the inherently higher frontal velocities in 2011 with lower velocities of the floating glacier tongue in 2007 and 2015.~~ Between 2007-10-23 and 2011-10-02 the ~~median~~ increase in median surface velocity along the profile was ~~~1.41.3~~ m d^{-1} or ~~~2930~~ % (Fig. 3a, b). If ignoring velocity change in the vicinity of the 2011 glacier front, the highest relative acceleration values (~32–35 %) were recorded between ~7 and ~11 km upstream of the 1996 grounding line (Fig. 3b). The amount of relative acceleration rises significantly at ~7 km and abruptly drops at ~11 km. Peak absolute acceleration values of ~1.6 m d^{-1} were found at ~8 km. If excluding measurements ~~at in the region of the 2011 front~~ frontal change between 2011 and 2015, no further marked changes in velocities ~~were could be~~ detected along our centreline profile after 2011 (Fig. 3 a, b, Fig. 1). Surface velocities on 2013-12-24 at the three measuring sites of Doake

(1975) ~50 km upstream (Fig. 1) were very similar to those measured in 1996 and 2008. Furthermore, the flow directions in 2013 were like those in 2008 and 1974 (Tab. 2). However, surface velocities derived from Landsat 8 feature tracking suggest that in 2015 velocities at the three measuring sites had increased by ~20% in comparison to 2008 (Walker and Gardner, 2017; Zhao et al., 2017) (Tab. 2).

5.2 Elevation change

Figure 4 shows elevation change rates on the Airy-Rotz-Seller-Fleming glacier system for the period between 2011 and 2014. The entire area undergoes a considerable drawdown. On Fleming Glacier the highest ice thinning rates were recorded in a zone extending from ~8 to ~14 km upstream. On Seller Glacier the ice loss exceeds ~6 m a⁻¹ at about 7 km upstream. In general, ice thinning decreases towards higher altitudes. A tendency to lower negative or even positive elevation change rates was observed on the lower parts of the joint Fleming and Seller glacier tongue between 0 and up to ~9 km upstream. The pattern was not as clear as on Airy Glacier, where a distinct area of low ice thinning rates positive elevation change was detected between 0 and ~4 km upstream.

Figures 5 a) and b) show comparisons of elevation change rates for the times prior to (2004–2008) and after the glacier acceleration (2011–2014). The location of the data is shown in Fig. 4. In Fig. 5 a) elevation change rates from PIB ATM-CAMS measurements (2004–2008) are plotted together with rates from ATM measurements in 2011 and 2014. The large scattering of the data is due to the highly crevassed surface of the glacier tongue, where a purely horizontal displacement of crevasses can cause apparent positive and negative elevation differences. Therefore, a median filter was applied to the data before adjustment of a cubic function. Fig. 5 b) shows ice thinning rates from ICESat tracks in 2004 and 2008 together with rates calculated from 2011–2014 TSX/TDX data. Note that the ATM data in Fig. 5 a) and the ICESat data in Fig. 5 b) refer to different profiles.

Figure 5 shows that prior to the speedup in 2008, Fleming Glacier has already been affected by pronounced surface lowering. Figure 5 shows that prior to the speedup (2004–2008) Fleming Glacier was already affected by pronounced surface lowering. A clear trend of increasing ice thinning rates towards the glacier front is visible for 2004–2008 on both profiles.

During this time period the maximum negative elevation change rates were found close to the 1996 grounding line. Here the cubic regression functions imply that the ice surface lowered at a maximum of $\sim 3.84 \pm 0.2$ m a⁻¹ for the CAMS-ATM measurements and at $\sim 4.6 \pm 0.4$ m a⁻¹ for the ICESat data. For all median elevation change rates presented below, we calculated the NMAD in order to account for the statistical dispersion of the input data. The median ice thinning rates measured during 2004–2008 for the cubic fits were 1.5 ± 0.62 m a⁻¹ on the CAMS-ATM flightpath and 1.9 ± 10.4 m a⁻¹ on the ICESat track. This shows that Fleming Glacier did not show any sign of reaching a new equilibrium even almost 20 years after the partial disintegration of the ice shelf in 1989. The OIB ATM and the TSX/TDX elevation change rates between 2011 and 2014 reveal a significant change in pattern for the time after the glacier flow acceleration. A tendency to lower ice thinning rates is present towards the glacier front and high negative elevation change rates can be found in a zone between

10—15 km upstream, with maximum ice losses of $\sim 3.74 \pm 0.2$ m a⁻¹ for the ATM, and $\sim 3.8 \pm 0.63$ m a⁻¹ for the TSX/TDX cubic regression functions. The median elevation change rates were -3.24 ± 0.683 m a⁻¹ and -2.6 ± 1.2 m a⁻¹ for the cubic fits of the 2011—2014 TSX/TDX data (Fig. 5 b), extracted along the ICESat track and, and -2.6 ± 0.2 m a⁻¹ the on the 2011—2014 ATM data (Fig. 5 a), respectively. flight path. Despite of lower ice thinning rates measured towards the ice front in 2011—2014, our data show an overall median increase of median ice thinning rates along the profiles of ~ 1.10 – 1.32 m a⁻¹ or ~ 60 –70 % between the periods from 2004 to 2008 and from 2011 to 2014. However, in some areas 10–15 km upstream, ice thinning rates even doubled in the latter period.

5.3 Floating area (hydrostatic height anomalies) and estimation of recent grounding line

Figure 6 depicts the results of the hydrostatic height anomaly calculations from PIB and OIB elevation and ice thickness data acquired before (2002–2004) and after the speedup of Fleming Glacier (2011–2014). Detailed plots showing the results of the hydrostatic height anomaly calculations along PIB and OIB flight lines can be found in the Supplemental Material, Fig. S5 a–e.

The hydrostatic height anomaly data of 2002 and 2004 (Fig. 6, Track 1 and 2) clearly reveal that the ice inland of the 1996 grounding line was not floating at these times. However, the same calculations for data acquired in 2011 and 2014 (Fig. 6, Track 3–5) as well as patterns of low and positive elevation change rates in the TSX/TDX 2011–2014 dh/dt map (Fig. 4, S7) suggest that after the final stage of glacier acceleration in 2011 an area of about 56 km² (referring to the front in 2014) of the formerly grounded glacier tongue of the Airy Rotz Seller Fleming system had been afloat. However, the same calculations for data acquired in 2011 and 2014 as well as patterns of low and positive elevation change rates in the TSX/TDX 2011–2014 dh/dt map prove that a vast part of the formerly grounded glacier tongue had been freely floating after the final stage of glacier acceleration in 2011. In 2014 the estimated total freely floating area of the Airy Rotz Seller Fleming glacier tongue was ~ 68 km².

The bedrock elevation model of Huss and Farinotti (2014) exhibits, that the boundary of the area showing flotation follows bedrock ridges (Fig. 6). Those confine a subglacial trough underneath the Airy-Rotz-Seller-Fleming glacier system. The ridges reach up to ~ 9 km upstream of are located several kilometres behind the 1996 grounding line. For most regions of the glacier tongue we estimate the current grounding line to coincide with these ridges at an elevation between ~ -400 and -500 m. On Fleming and Seller Glacier our estimation of the recent grounding line also largely coincides with the extent of lower ice elevation change rates apparent in the TSX/TDX 2011–2014 dh/dt map (Fig. S7). However, on Airy Glacier a distinct area of low ice thinning rates on the lower part of the glacier tongue indicates floatation (Fig. 4, S7), whereas hydrostatic equilibrium in 2011 suggests that the glacier is grounded on a hump which reaches to the subglacial trough (Fig. 6).

We extracted data of surface velocities, TDX/TSX 2011–2014 elevation change rates and, bedrock topography and if available hydrostatic height anomalies along four profiles on Airy and Fleming Glacier in order to estimate recent grounding line positions and those in 2008 after the first acceleration phase (Fig. S6 a–d). The locations of the profiles as well as the deduced grounding line locations are shown in Fig. 6. After the first acceleration phase in 2008 the front of Fleming Glacier

had retreated behind the 1996 grounding line for the first time. Hence, on Fleming Glacier the grounding line must have been situated upstream of the 1996 position at this time. The profile-plots in Fig. S6 b, c suggest that after the first acceleration phase in 2008 the grounding line ~~was not located as far upstream had not yet retreated as far~~ as after the second acceleration phase between 2010 and 2011. ~~However, since the estimation of the 2008 grounding line positions was based on surface velocities and modelled bedrock topography only, their precise locations remain unclear. Hence, the 2008 grounding line positions indicated in Fig. 6 are just a best guess based on the data we have in hand. A more detailed discussion on how the grounding line positions were finally decided from the profiles is provided in the Supplemental Material, Fig. S6 a-d. For Seller Glacier no information on the 2008 grounding line position was deducible. For each profile possible and likely grounding line positions after the first acceleration in 2008 and today are shown in Fig. 6. All in all, we estimated the current grounding line of Fleming Glacier to be located ~6–9 km upstream of its 1996 position. Its likely recent location is consistent with the maximum extent of upstream propagation of high velocities in 2008 on the centreline profile (Fig. 6). The plots of the extracted data as well as more detailed information on the interpretation of the data in respect of the estimation of recent and previous grounding line positions are provided in the Supplemental Material, Fig. S6 1–4.~~

~~On the central part of Fleming Glacier velocities extracted along the centreline profile as well as along profile 3 (Fig. 6, Fig. S6 3) show that in 2011 high velocities had propagated upstream until to a chain of bedrock hills located up to 13 km upstream. However, hydrostatic height anomalies calculated from 2014 OIB data as well as the TSX/TDX elevation change rate map show no indication of ice floatation further upstream than ~6–9 km. This location is consistent with the propagation of high velocities in 2008 on the centreline profile and is estimated to be the likely recent grounding line position. A possible location of the grounding line after the initial ungrounding in 2008, however, is a gentle hill, ~2.5 km upstream of the grounding line in 1996.~~

6 Discussion

Our results confirm the previously detected acceleration of Fleming Glacier ~~in response to the until 2007 after the~~ stepwise break-up and disintegration of Wordie Ice Shelf (Rignot et al. 2005). ~~Median elevation change rates of -1.5 m a^{-1} and -1.9 m a^{-1} between 2004 and 2008 may suggest that Fleming Glacier had not reached a new equilibrium even almost 20 years after the partial disintegration of the ice shelf in 1989. Nevertheless, our dense velocity time series shows that surface velocities remained fairly stable between 1994 and 2007.~~

~~Between January and April 2008 the glacier had abruptly accelerated and high velocities had propagated upstream. Between March 2010 and early 2011 a second phase of acceleration was detected during which the speedup gradually propagated further upstream. The remarkable median speedup of $\sim 1.3 \text{ m d}^{-1}$ which we recorded between 2007 and 2011 is in good agreement with an acceleration of $\sim 400\text{--}500 \text{ m a}^{-1}$ reported by Walker and Gardner (2017) and Zhao et al. (2017) for the period 2008–2014/2015. No~~However, a comparison of our velocities in 2013 with their velocities in 2015 at the three

measuring sites of Doake (1975) ~50 km upstream suggests that ~~significant acceleration since 1996 could be found at the three measuring sites of Doake (1975) ~ 50 km glacier upstream, indicating the recent speedup had~~ not yet propagated up until to these locations prior to 2015. Like Wendt et al. (2010), we also cannot confirm the change in flow direction in 1996 proposed by Rignot et al. (2005).

5 Abrupt speedups of tributary glaciers are often recorded as a direct consequence of loss of the buttressing force or major calving events (e.g. Seehaus et al., 2015). However, we did not observe any major calving event, which could have been responsible for the observed acceleration in 2008 or afterwards. In Greenland seasonal velocity fluctuations have been linked to both enhanced basal sliding due to the penetration of surface melt water to the ice-bedrock interface and inter-annual differences in drainage efficiency (Moon et al., 2014; Sundal et al., 2011; Zwally et al., 2002). However, the meltwater
10 production on Fleming Glacier is considered to be generally not sufficient to percolate to the glacier bed (Rignot et al., 2005). Furthermore, a trend of cooling air temperatures is reported for the Antarctic Peninsula since the end of the 1990s (e.g. Turner et al., 2016). Although decadal mean surface temperatures in the period 2006–2015 were 0.2 °C higher than in 1996–2005 at San Martin station (~120 km north of Wordie Ice Shelf), warming rates have decreased markedly since the decade 1996–2005 and show a cooling trend in 2006–2015 (Oliva et al., 2017). This may have further reduced surface melt
15 during recent years. All in all, enhanced basal sliding due to percolating meltwater is likely not the explanation for the observed increase in flow velocities. However, we do not rule out that as a consequence of the acceleration, basal sliding increased in grounded areas by meltwater generated from greater basal frictional heat, which likely further reduces surface melt. A possibly enhanced basal sliding as e.g. reported from Greenland (e.g. Moon et al., 2014) can hence be ruled out as explanation for the increase in flow velocities. Hydrostatic height anomalies calculated from OIB ice thickness and surface
20 elevation data, TSX/TDX elevation change rates, surface velocities and modelled bedrock topography suggest that the current grounding line of Fleming Glacier is located ~6-9 km upstream of its 1996 position, following the edges of a subglacial trough. When ungrounding causes parts of the glacier tongue to go afloat, buttressing and basal friction is reduced. This in turn provokes the glacier to speed up and to dynamically thin. We propose that unpinning and grounding line retreat are the main causes of the observed strong acceleration of Fleming Glacier.

25 Fairly stable velocities between 1994 and 2007 as well as hydrostatic height anomalies in 2002 and 2004 do not indicate that ungrounding from the 1996 grounding line position had happened prior to 2008. Although we were not able to give a precise estimate of the grounding line in 2008, the fact that during the acceleration phase in 2008 the glacier front had retreated behind the 1996 grounding line for the first time, shows that the 2008 grounding line must have been located upstream of the 1996 position. However, our data suggest that in 2008 the grounding line had not retreated to the edge of the subglacial
30 trough at ~6–9 km upstream, yet. The rapidity of the acceleration in 2008 indicates that resistance to glacier flow must have abruptly been reduced. This is characteristic of a response to sudden unpinning rather than to gradual grounding line retreat. We hence propose that in 2008 the frontal part of the glacier abruptly detached from a pinning point (likely a sill) located at the 1996 grounding line. A cavity underneath the ice has probably already existed. Between 2008 and early 2010 Fleming Glacier was possibly grounded and stabilized on a gentle hill ~2.5 – 4 km upstream of the 1996 grounding line. The second

phase of gradual acceleration and upstream propagation of high velocities between March 2010 and early 2011 is likely a response to further gradual grounding line retreat to the recent position.

Rignot et al. (2002) demonstrated that if ungrounding occurs, the flow acceleration usually affects both the floating and the grounded part of the glacier, but is largest near the grounding zone. This is consistent with our observation of a highest relative speedup by ~32–35 % between 7 and 11 km upstream, which is in the vicinity of our estimated recent grounding line location. Furthermore, 60–70% higher median ice thinning rates in the period between 2011 and 2014 in comparison to the period 2004–2008 point to an increased dynamic thinning and mass loss after in response to the grounding line retreat. The highest negative elevation change rates migrated upstream and can now be found in the vicinity of the estimated current grounding line. In the grounding zone basal melt adds to the dynamic thinning.

A tendency of lower ice thinning rates towards the glacier front, which was detected along the 2011–2014 OIB ATM profiles and in the 2011–2014 TSX/TDX dh/dt map, indicates floatation of the glacier tongue. This is in contrast to Walker and Gardner (2017), who do not see the positive trend of elevation change for the same OIB ATM dataset. Their approach of averaging elevation change in 5 km intervals probably filtered out the positive trend that is most prominent on the lowest 3 km of the profile.

Rignot et al. (2013) and Depoorter et al. (2013) reported high basal melt rates of $23.6 \pm 10 \text{ m a}^{-1}$ and $14.79 \pm 5.26 \text{ m a}^{-1}$ for the remaining parts of Wordie Ice Shelf, respectively. The magnitude of basal thinning is comparable to those found for ice shelves in the Amundsen Sea sector, where the influx of relatively warm CDW onto the continental shelf is thought to be the dominant driver for recent substantial grounding line retreat, acceleration and dynamic thinning of several glaciers (Turner et al., 2017). Periodical pulses of warm CDW are also known to flood onto the continental shelf of Marguerite Bay (Holland et al., 2010). Significant warming of Antarctic Continental Shelf Bottom Water (ASBW) of 0.1° to $0.3^\circ\text{C decade}^{-1}$ since the 1990s were recorded in the Bellingshausen Sea region and linked to increased warming and shoaling of CDW (Schmidtke et al., 2014). Cook et al. (2016) proposed that oceanic melt induced by an increased shoaling of relatively warm CDW is responsible for an accelerated frontal retreat of tidewater glaciers in the south-western Antarctic Peninsula since the 1990s.

Other studies reported considerable thinning of the nearby George VI Ice Shelf (Hogg et al., 2017; Holt et al., 2013) and other ice shelves on the south-western Antarctic Peninsula (e.g. Rignot et al., 2013) due to increased basal melt. The onset of Fleming Glacier's speedup between January–April/March 2008 corresponds well with observations of Wouters et al. (2015). They reported first signs of a near simultaneous increase of ice mass loss for glaciers all across the western Antarctic Peninsula south of -70° since 2008 and an unabated rapid dynamic ice loss since 2009. For the glaciers on Western Palmer Land Hogg et al. (2017) showed that ~35% of the ice loss after 2009 can be attributed to dynamic thinning triggered by ocean driven melt.

Walker and Gardner (2017) found that in 2008/2009 and 2010/2011 exceptional warm water intrusions into Wordie Bay occurred due to upwelling CDW in response to phases of anomalously strong north-westerly winds during strong La Niña and positive SAM (Southern Annular Mode) events. Highest temperatures were not only recorded at depths between 100–200 m but also below 400 m, where Fleming Glacier is grounded. The coincident timing with the two phases of glacier

acceleration substantiates the link between ocean warming and our observed dynamic changes. It is very likely that submarine ice melting was increased during phases of strong CDW upwelling and that this has triggered unpinning from the 1996 grounding line position in 2008 as well as further gradual grounding line retreat in 2010–2011.

The strong basal melt rates proposed by Rignot et al. (2013) and Depoorter et al. (2013) further suggest that basal melt probably has already occurred prior to 2008. This, together with increased dynamic thinning towards the ice front between 2004 and 2008 has likely weakened the ice at the pinning point, which may have fostered unpinning in 2008. Furthermore, the bed topography reveals that the trough underneath the joint Airy-Rotz-Seller-Fleming glacier tongue has a retrograde slope. Such a bed topography is known to be an unstable configuration for the glacier (e.g. DeConto and Pollard, 2016; Favier et al., 2014; Rignot et al., 2014; Schoof, 2012), which may have promoted gradual grounding line retreat between 2010 and 2011. Figure S8 summarizes our interpretation of grounding line retreat at Fleming Glacier.

The bedrock topography of Fleming Glacier also shows a retrograde bed slope starting at ~3–4 km upstream of the current grounding line, which transitions into a pronounced deep trough (up to 1100 m below sea level) at about 10 km upstream. Hence, ~~if grounding line retreat exceeds the edge of this trough, further thinning and subsequent ungrounding may have strong consequences and respective unstable conditions destabilisation~~ like on Thwaites Glacier and in the Pine Island Bay region ~~is possible, which would involve further rapid grounding line retreat and amplified mass loss in the future (Favier et al., 2014; Rignot et al., 2014).~~

~~.- In austral summer of 2007/2008 the lowest sea ice extent for the period 1978–2010 was recorded in the Bellingshausen/Amundsen Seas (Parkinson and Cavalieri, 2012). The bedrock topography reveals that the limits of hydrostatic equilibrium follow the edges of a subglacial trough underneath the joint Airy-Rotz-Seller-Fleming glacier tongue. We interpret this unfavourable bed topography to have promoted the grounding line retreat. Furthermore, before the speedup occurred, the highest ice thinning rates were recorded close to the grounding line position in 1996. This points to pronounced basal melt at the grounding line prior to the ungrounding. The bedrock topography of Fleming Glacier also shows a retrograde bed slope starting at ~3–4 km upstream of the current grounding line, which transitions into a pronounced deep trough (up to 1000 m below sea level) at about 10 km upstream. Hence, further thinning and subsequent ungrounding may have strong consequences and respective unstable conditions like on Thwaites Glacier and in the Pine Island Bay region. Wilkins Ice Shelf also showed various break-up events from beginning 2008 onwards after 10 years of absence of break-up events (Braun et al., 2009). Rankl et al. (2017) revealed a considerable speed-up of Wilkins Ice Shelf from 2007 to 2008. These tremendous changes were likely triggered by upwelling of relatively warm CDW which led to enhanced basal melt underneath of ice shelves as well as to a reduction of sea ice. Cook et al. (2016) proposed that oceanic melt induced by an increased shoaling of relatively warm CDW is the dominant driver for an accelerated frontal retreat of tidewater glaciers in the south-western Antarctic Peninsula since the 1990s. Other studies reported considerable thinning of the nearby George VI Ice Shelf (Hogg et al., 2017; Holt et al., 2013) and other ice shelves on the south-western Antarctic Peninsula (e.g. Rignot et~~

al., 2013) due to increased basal melt. All in all, our observations show similarities with those made in the Amundsen Sea Sector,

7 Conclusions

We presented a detailed history of the glacier dynamics of Fleming Glacier after the retreat and disintegration of Wordie Ice Shelf by analysing glacier extent, surface velocities, elevation change rates and hydrostatic equilibrium. Especially our dense SAR time series enabled us to precisely date events of velocity change jointly with glacier retreat, ~~and~~ elevation changes and grounding line retreat.

Our results show that until 2008 the dynamics of Fleming Glacier were primarily controlled by the impacts of break-up events of Wordie Ice Shelf before the early 1990s. The retreat of the ice shelf reduced glacier buttressing and led to an increase in surface velocities (Rignot et al., 2005), which in turn caused the glacier to dynamically thin. The last floating ice shelf parts were lost between 1998 and 1999, ~~but t-~~ This showed no detectable effects on glacier flow dynamics.

After two decades of rather stable velocities, the glacier abruptly accelerated between January and April 2008. Our interpretation is that this happened due to the detachment of the glacier tongue from a pinning point located at the 1996 grounding line position ~~an ungrounding of the glacier tongue over extended areas. The unpinning was likely fostered by weakening of the ice due to basal melt and dynamic thinning prior to 2008.~~ Further gradual retreat of the grounding line between 2010 and 2011, an increase in surface velocities of ~~~2930~~ % as well as ~~~60~~–70 % higher ice thinning rates show that ungrounding in 2008 has initiated a new phase of dynamic ~~imbalance~~ instability. The unfavourable ~~retrograde trough-shaped bed slope topography~~ underneath Fleming Glacier probably amplified the grounding line retreat.

The coincident timing of reported strong upwelling events of warm CDW with the two phases of acceleration and grounding line retreat shows ~~Our data suggest~~ that enhanced basal melt ~~at the grounding line~~ due to increased shoaling of warm CDW most likely played a major role for the recent changes at Fleming Glacier. The reduction in buttressing due to unpinning and grounding line retreat is able to explain why the magnitude of velocity change was much higher than in other places at the western Antarctic Peninsula during this time. The ungrounding in 2008 coincides in time with accelerations of ice shelf tributary glaciers further in the south west of the Antarctic Peninsula, break up events and ice shelf speedup at Wilkins Ice Shelf as well as records of very low sea ice extent in the Bellingshausen and Amundsen Sea regions.

Today Fleming Glacier and the other glaciers of the Airy-Rotz-Seller-Fleming glacier system are far away ~~from~~ reaching a new equilibrium. The modelled subglacial topography of Fleming Glacier upstream of the recent grounding line is characterized by some smaller landward deepening troughs which are separated by chains of gentle hills. ~~As p~~ Pronounced oceanic forcing will presumably ~~is likely to~~ continue, since the SAM is forecasted to be shifted further poleward, which will foster conditions like those during the strong La Niña/+SAM events in 2008/2009 and 2010/2011 (Abram et al., 2014; Fogt et al., 2011; Walker and Gardner, 2017). Thus, further retreat of the grounding line and more dynamic thinning are expected on Fleming Glacier. If the ungrounding would reach upstream to the retrograde bed slope at about ~3–4 km from the current

grounding line and further to the deep subglacial trough, this ~~can trigger a positive feedback loop of rapid grounding line retreat, flow acceleration, dynamic thinning, increased calving and mass loss~~ could have fatal effects on the stability and sea level contribution of Fleming Glacier. However, on Airy and Seller glaciers the more favourable subglacial geometry of a landward steepening slope may slow down or prevent further grounding line retreat in the near future

5 Author contributions

Peter Friedl processed and analysed all SAR, optical, elevation and ground penetrating radar data. Thorsten C. Seehaus processed the ALOS PALSAR data and supported the interferometric DEM generation. Anja Wendt provided the CAMS LiDAR data and gave valuable comments. Matthias H. Braun coordinated the research and wrote the manuscript jointly with Peter Friedl. Kathrin Höppner initiated the project, co-coordinated the research in close cooperation with Matthias H. Braun and contributed to discussions of the results throughout. All authors revised the manuscript.

Competing interests

The authors declare no competing financial interests.

Acknowledgements

This project was mainly funded by the DLR VO-R Young Investigator Group “Antarctic Research”. Thorsten C. Seehaus was funded by Deutsche Forschungsgemeinschaft (DFG) in the framework of the priority program "Antarctic Research with comparative investigations in Arctic ice areas" by a grant to M.B. (BR 2105/9-1). Matthias H. Braun and Thorsten C. Seehaus would like to thank the HGF Alliance “Remote Sensing of Earth System Dynamics” (HA-310) and Marie-Curie-Network International Research Staff Exchange Scheme IMCONet (EU FP7-PEOPLE-2012-IRSES, Project Reference: 318718) for additional support. TerraSAR-X and TanDEM-X data were provided by DLR through the projects HYD3008 and XTI_GLAC7015. We would like to thank Rainer Lorenz (German Remote Sensing Data Centre, DLR) for his support retrieving the ERS-1/-2 SLC data from the ESA ERS-1/-2 Processing and Archiving Facility (D-PAF) at DLR and the ENVISAT ASAR SLC data from the ESA ENVISAT Processing and Archiving Centre (D-PAC) at DLR. All ERS, ENVISAT and TSX/TDX data were received at the GARS (German Antarctic Receiving Station) O’Higgins of DLR. The CAMS data were acquired and made available by the Centro de Estudios Científicos, Chile. Furthermore, we are grateful to Christian Kienholz (University of Alaska Fairbanks) and Bill Hauer (Alaska Satellite Facility) for providing the Radarsat-1 SLC data. Other remote sensing data was kindly made available by various institutions, e.g. ESA, NASA, JAXA, USGS and CReSIS. Finally, we would like to thank Paul Wachter, and Saurabh Vijay ~~for helpful comments~~ and the three anonymous reviewers for their helpful comments on the manuscript.

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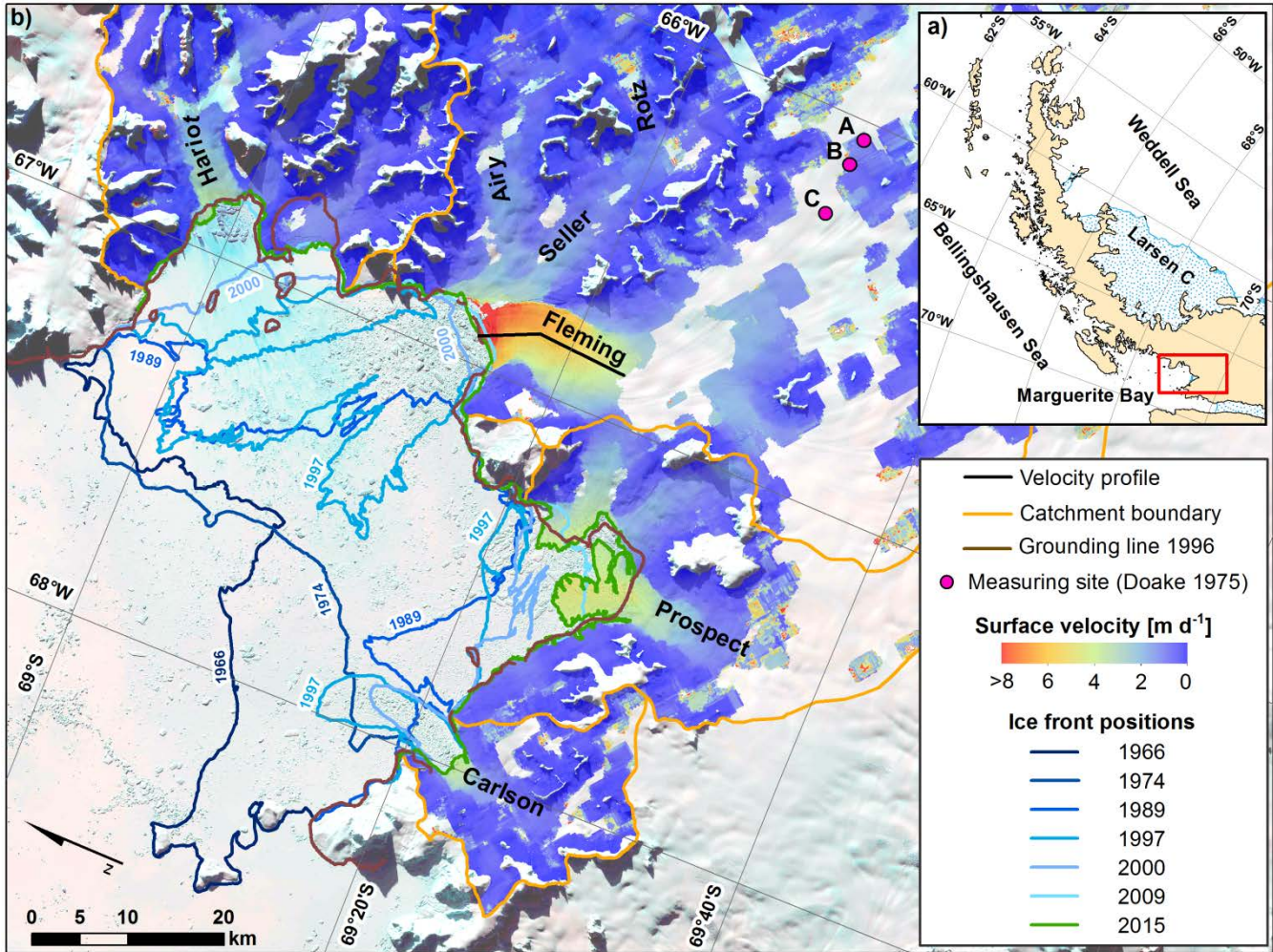


Figure 1: (a) Location of Wordie Bay at the Antarctic Peninsula. Map base: SCAR Antarctic Digital Database, version 6.0. (b) Surface velocity field and frontal positions of Wordie Ice Shelf between 1966 and 2015. Surface velocities were derived from Sentinel-1 acquisitions acquired on 28-08-2015 and 09-09-2015. Front positions (blue and green lines) (coloured lines) were taken from existing datasets or manually mapped from calibrated and multi-looked SAR intensity images. For detailed information on the data sources used for the frontal delineation see Supplemental material Tab. S1. The grounding line in 1996 (brown line) was derived from ERS-1/2 double difference interferometry (Rignot et al., 2005; Rignot et al., 2011a). Black line: Extraction profile for the velocity time series presented in Sect. 5. Orange Line: Glacier system catchment boundaries from the SCAR Antarctic Digital Database, version 6.0. Pink dots: Sites of the velocity measurements undertaken by Doake, 1975 in 1974. Background: Mosaic of two Landsat-8 „Natural Colour“ images, acquired on 2015-09-16 ©USGS.

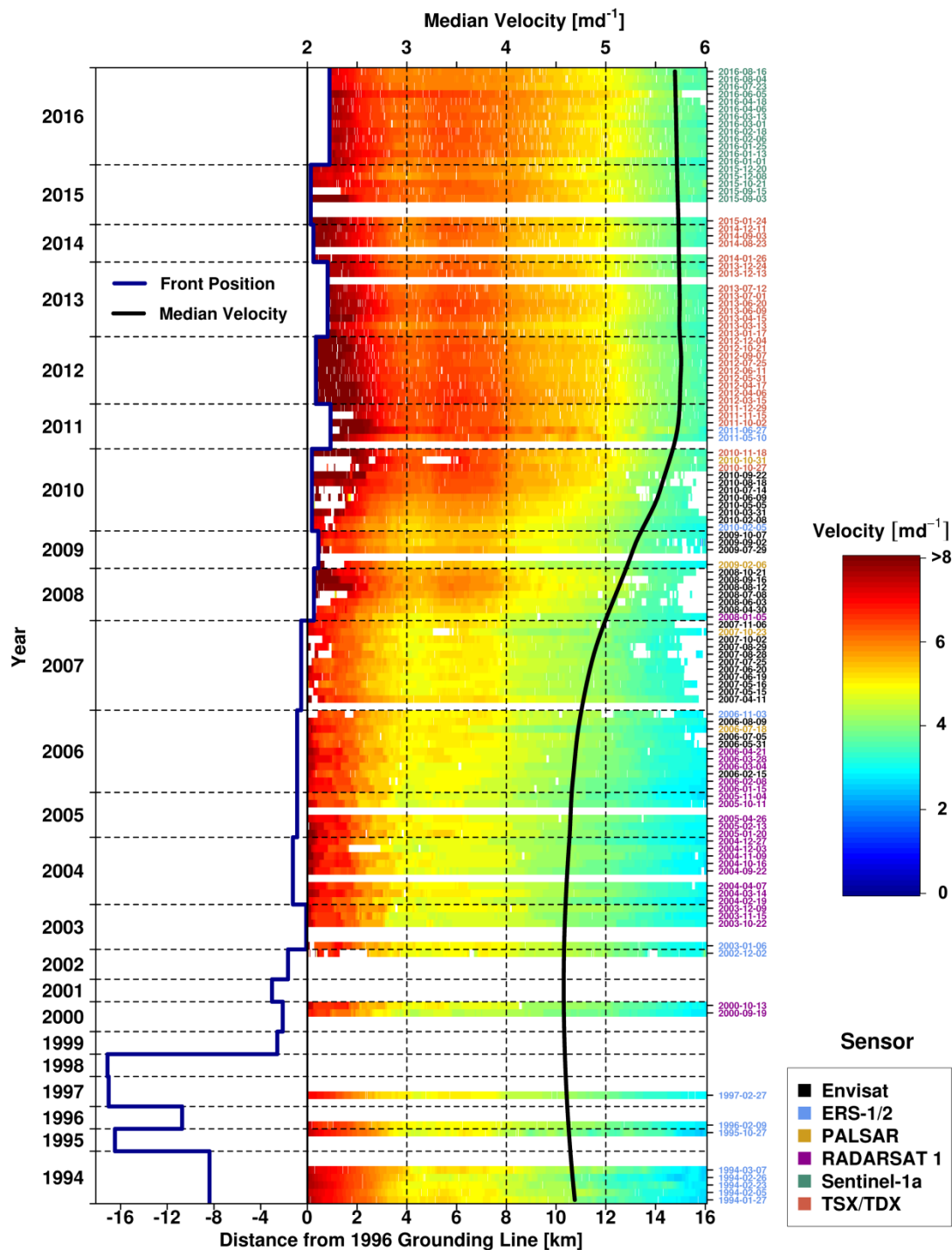


Figure 2: Left side (blue line): relative distance of the glacier front to the grounding line position in 1996. See Tab. S1 for data used for front mapping. -Right side: velocity time series and smoothed median velocities (black line) derived from multi sensor SAR intensity tracking along a centreline profile on Fleming Glacier starting at the 1996 grounding line (Fig. 1).

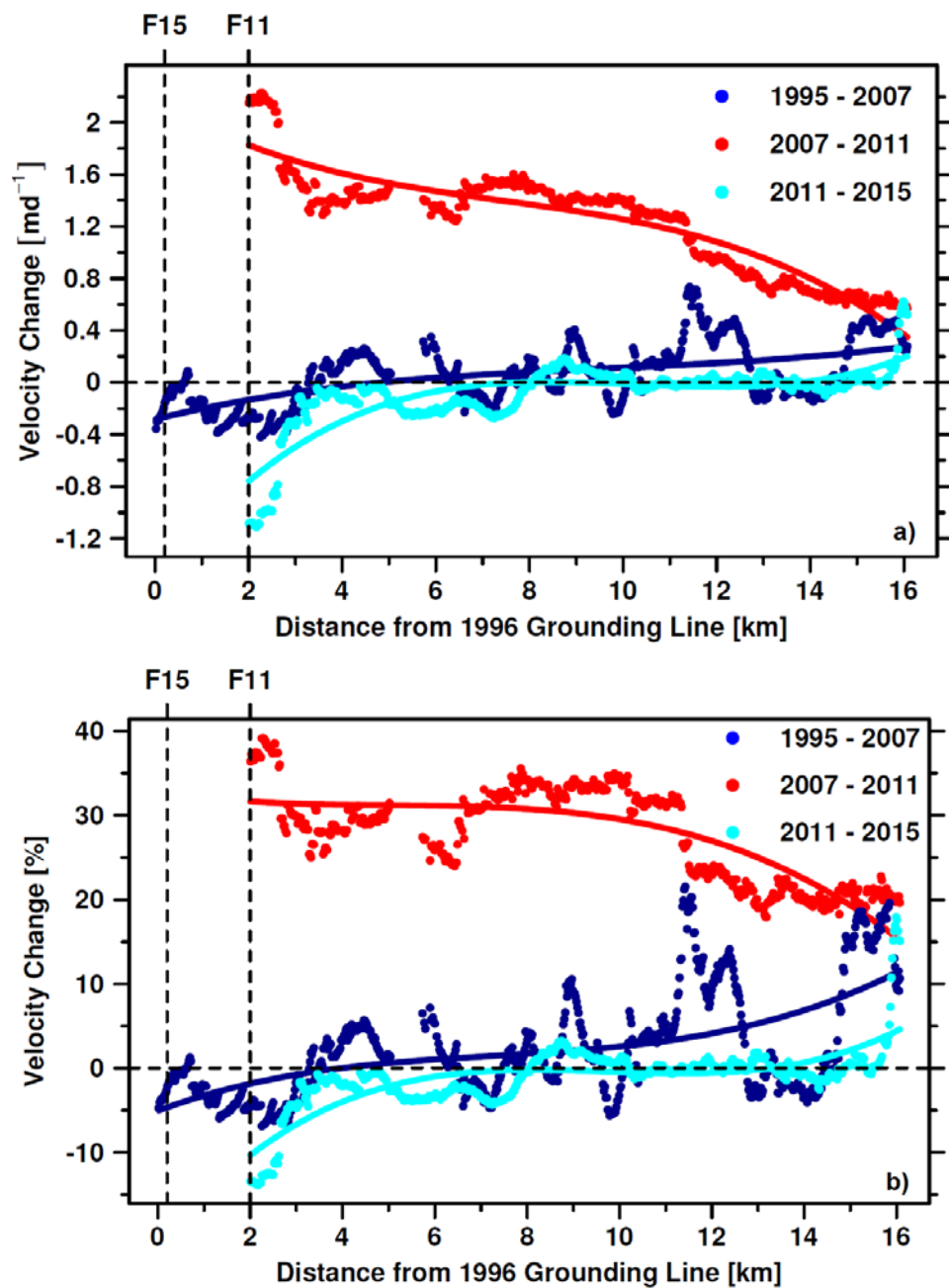


Figure 3: Absolute (a) and relative (b) velocity change along the centreline profile on Fleming Glacier (Fig. 1) for 1995-10-27 to 2007-10-23, 2007-10-23 to 2011-10-02 and 2011-10-02 to 2015-10-21. Coloured lines show cubic functions fitted to the data. F11: front position in 2011. F15: Front position in 2015. See Tab. S1 for data used for front mapping.

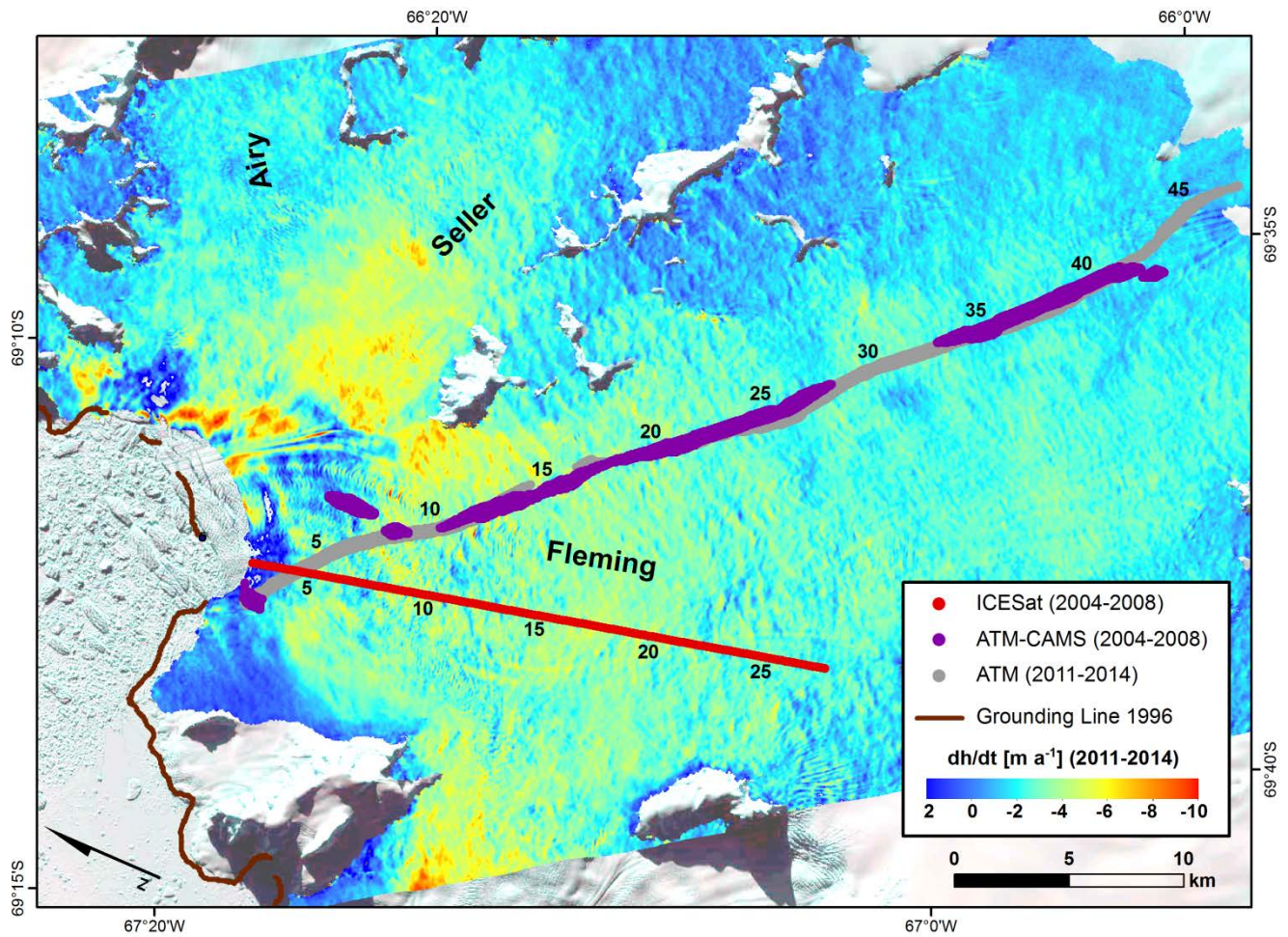
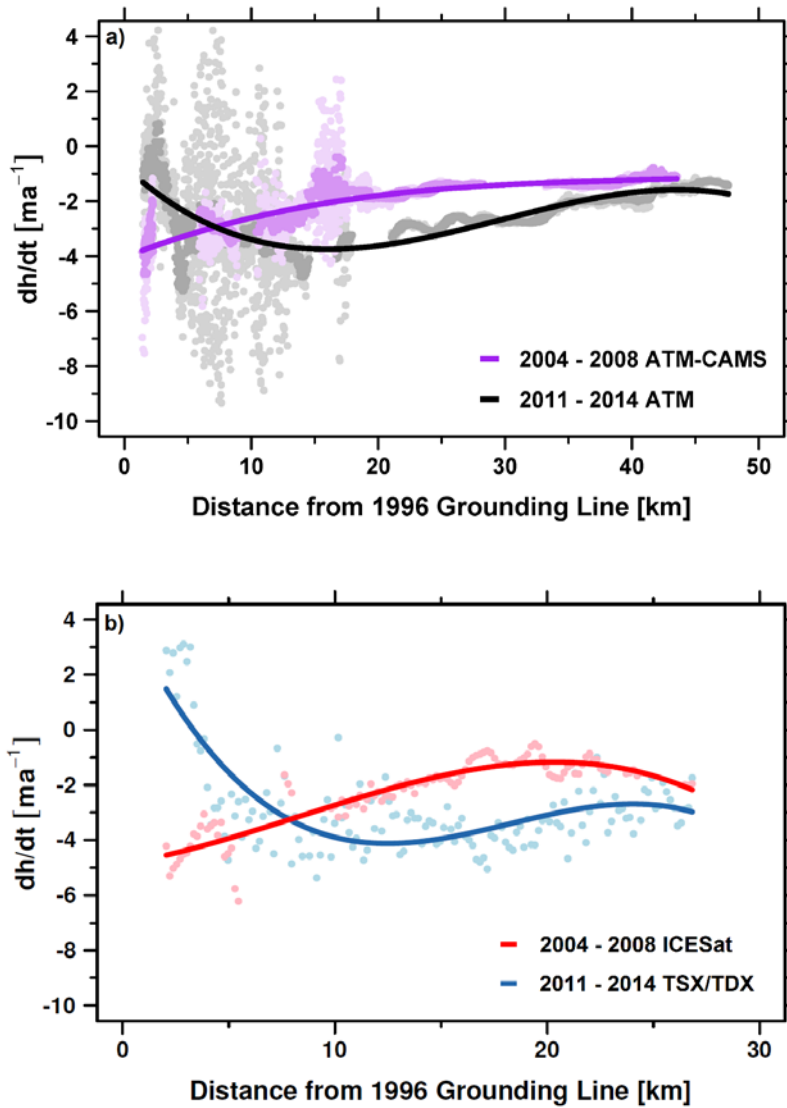


Figure 4: Glacier surface elevation change on Fleming Glacier between 2011 and 2014 derived from TSX/TDX bistatic and monostatic acquisitions. Red dots: ICESat track on 2008-10-04 taken as reference for ICESat 2004-2008 dh/dt calculations. Purple dots: common locations of PIB ATM and CAMS LiDAR measurements in 2004 and 2008. Grey dots: common locations of OIB ATM LiDAR measurements in 2011 and 2014. -Brown line: -grounding line in 1996 from Rignot et al. (2005) and Rignot et al. (2011a). Numbers indicate distances to the 1996 grounding line. Background: Mosaic of two Landsat-8 „Natural Color” images, acquired on 2015-09-16 ©USGS.



5 **Figure 5: (a)** Elevation change rates on Fleming Glacier 2004–2008 and 2011–2014 plotted against distance from the 1996 grounding line. Light purple dots: change rates from PIB ATM and CAMS LiDAR measurements in 2004 and 2008. Purple dots: median filtered elevation change rates 2004–2008. Light grey dots: change rates from OIB ATM LiDAR measurements in 2011 and 2014. Dark grey dots: median filtered elevation change rates 2004–2008. See Fig. 4 for flight path locations. Purple and black lines: Cubic functions fitted to the median filtered elevation change rates both datasets. **(b)** Elevation change rates on Fleming Glacier 2004–2008 and 2011–2014 plotted against distance from the 1996 grounding line. Light red dots: change rates from ICESat measurements in 2004 and 2008. Light blue dots: change rates between 2011 and 2014, extracted from the TSX/TDX dh/dt map along the 2008 ICESat track (see Fig. 4 for location). Red and blue lines: Cubic functions fitted to both datasets.

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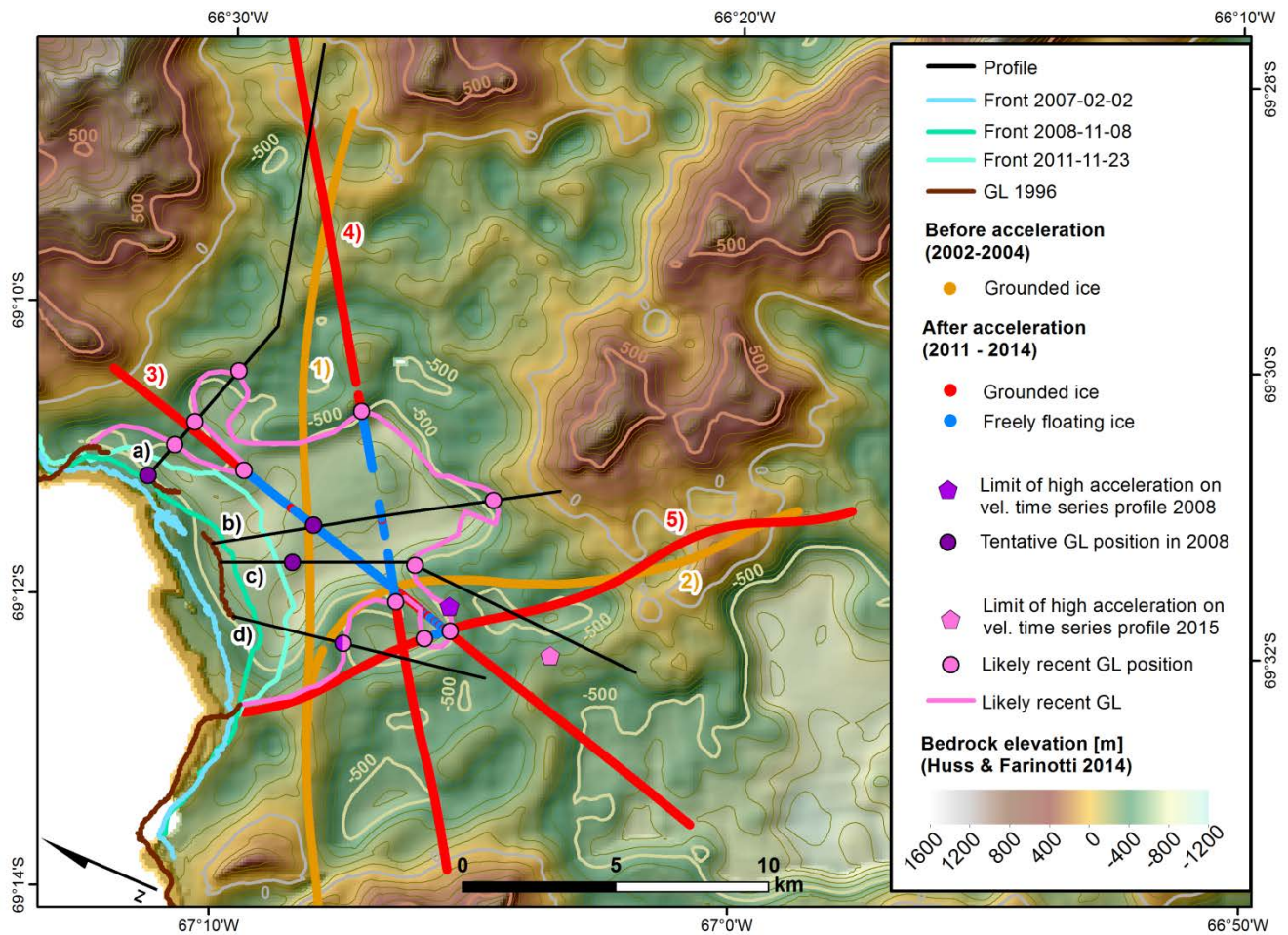


Figure 6: Floating area of Fleming Glacier and estimation of the recent grounding line. Black lines 1–4: Profiles for extraction of modeled bedrock elevations, surface velocities and elevation change rates (see Fig. S6 a1–d4). Elevation contours are shown at an interval of 100 m. Light blue and cyan lines: glacier fronts on 2007-02-02 and 2011-11-23. Brown line: grounding line in 1996 (Rignot et al., 2005; Rignot et al., 2011a). Orange dots: Grounded ice before acceleration as derived from PIB LiDAR and ice thickness data. Orange dots: Fulfillment of floating condition before acceleration in 2008, derived from PIB LiDAR and ice thickness data. Dates of PIB flights: 1a) 2002-11-26, 2b) 2004-11-18. Background: bedrock elevation from Huss and Farinotti (2014). Blue and red dots: Freely floating and grounded ice after acceleration as derived from OIB laser altimeter and ice thickness data. Blue and red dots: Fulfillment of the hydrostatic equilibrium assumption after the acceleration derived from OIB laser altimeter and ice thickness data. Dates of OIB flights: 3e) 2011-11-17, 4d) 2014-11-16, 5e) 2014-11-10. Purple pentagon: Location of the limit of high acceleration on the velocity time series profile in 2008 (Fig. 1 and 2). Purple circle: likely grounding line location in 2008. Purple circles with cross: possible grounding line locations in 2008. Pink pentagon: Location of the limit of acceleration on the velocity time series profile in 2015 (Fig. 1 and 2). Pink circles: Estimated positions of the recent grounding line obtained from buoyancy calculations, surface velocities, elevation change rate patterns and/or modeled bedrock elevations on profiles a–d. Pink line: Final solution of the likely recent grounding line location.

Tables

Table 1: Sensors and data used in this study, and their main specifications. Intensity tracking parameters are provided in pixels [p] in slant range geometry.

SAR sensors

Platform	Sensor	Mode	SAR band	Repetition cycle [d]	Time interval [yyyy-mm-dd] = [yyyy-mm-dd]	Tracking patch sizes [p x p]	Tracking step size [p x p]
ERS-1/2	AMI SAR	IM	C band	35/3/1	1994-01-26 = 2011-06-29	64x320	5x25
RADARSAT 1	SAR	ST	C band	24	2000-09-07 = 2008-01-17	128x512	5x20
Envisat	ASAR	IM	C band	35	2006-02-15 = 2010-10-10	64x320	5x25
ALOS	PALSAR	FBS	L band	46	2006-06-25 = 2010-11-23	128x384	10x30
TerraSAR-X/ TanDEM-X	SAR	SM	X band	11	2008-10-14 = 2015-01-30	512x512	25x25
Sentinel-1a	SAR	IW	C band	12	2015-08-28 = 2016-08-22	640x128	50x10

LiDAR/Laser Altimeter

Mission	Sensor	Type	Wavelength [nm]	Footprint [m]	Dates [yyyy-mm-dd]	Estimated accuracy	Reference
Pre-IceBridge (PIB)	ATM	LIDAR	532	1	2002-11-26 2004-11-18	0.20 m	Krabill (2012)
ICESat	GLAS	Laser Altimeter	1064	70	2004-05-18 2008-10-04	0.20 m	Zwally et al. (2014)
CECS/FACH	CAMS	LiDAR	900	1	2008-12-07 2011-11-17	0.25 m	Wendt et al. (2010)
Operation IceBridge (OIB)	ATM	LiDAR	532	1	2014-11-10 2014-11-16	0.20 m	Krabill (2010, updated 2016)

Ice Thickness

Mission	Sensor	Type	Bandwidth [MHz]	Sample spacing [m]	Dates [yyyy-mm-dd]
Pre-IceBridge (PIB)	ICoRDS-2	Radar	141.5-158.5	~130	2002-11-26
Pre-IceBridge (PIB)	ACoRDS	Radar	140-160	~30	2004-11-18
Operation IceBridge (OIB)	MCoRDS	Radar	180-210	~15	2011-11-17
Operation IceBridge	MCoRDS 2	Radar	165-215	~15	2014-11-10

Table 2: Comparison of surface velocities and flow directions at Fleming Glacier obtained by an optical survey in 1974 (Doake, 1975), SAR interferometry in 1996 (Rignot et al., 2005),GPS measurements in 2008 (Wendt et al., 2010) and SAR intensity tracking in 2013. Velocities and flow directions in 2013 were derived from intensity tracking applied on two TSX/TDX acquisitions on 2013-12-19 and 2013-12-30. Velocities in 2015 are from (Zhao et al., 2017). For the locations of the measuring sites see Fig. 1.

Location	1974		1996		2008		2013		<u>2015</u>	
	Magnitude	Direction	Magnitude	Direction	Magnitude	Direction	Magnitude	Direction	<u>Magnitude</u>	<u>Direction</u>
	m a ⁻¹	°	m a ⁻¹	°	m a ⁻¹	°	m a ⁻¹	°	<u>m a⁻¹</u>	<u>°</u>
A (69.505° S, 66.049° W)	146 ± 4	277 ± 5	244 ± 10	285	205.5 ± 0.02	275.8 ± 0.1	205 ± 22	272	<u>271 ± 20</u>	<u>NA</u>
B (69.502° S, 66.123° W)	175 ± 4	272 ± 5	271 ± 10	287	<u>NA</u>	<u>NA</u>	244 ± 22	271	<u>299 ± 20</u>	<u>NA</u>
C (69.500° S, 66.267° W)	201 ± 4	283 ± 5	306 ± 10	300	312.8 ± 0.04	286.3 ± 0.1	323 ± 22	284	<u>356 ± 20</u>	<u>NA</u>

Supplement

S1 Data used for front line delineation

Table S1: Data used for mapping the front positions in Fig. 1, 2, 3, ~~and 6~~ and S6.

Year	Platform	Reference	Date
1966	Aerial photography	Ferrigno, 2008	11/12 1966
1974	Landsat-1	Ferrigno, 2008	1974-01-06
1989	Landsat-3	Ferrigno, 2008	1989-02-20
1994	ERS-1/2		1994-02-01
1995	ERS-1/2		1995-10-27
1996	ERS-1/2		1996-02-10
1997	ERS-1/2		1997-01-29
1998	ERS-1/2		1998-01-30
1999	ERS-1/2		1999-11-10
2000	ERS-1/2		2000-02-20
2001	Landsat-7		2001-01-04
2002	ERS-1/2		2002-02-23
2003	ERS-1/2		2003-01-24
2004	ERS-1/2		2004-03-03
2005	ERS-1/2		2005-01-28
2007	ERS-1/2		2007-02-02
2008*	Envisat	Wendt et al., 2010	2008-04-13
<u>2008*</u>	<u>Envisat</u>		<u>2008-11-08</u>
2009	ASTER	Wendt et al., 2010	2009-02-02
2010	ERS-1/2		2010-02-26
2011	TSX/TDX		2011-11-23
2012	TSX/TDX		2012-10-16
2013	TSX/TDX		2013-12-08
2014	TSX/TDX		2014-11-03
2015	Sentinel 1a		2015-09-09
2016	Sentinel 1a		2016-01-31

*The 2008-04-13 front position is shown in Fig. 1 and the 2008-11-08 front position is shown in Fig. 6 and S6.

S2 Error estimation of surface velocity measurements

The corresponding errors of the velocity measurements were estimated as described in detail in Seehaus et al. (2015). It is assumed that the resulting uncertainties for each velocity field are induced by two major sources: the coregistration process and the tracking algorithm itself. The error caused by residual inaccuracies of the coregistration (σ_V^C) was determined by calculating the median velocity for 19 to 64 points on stable non-moving surfaces (e.g. rock outcrops) (Fig. S1). The error induced by the tracking algorithm (σ_V^T) was estimated according to the following formula, modified from McNabb et al. (2012):

$$\sigma_V^T = \frac{C\Delta x}{z\Delta t} \quad (1)$$

where C is the uncertainty of the tracking algorithm (assumed to be 0.4 pixels), Δx is the image resolution (mean values for each sensor are listed in Tab. S2), z is the oversampling factor (we applied a factor of two) and Δt is the temporal baseline between the SAR images. The total error (σ_V) of the velocity measurement is the sum of σ_V^C and σ_V^T . Table S3 lists the values Δt , σ_V^C , σ_V^T and σ_V for each velocity field. As in Seehaus et al. (2015) the quite large σ_V^T values for ERS-1/2 measurements during one of the sensor's "Tandem" or "Ice Phases", where the satellites orbited in 1- or 3-day repeat passes, were excluded from our estimation of σ_V .

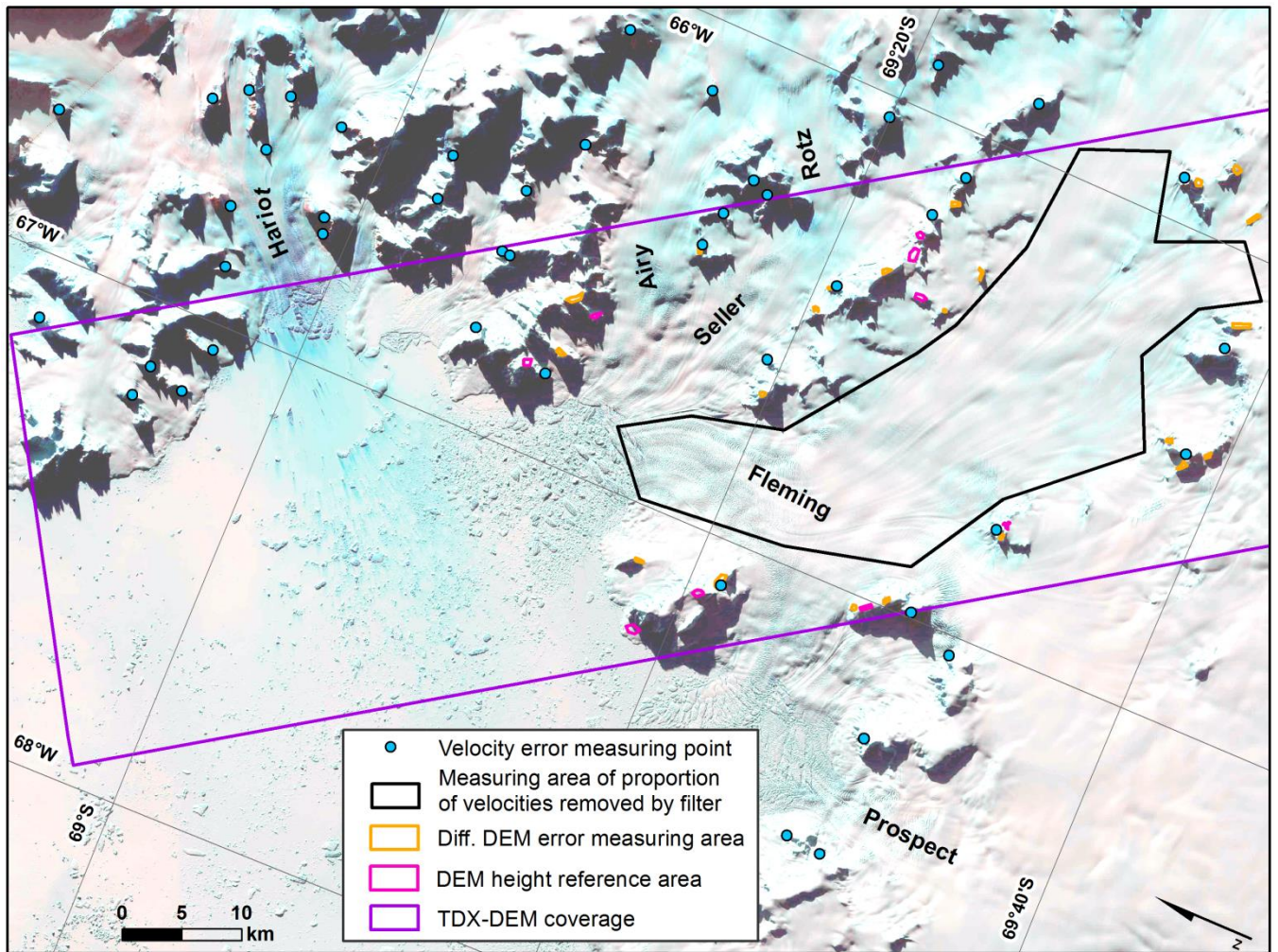


Figure S1: Blue dots: Stable points used for the coregistration error estimation of the intensity tracking results; Black polygon: -reference area for the calculation of the proportion of velocity measurements removed by the filter of Burgess et al. (2012) (values of each velocity field are listed in Tab. S.3). Orange polygons: stable areas on nunataks and hills for accuracy assessment of elevation change measurements. Pink polygons: stable areas on nunataks and hills for vertical height referencing of the TanDEM-X DEMs. Purple polygon: and spatial coverage of the TanDEM-X DEMs. Background: Mosaic of two Landsat-8 „Natural Color“ images, acquired on September 16, 2015 ©USGS.

Table S2: Ground range resolution Δx_{GR} , azimuth resolution Δx_{AZ} , and image resolution Δx used in Formula 1 (S2) for calculating velocity errors. For most of the sensors Δx_{GR} is coarser than Δx_{AZ} , except for Sentinel-1a that has a coarser resolution in azimuth direction and TSX/TDX which have fairly equal resolutions in both directions. In order to make a conservative estimate of velocity errors, always the coarser resolution value was chosen to be Δx . Mean values of image resolution (Δx) at the study site used for the calculation of the velocity error induced by the tracking algorithm for each sensor. The tracking was done in slant range geometry without applying multilooking. For all sensors except for Sentinel 1a the ground resolution in range direction was coarser than in azimuth direction. Consequently a calculated mean value of the azimuth resolution was taken as upper bound approximation of Δx for Sentinel 1a tracking results and a mean value of the ground range resolution for the results obtained from data of the other sensors.

Sensor	$\Delta x_{GR} [m]$	$\Delta x_{AZ} [m]$	$\Delta x [m]$
AMI SAR (ERS-1/2)	<u>20</u>	<u>4</u>	<u>20</u>
R1 (Radarsat 1)	<u>20</u>	<u>4</u>	<u>20</u>
ASAR (Envisat)	<u>17</u>	<u>4</u>	<u>17</u>
PALSAR (ALOS)	<u>7</u>	<u>3</u>	<u>7</u>
TSX/TDX(TerraSAR-X/TanDEM-X)	<u>2</u>	<u>2</u>	<u>2</u>
S1 (Sentinel-1a)	<u>3</u>	<u>14</u>	<u>14</u>

Table S3: Uncertainty σ_V of processed velocity fields [and proportion of removed velocity measurements by filter \(reference area is shown in Fig. S.1\)](#). Date: Mean date of SAR acquisitions; Δt : Time interval in days between repeat SAR acquisitions; σ_V^C : Uncertainty of image coregistration; σ_V^T : Uncertainty of intensity tracking processing; ERS velocity fields with $\Delta t \leq 3d$: $\sigma_V = \sigma_V^C$. The table is continued on the next pages.

Date [yyyy-mm-dd]	Sensor	Δt [d]	σ_V^C [m d ⁻¹]	σ_V^T [m d ⁻¹]	σ_V [m d ⁻¹]	Removed by filter [%]
1994-01-27	AMI SAR	3	0.13	1.33	0.13	1.47
1994-02-05	AMI SAR	3	0.19	1.33	0.19	1.01
1994-02-23	AMI SAR	3	0.14	1.33	0.14	3.85
1994-02-26	AMI SAR	3	0.34	1.33	0.34	4.41
1994-03-07	AMI SAR	3	0.2	1.33	0.2	1.93
1995-10-27	AMI SAR	1	0.25	4	0.25	1.34
1996-02-09	AMI SAR	35	0.12	0.11	0.23	3.88
1997-02-27	AMI SAR	35	0.12	0.11	0.23	4.91
2000-09-19	R1	24	0.11	0.14	0.25	7.1
2000-10-13	R1	24	0.11	0.14	0.25	4.58
2002-12-02	AMI SAR	35	0.19	0.11	0.3	1.64
2003-01-06	AMI SAR	35	0.13	0.11	0.24	1.83
2003-10-22	R1	24	0.12	0.14	0.26	2.98
2003-11-15	R1	24	0.19	0.14	0.33	4.21
2003-12-09	R1	24	0.09	0.14	0.23	4.38
2004-02-19	R1	24	0.08	0.14	0.22	2.63
2004-03-14	R1	24	0.12	0.14	0.26	3.19
2004-04-07	R1	24	0.17	0.14	0.31	3.36
2004-09-22	R1	24	0.1	0.14	0.24	3.61
2004-10-16	R1	24	0.21	0.14	0.35	3.91
2004-11-09	R1	24	0.09	0.14	0.23	2.63
2004-12-03	R1	24	0.14	0.14	0.28	3.49
2004-12-27	R1	24	0.15	0.14	0.29	13.79
2005-01-20	R1	24	0.26	0.14	0.4	11.63
2005-02-13	R1	24	0.07	0.14	0.21	14.86
2005-04-26	R1	24	0.19	0.14	0.33	2.59
2005-10-11	R1	24	0.13	0.14	0.27	18.38
2005-11-04	R1	24	0.17	0.14	0.31	1.75
2006-01-15	R1	24	0.12	0.14	0.26	10.81
2006-02-08	R1	24	0.16	0.14	0.3	1.7
2006-02-15	ASAR	35	0.15	0.11	0.26	3.29
2006-03-04	R1	24	0.06	0.14	0.2	1.23

Date [yyyy-mm-dd]	Sensor	Δt [d]	σ_V^C [m d ⁻¹]	σ_V^T [m d ⁻¹]	σ_V [m d ⁻¹]	Removed by filter [%]
2006-03-28	R1	24	0.23	0.14	0.37	1.27
2006-04-21	R1	24	0.17	0.14	0.31	1.25
2006-05-31	ASAR	35	0.09	0.11	0.2	3.35
2006-07-05	ASAR	35	0.11	0.11	0.22	3.2
2006-07-18	PALSAR	46	0.06	0.03	0.09	27.76
2006-08-09	ASAR	35	0.14	0.11	0.25	3.82
2006-11-03	AMI SAR	35	0.18	0.11	0.29	3.1
2007-04-10	ASAR	35	0.17	0.11	0.28	11.26
2007-05-15	ASAR	35	0.06	0.11	0.17	1.29
2007-05-16	ASAR	35	0.09	0.11	0.2	3.44
2007-06-19	ASAR	35	0.05	0.11	0.16	3.44
2007-06-20	ASAR	35	0.08	0.11	0.19	1.33
2007-07-25	ASAR	35	0.05	0.11	0.16	3.92
2007-08-28	ASAR	35	0.1	0.11	0.21	1.41
2007-08-29	ASAR	35	0.13	0.11	0.24	4.08
2007-10-02	ASAR	35	0.08	0.11	0.19	1.45
2007-10-03	ASAR	35	0.11	0.11	0.22	4.25
2007-10-23	PALSAR	46	0.05	0.03	0.08	20.19
2007-11-06	ASAR	35	0.06	0.11	0.17	1.63
2008-01-05	R1	24	0.2	0.14	0.34	1.95
2008-04-30	ASAR	35	0.11	0.11	0.22	5.52
2008-06-03	ASAR	35	0.08	0.11	0.19	1.97
2008-07-08	ASAR	35	0.06	0.11	0.17	2.05
2008-08-12	ASAR	35	0.08	0.11	0.19	2.37
2008-09-16	ASAR	35	0.08	0.11	0.19	2.22
2008-10-21	ASAR	35	0.07	0.11	0.18	2.45
2009-02-06	PALSAR	46	0.15	0.03	0.18	26.83
2009-04-14	ASAR	35	0.22	0.11	0.33	6.95
2009-07-29	ASAR	35	0.09	0.11	0.2	6.57
2009-09-02	ASAR	35	0.09	0.11	0.2	6.44
2009-10-07	ASAR	35	0.09	0.11	0.2	6.63
2010-02-08	ASAR	35	0.08	0.11	0.19	4.69
2010-02-11	ASAR	35	0.19	0.11	0.3	0.02
2010-03-31	ASAR	35	0.09	0.11	0.2	5.16
2010-05-05	ASAR	35	0.1	0.11	0.21	5.49

Date [yyyy-mm-dd]	Sensor	Δt [d]	σ_V^C [m d ⁻¹]	σ_V^T [m d ⁻¹]	σ_V [m d ⁻¹]	Removed by filter [%]
2010-06-09	ASAR	35	0.08	0.11	0.19	5.29
2010-07-14	ASAR	35	0.09	0.11	0.2	5.39
2010-08-18	ASAR	35	0.1	0.11	0.21	5.87
2010-09-22	ASAR	35	0.09	0.11	0.2	6.04
2010-10-27	TSX/TDX	33	0.01	0.01	0.02	3.7
2010-10-31	PALSAR	46	0.33	0.03	0.36	28.43
2010-11-18	TSX/TDX	11	0.02	0.04	0.06	4.06
2011-05-10	AMI SAR	3	0.25	1.33	0.25	2.52
2011-06-27	AMI SAR	3	0.36	1.33	0.36	6.46
2011-10-02	TSX/TDX	11	0.02	0.04	0.06	2.94
2011-11-15	TSX/TDX	11	0.03	0.04	0.07	5.02
2011-12-29	TSX/TDX	11	0.01	0.04	0.05	6.47
2012-03-15	TSX/TDX	11	0.02	0.04	0.06	3.55
2012-04-06	TSX/TDX	11	0.02	0.04	0.06	4.56
2012-04-17	TSX/TDX	11	0.02	0.04	0.06	3.66
2012-05-31	TSX/TDX	11	0.03	0.04	0.07	4.14
2012-06-11	TSX/TDX	11	0.01	0.04	0.05	4.01
2012-07-25	TSX/TDX	11	0.02	0.04	0.06	1.69
2012-09-07	TSX/TDX	11	0.02	0.04	0.06	4.53
2012-10-21	TSX/TDX	11	0.02	0.04	0.06	4.34
2012-12-04	TSX/TDX	11	0.02	0.04	0.06	3.37
2013-01-17	TSX/TDX	11	0.01	0.04	0.05	3.94
2013-03-13	TSX/TDX	11	0.03	0.04	0.07	3.8
2013-03-23	TSX/TDX	11	0.08	0.04	0.12	2.19
2013-04-15	TSX/TDX	11	0.03	0.04	0.07	3.47
2013-06-09	TSX/TDX	11	0.02	0.04	0.06	2.56
2013-06-20	TSX/TDX	11	0.02	0.04	0.06	3.94
2013-07-01	TSX/TDX	11	0.02	0.04	0.06	2.37
2013-07-12	TSX/TDX	11	0.02	0.04	0.06	2.2
2013-11-20	TSX/TDX	11	0.04	0.04	0.08	0.71
2013-12-13	TSX/TDX	11	0.03	0.04	0.07	4.25
2013-12-24	TSX/TDX	11	0.02	0.04	0.06	3.57
2014-01-26	TSX/TDX	11	0.01	0.04	0.05	3.2
2014-08-23	TSX/TDX	11	0.02	0.04	0.06	4.67
2014-09-03	TSX/TDX	11	0.03	0.04	0.07	4.52

Date [yyyy-mm-dd]	Sensor	Δt [d]	σ_V^C [m d ⁻¹]	σ_V^T [m d ⁻¹]	σ_V [m d ⁻¹]	Removed by <u>filter [%]</u>
2014-12-11	TSX/TDX	11	0.01	0.04	0.05	<u>4.05</u>
2015-01-24	TSX/TDX	11	0.03	0.04	0.07	<u>2.2</u>
2015-09-03	S1	12	0.06	0.23	0.29	<u>2.47</u>
2015-09-15	S1	12	0.05	0.23	0.28	<u>2.51</u>
2015-10-21	S1	12	0.15	0.23	0.38	<u>2.51</u>
2015-12-08	S1	12	0.11	0.23	0.34	<u>2.03</u>
2015-12-20	S1	12	0.07	0.23	0.3	<u>2.13</u>
2016-01-01	S1	12	0.09	0.23	0.32	<u>2.21</u>
2016-01-13	S1	12	0.05	0.23	0.28	<u>2.08</u>
2016-01-25	S1	12	0.08	0.23	0.31	<u>2.18</u>
2016-02-06	S1	12	0.08	0.23	0.31	<u>2.15</u>
2016-02-18	S1	12	0.08	0.23	0.31	<u>2.1</u>
2016-03-01	S1	12	0.09	0.23	0.32	<u>2.15</u>
2016-03-13	S1	12	0.08	0.23	0.31	<u>2.11</u>
2016-04-06	S1	12	0.09	0.23	0.32	<u>2.01</u>
2016-04-18	S1	12	0.07	0.23	0.3	<u>2.12</u>
2016-06-05	S1	12	0.12	0.23	0.35	<u>1.97</u>
2016-07-23	S1	12	0.1	0.23	0.33	<u>2.09</u>
2016-08-04	S1	12	0.08	0.23	0.31	<u>2.07</u>
2016-08-16	S1	12	0.06	0.23	0.29	<u>2.17</u>

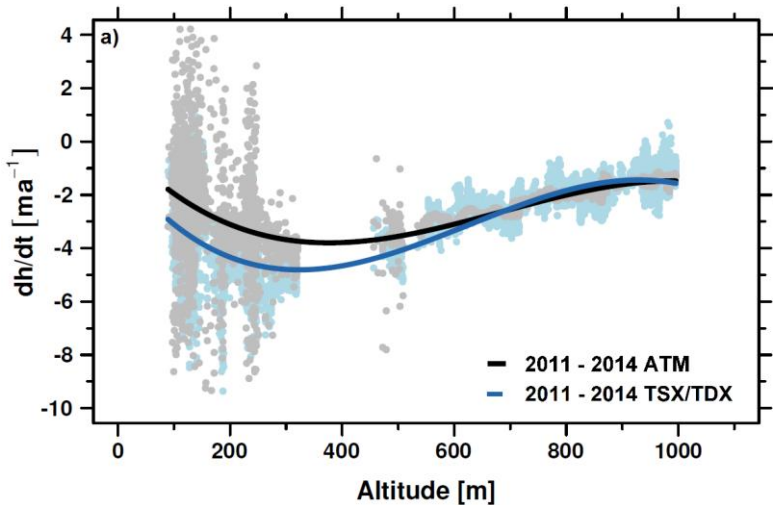
S3 Uncertainty estimation of variables for the hydrostatic height anomaly calculations and assessment of error propagation

In order to assess the propagation of uncertainties for the calculation of hydrostatic height anomalies, we estimated the error of each variable of Formula 2. The accuracy of the ATM elevations was estimated to be ± 0.2 m. The overall uncertainty of the EIGEN-6C4 geoid is 0.24 m (<http://icgem.gfz-potsdam.de/ICGEM/>). Accounting for an unknown additional error induced by kriging of the geoid values, we assumed a total accuracy of ± 0.5 m for e . Following the recommendation of the CReSIS Radar Depth Sounders (RDS) user guide (ftp://data.cresis.ku.edu/data/rds/rds_readme.pdf) we defined the error of H_i as the sum of the RMS error of the sensor's range resolution and the RMS error of the dielectric. Depending on the sensor, the uncertainty of H_i varies between ~ 6 and ~ 14 m for an ice thickness of 750 m, which is the approx. mean ice thickness in the vicinity of the current grounding line measured on our OIB profiles. For ρ_i we used a value of 917 kg m^{-3} , which is the standard density of pure ice (Benn and Evans, 2013). However, since impurities in the ice can cause this value to vary by around $\pm 5 \text{ kg m}^{-3}$ (Griggs and Bamber, 2011), we chose this rate to be the uncertainty of ρ_i . The global mean density of sea water is 1027 kg m^{-3} , but this value can vary locally. According to Griggs and Bamber (2011) we therefore assumed an error of $\pm 5 \text{ kg m}^{-3}$ for ρ_w . The firn density correction factor for pure glacier ice is 0. In situ values of about 10 m have been measured for δ on Larsen C Ice Shelf (Griggs and Bamber, 2009) and firn density correction factors > 20 m have been reported for areas of convergent flow on the Ross Ice Shelf (Bamber and Bentley, 1994). Since δ can vary spatially and we had no information on firn densities on the Airy-Rotz-Seller-Fleming glacier system, we used a mean firn density correction factor of 10 m throughout our calculations and noted an uncertainty of ± 10 m. In order to quantify the total error of Δe and to consider the propagation of uncertainties, we run a Monte Carlo simulation based on Formula 1 and 2 with 100.000 runs. For all possible sensor-depending errors of $H_i=750$ m the Monte Carlo simulation yielded a standard deviation of < 12 m for Δe . Thus we assumed the total ~~uncertainty error~~ of Δe to be $\pm \sim 12$ m. Consequently, we assigned locations on our OIB and PIB profiles to be freely floating ice, if the calculated values of Δe lay within this range. In the vicinity of the grounding zone the TDX global DEM has a minimum slope of 1° , and therefore we estimate the uncertainty in the horizontal position of the transition from grounded to freely floating ice to be ~ 700 m.

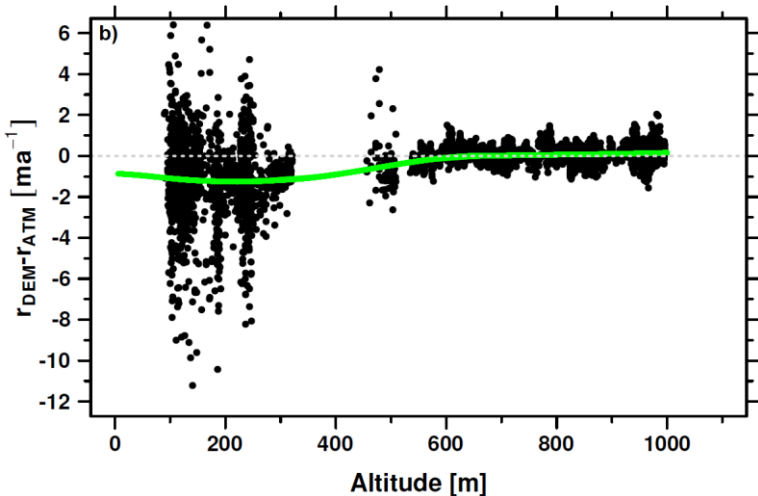
S4 Penetration bias correction of TSX/TDX 2011—2014 elevation change rates

Figure S4 a–c: Penetration bias correction of TSX/TDX 2011—2014 elevation change rates

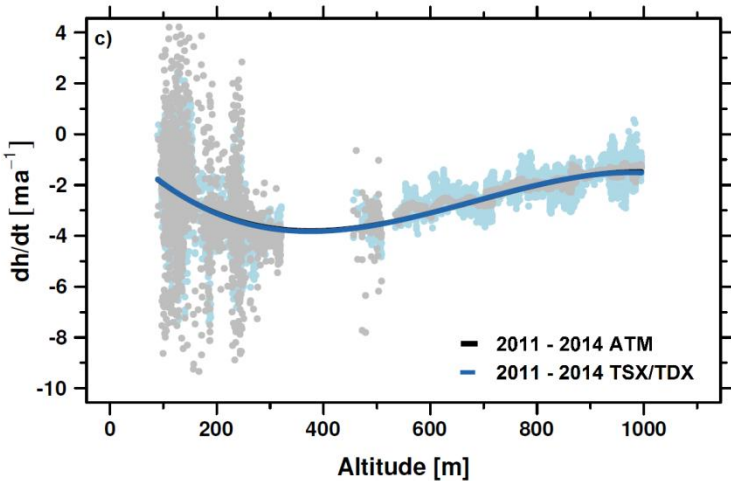
(a) Comparison of yearly elevation change rates obtained from OIB ATM LiDAR measurements (2011-11-17/2014-11-10) and TSX/TDX DEMs (2011-11-21/2014-11-03) after vertical registration of the TSX/TDX DEMs. Data is plotted against absolute ellipsoidal elevations from the resampled Bedmap 2 DEM (Fretwell et al., 2013). Grey dots: elevation change rates between 2011 and 2014 from OIB ATM. Black line: cubic function fitted to the OIB ATM measurements. Light blue dots: elevation change rates between 2011 and 2014 from TSX/TDX. Blue line: cubic function fitted to the TSX/TDX measurements



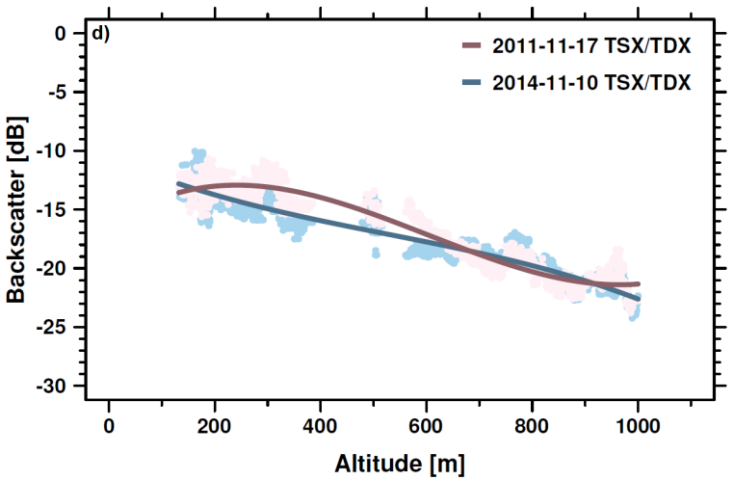
(b) Penetration bias correction model for TSX/TDX change rates. Data is plotted against absolute ellipsoidal elevations from the resampled Bedmap 2 DEM (Fretwell et al., 2013). Black dots: differences between TSX/TDX elevation change rates (r_{DEM}) and OIB ATM rates (r_{ATM}). Green line: local polynomial model fitted to the measurements.



(c) Comparison of yearly elevation change rates obtained from OIB ATM LiDAR measurements (2011-11-17/2014-11-10) and TSX/TDX DEMs (2011-11-21/2014-11-03) after vertical registration and penetration depth bias correction of the TSX/TDX elevation change map. Data is plotted against absolute ellipsoidal elevations from the resampled Bedmap 2 DEM (Fretwell et al., 2013). Grey dots: elevation change rates between 2011 and 2014 from OIB ATM. Black line: cubic function fitted to the OIB ATM measurements. Light blue dots: elevation change rates between 2011 and 2014 from TSX/TDX. Blue line: cubic function fitted to the TSX/TDX measurements

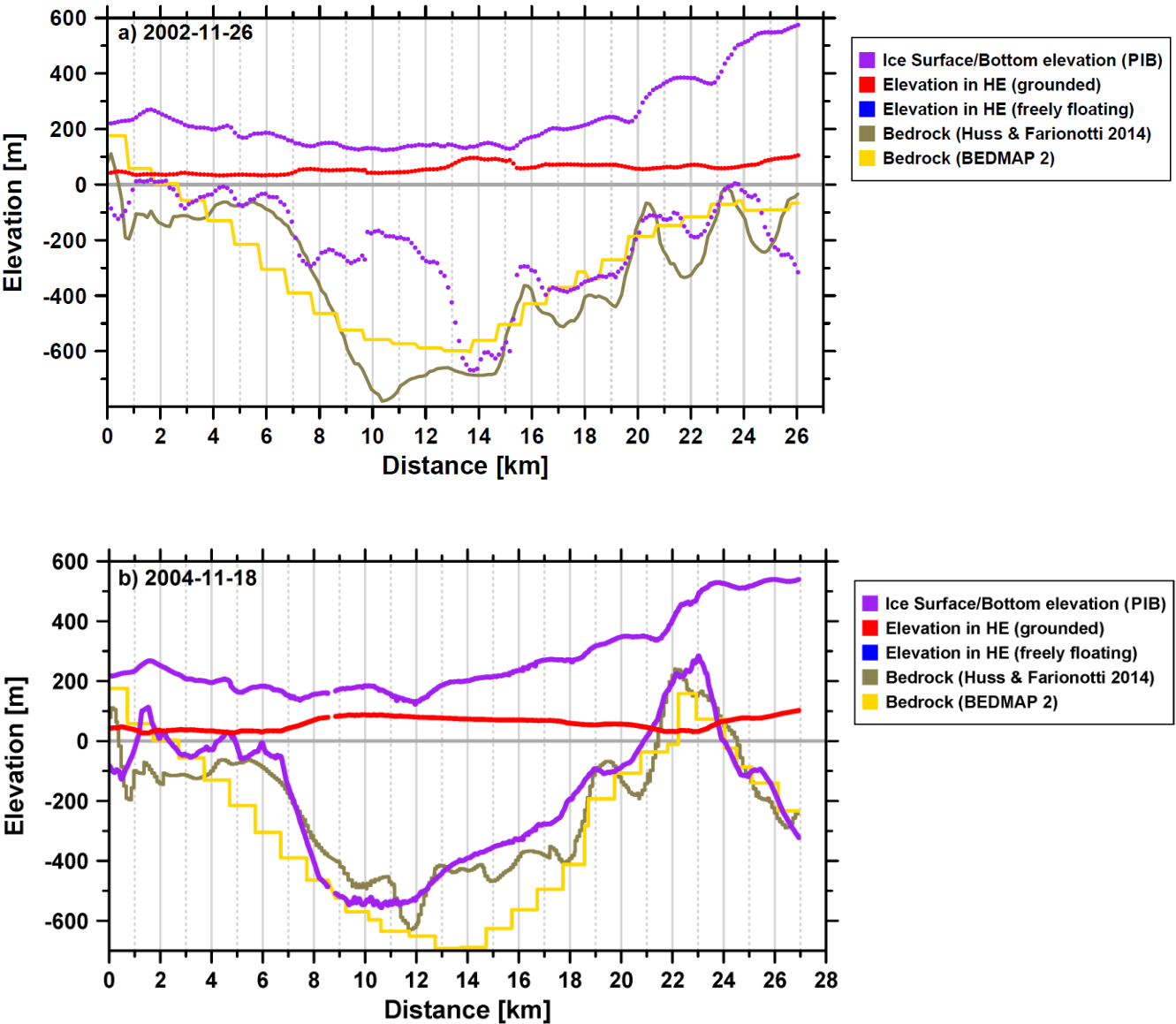


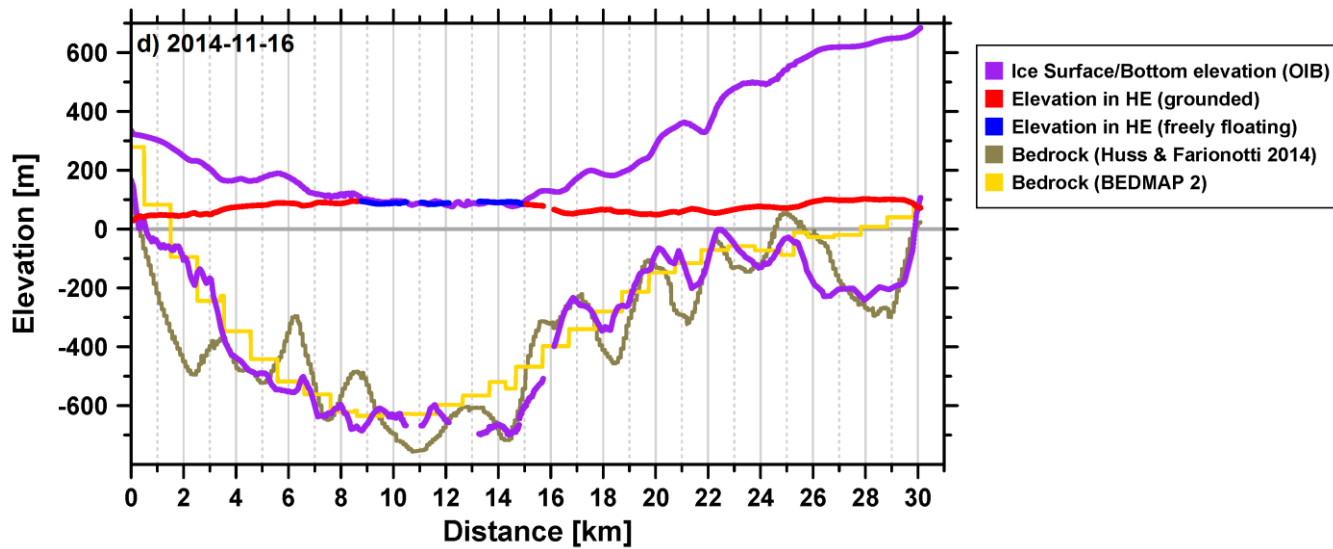
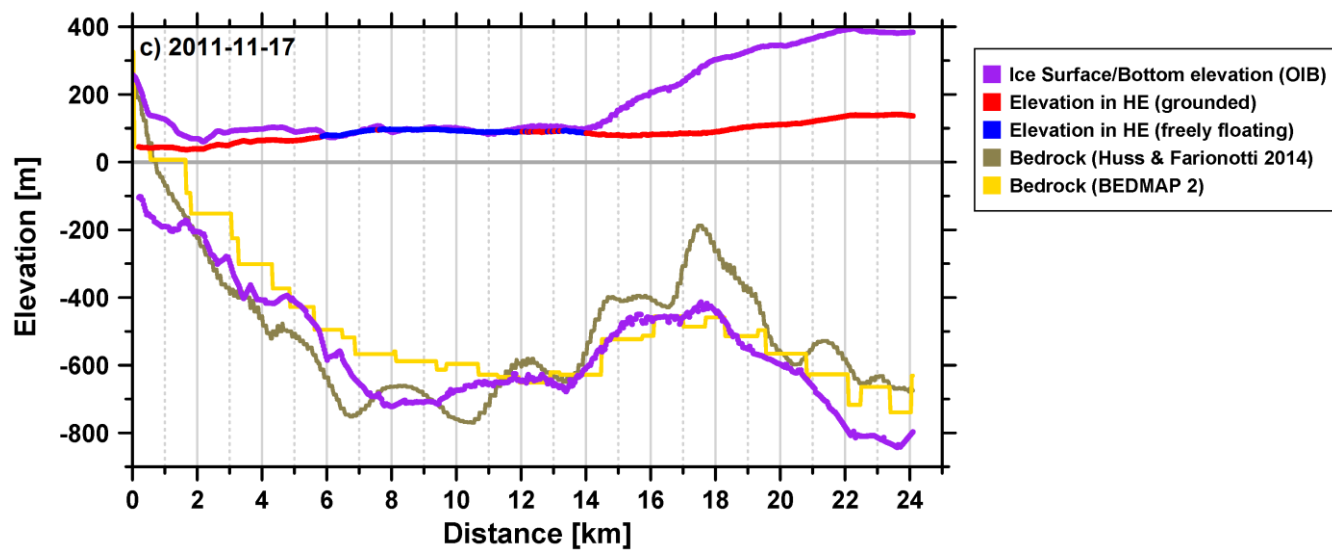
(d) Comparison of median filtered backscatter values of TSX/TDX acquisitions on 2011-11-21 and 2014-11-03. Data is plotted against absolute ellipsoidal elevations from the resampled Bedmap 2 DEM (Fretwell et al., 2013). Light pink dots: backscatter of the master image on 2011-11-21. Brown line: cubic function fitted to the 2011-11-21 backscatter values. Light blue dots: backscatter of the master image on 2014-11-03. Dark blue line: cubic function fitted to the 2014-11-03 backscatter values.

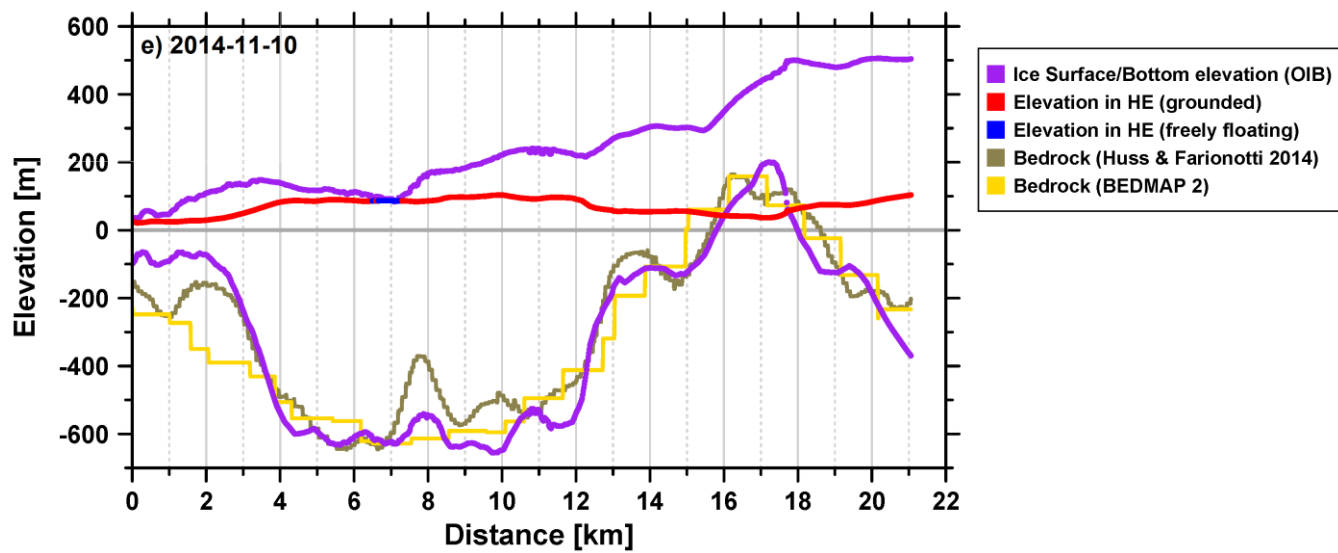


S5 Results of the hydrostatic height anomaly calculations

Figure S5 a–e: Fulfillment of the hydrostatic equilibrium assumption from hydrostatic height anomaly calculations along PIB and OIB profiles. Profiles are to read from left to right (in upstream direction). Dates of PIB and OIB flights: a) 2002-11-26, b) 2004-11-18, c) 2011-11-17, d) 2014-11-16, e) 2014-11-10. Purple dots: PIB/OIB ice surface/bottom elevations. Ice surface elevation is taken from PIB/OIB ATM measurements and ice bottom elevation is calculated by subtracting OIB/PIB ice thickness from ice surface elevation. Brown line: Bedrock elevation from Huss and Farinotti, 2014. Yellow line: Bedrock elevation from Bedmap 2 (Fretwell et al., 2013). Red and blue dots: calculated ice surface elevation in hydrostatic equilibrium e_{he} and information on hydrostatic equilibrium (blue: freely floating ice, red: grounded ice)







S6 Estimation of grounding line positions along profiles of surface velocity, bedrock topography, ice elevations and hydrostatic height anomalies

Figure S6.1-4S6 a-d: Estimated grounding line positions on profiles a-d-1-5 (for location see Fig. 6) based on surface velocities and elevation change rates from TDX (black line) and hydrostatic height anomalies. Distances for Fig. S6.1-4S6 a-d are relative to the 1996 grounding line glacier front in 2007 (Fig. 6). OIB ice surface/bottom elevations and modeled bedrock topography (Fretwell et al., 2013; Huss and Farinotti, 2014). F: Front, GL grounding line. Brown line: Bedrock elevation from Huss and Farinotti (2014). Yellow line: Bedrock elevation from Bedmap 2 (Fretwell et al. 2013). Green: freely floating ice, Red: partially floating/grounded ice, Blue lines: OIB/PIB ice surface/bottom elevations.

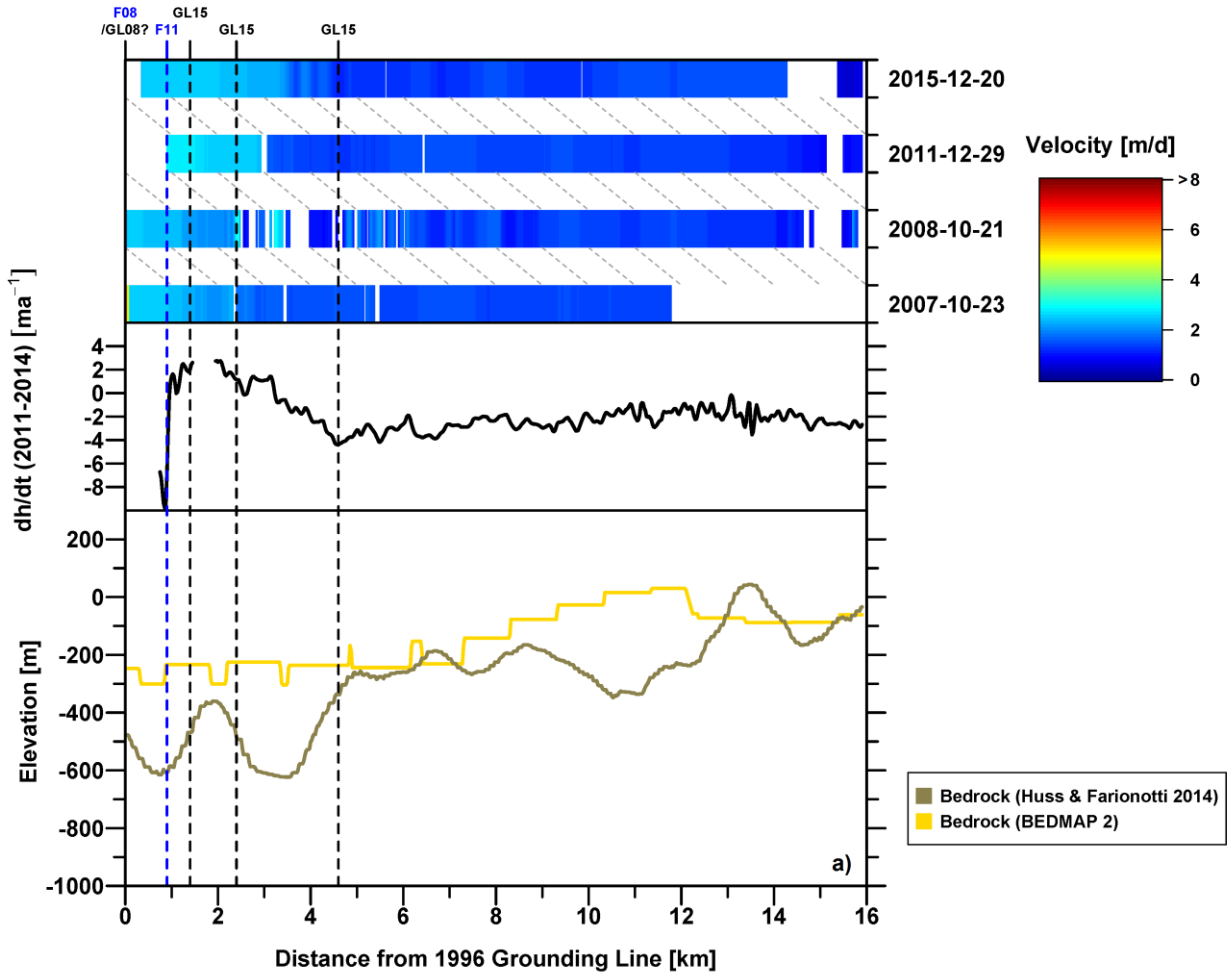


Figure S6 a exhibits that in contrast to Fleming Glacier, on Airy Glacier, velocity data extracted along the center line Profile a1 (Fig. 6) does not show signs of acceleration between 2007 and 2008. Furthermore in 2008 the front in 2008 was still directly located at the 1996/2007 grounding line. Hence, we assume It is likely that in 2008 at this time Airy Glacier was still grounded at the 1996/2007 grounding line position. However, in 2011 the front had retreated behind the 1996/2007

grounding line. This shows that the grounding line must have had retreated from its 1996 position by this time. Hydrostatic height anomalies in 2011 (Fig. 6, Track c) indicate that the ice is recently grounded on a hill, located ~2 km upstream, which reaches to the subglacial trough. However, An abrupt change in the 2011–2014 TSX/TDX elevation change rates is visible at 4–5 km upstream, which coincides with the edge of the subglacial trough in the bedrock data. A similar extent of accelerated surface velocities is visible on the velocity profiles in 2011 and 2015. Hence, this location is likely the recent location of the grounding line. A distinct area of low elevation change rates which is visible on the TSX/TDX dh/dt map (Fig. S7) suggests that the entire Airy Glacier tongue may be currently floating up to ~4 km upstream of the 1996 grounding line. However, this is in contrast to the buoyancy calculations (Fig. 6, Track 3) which indicate that in 2011 the ice was grounded on a hill ~2 km upstream, which reaches to the subglacial trough. Since we assume that the hydrostatic height anomalies from 2011 are the most reliable source of information for grounding line detection we have, we delineated the recent grounding line accordingly. However, we cannot rule out that the ice which rested on the hill in 2011 is afloat today, as well as recent surface velocities point to a maximum upstream location of the current grounding line on Airy Glacier at the limit of the subglacial trough ~4–5 km upstream.

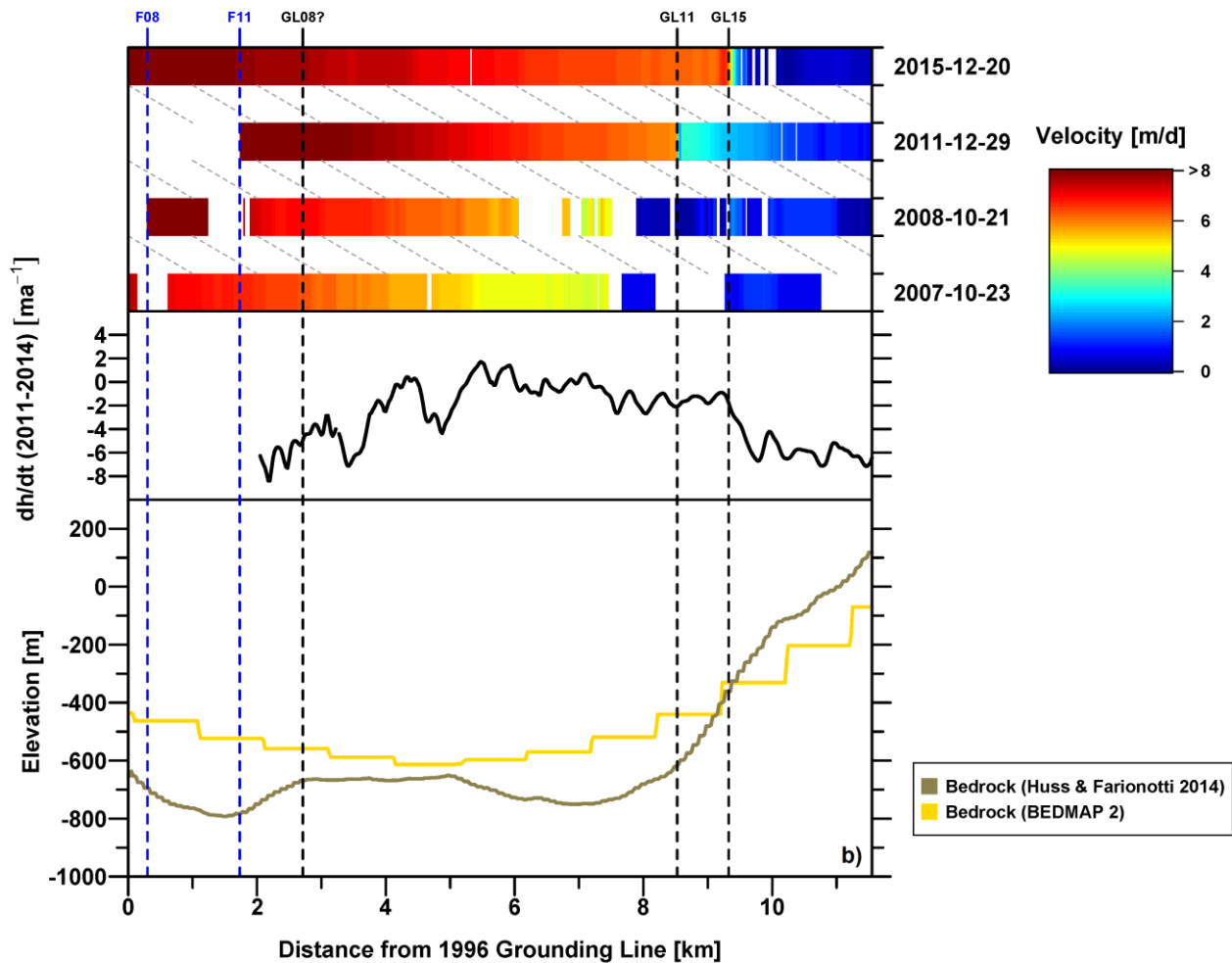


Figure S6 b depicts data extracted along Profile b (Fig. 6), close to the confluence of Fleming and Seller Glacier. Here velocity data extracted along Profile 2 (Fig. 6) shows only slight acceleration between 2007 and 2008. While the glacier had accelerated, the glacier front had retreated behind the 1996 grounding line. Hence, in 2008 the grounding line must have been located upstream of the 1996 position. The upstream extent of acceleration visible on the velocity profile suggests that in 2008 the grounding line had not retreated as far as in 2011 and 2015, yet. Although it is difficult to tell where the grounding line exactly was in 2008 in this region, the data suggest that in 2008 the grounding line had not retreated until to the edge of the trough ~10-11 km upstream, yet. A reasonable conservative estimate would be that the grounding line in 2008 was located on a gentle hill apparent in the bedrock data close to the 2011 front, ~4 km upstream. However, given the limitations of the modelled bedrock topography (Sect. 3) and the lack of further information, the exact grounding line position in 2008 remains vague.

In 2011 ~~however~~, the glacier had substantially accelerated and in 2015 high velocities had further propagated inland by ~1 km along the profile up to the edge of the subglacial trough at ~9 km upstream. The limit of high velocities in 2015 coincides with a marked change of the TSX/TDX 2011–2014 elevation change rates, indicating that at this position the ice starts to float. Taking into account elevation change rates and information from hydrostatic height anomaly calculations (Fig. 6, Track d). Based on these evidences, we estimate that the grounding line position in 2011 was likely already located at the edge of the subglacial trough and had probably further retreated by ~1 km in 2015.

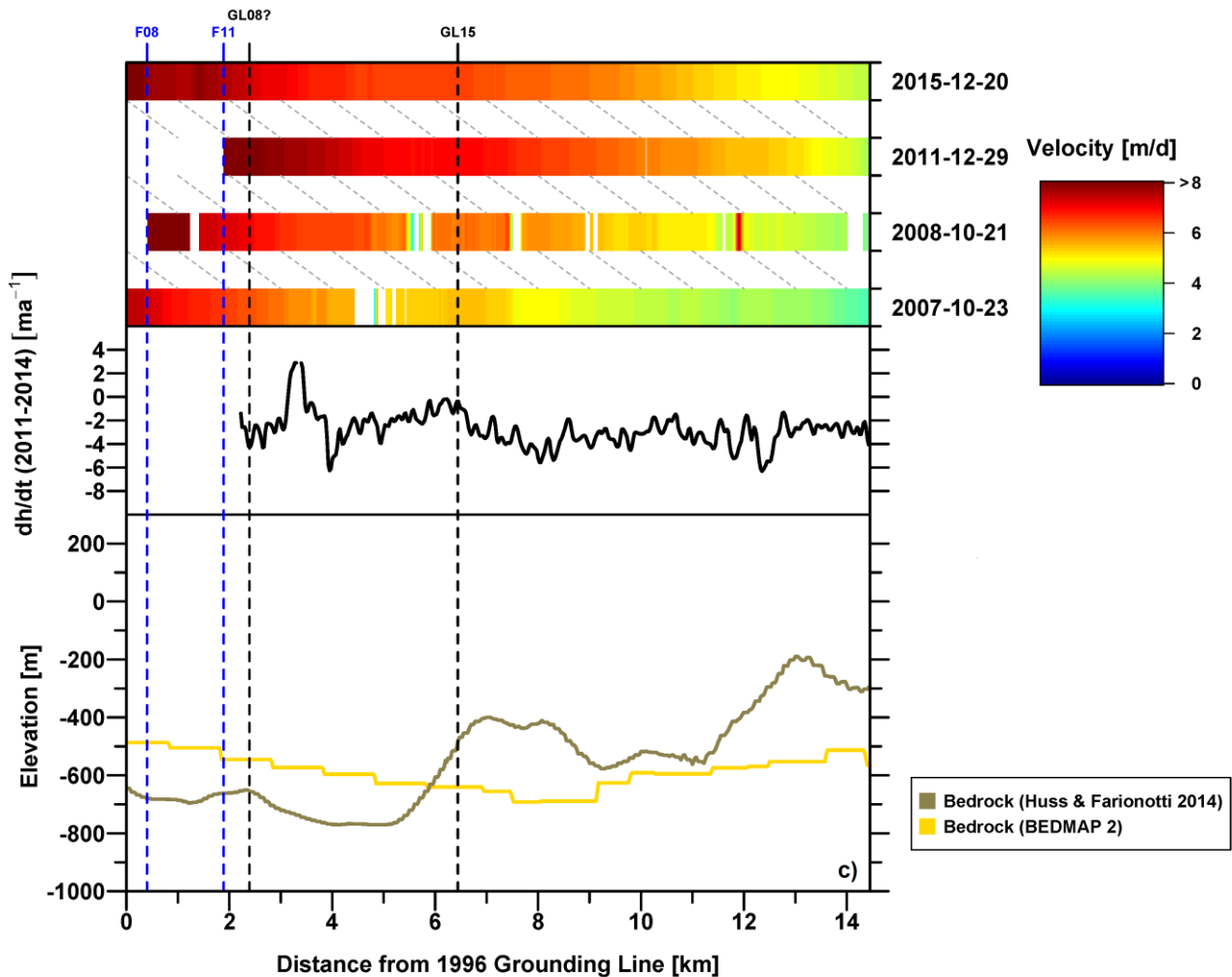


Figure S6 c reveals that ~~v~~Velocities extracted along Profile 3 (Fig. 6) show an upstream propagation of high velocities between 2007 and 2008 of up to 12 km. In 2008, the glacier front had retreated behind the 1996 grounding line. This indicates that in 2008 the grounding line must have been located ~~had already migrated~~ upstream from its 1996 ~~2007~~ position ~~in 2008. However, due to the limitations of estimating the grounding line based on velocity information only, the extent of~~

~~the retreat in 2008 remains unclear. Since the acceleration had not yet propagated as far upstream as in 2011 and 2015, we assume that the 2008 grounding line was located more seawards than in 2011 and 2015 respectively. A possible grounding line location position in 2008 is a smaller hill visible in the bedrock data close to the front in 2011 at ~2, 5 km upstream, close to the front in 2011. However, due to the limitations of the bedrock data (Sect. 3) and the lack of further evidence, the precise 2008 position remains unclear. ~4 km upstream of the front in 2007.~~

~~In 2011 and 2015 the glacier had markedly accelerated, indicating that the grounding line had further retreated. The 2011/2014 TSX/TDX dh/dt profile shows a slight drop in ice thinning rates at ~6 km upstream of the 1996 grounding line which coincides with the edge of the subglacial trough in the bedrock data. Although the change of elevation change rates is not as pronounced as on the other profiles, hydrostatic height anomalies along an OIB-flight path running close to this position (Fig. 6, Track 4) show that in 2014 the glacier was freely floating downstream. OIB data acquired in 2011 and 2014 (Fig. 6, Track 3 and 5) indicate that the ice is currently grounded upstream of the hill chain. We hence estimate the recent grounding line position to be located at the edge of the subglacial trough ~6 km upstream.~~

~~In 2011 the glacier front had retreated behind the 1996/2007 grounding line. Although no clear pattern of lower ice elevation change rates is visible on the extracted 2011–2014 dh/dt data, hydrostatic height anomalies calculated for an OIB flight path in 2014 (Fig. 6, Track d) show that the glacier tongue downstream of the hill chain ~8 km upstream was freely floating in 2014. Although in 2011 high velocities had gradually propagated up to 14 km upstream of the front in 2007, OIB data acquired in 2014 (Fig. 6, Track e) reveal that the ice upstream of the hill chain at ~8 km is not freely floating.~~

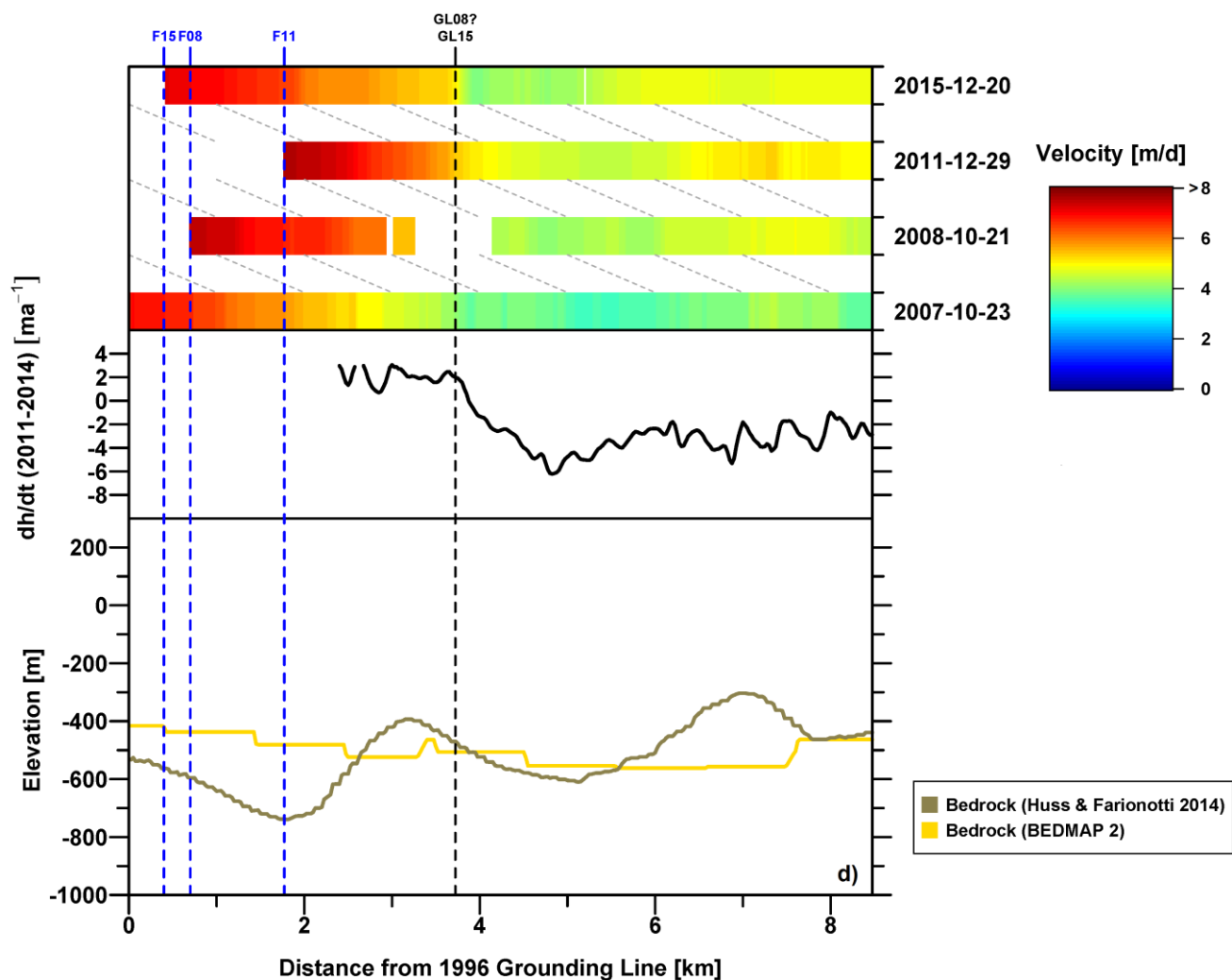


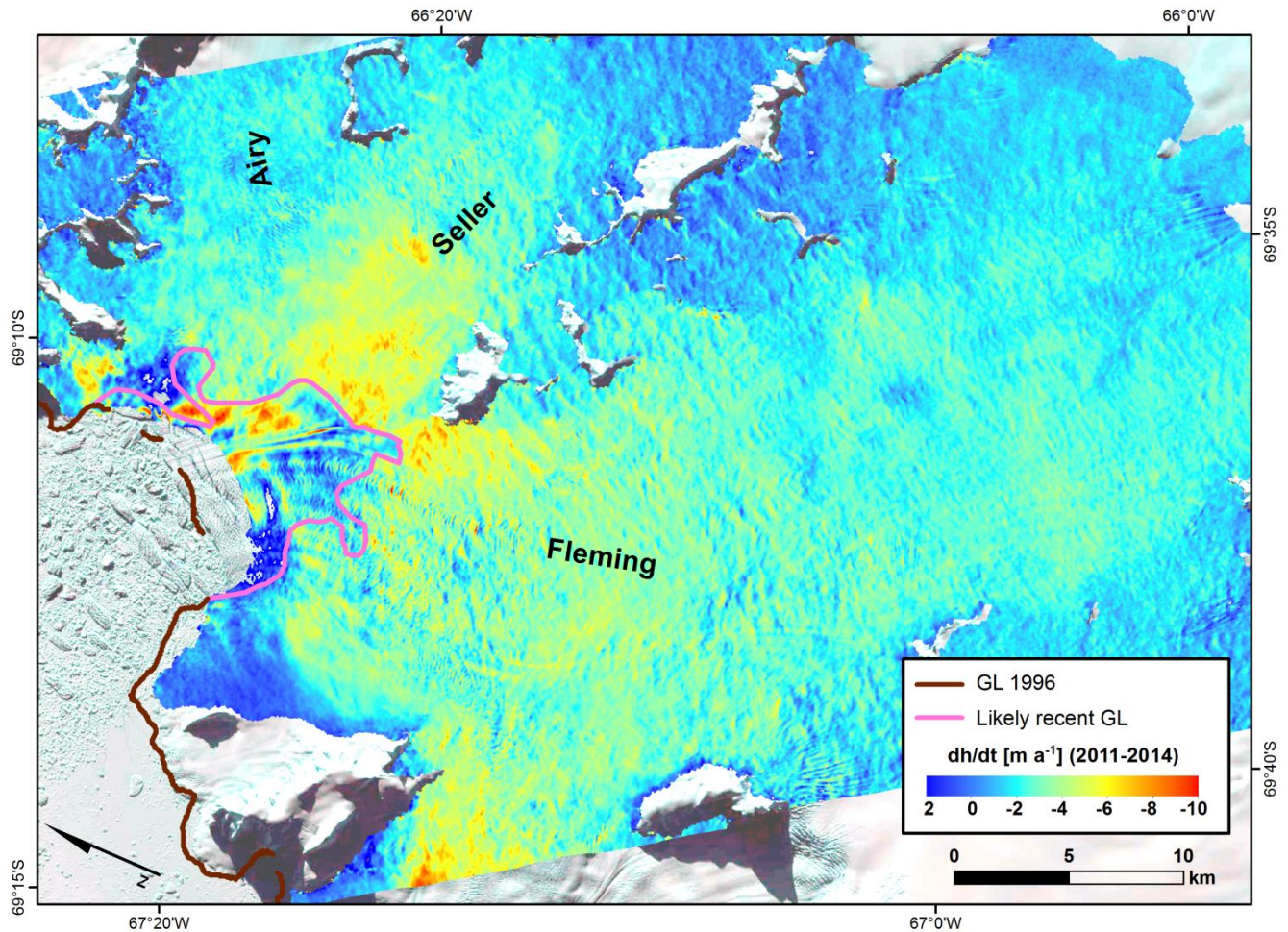
Figure S6 d shows that between 2007 and 2008 a pronounced acceleration is detected along Profile d (Fig. 6). In 2008 the glacier front had retreated behind the 1996 grounding line for the first time. Thus, the grounding line in 2008 must have been located upstream of the 1996 position in 2008. The velocity patterns look similar in 2008, 2011 and 2015, suggesting that no further grounding line retreat occurred after 2008. We hence assume that the 2008 grounding line position is more or less identical with the recent location. However, given the lack of further evidence the precise grounding line position of 2008 remains vague.

A sharp increase in surface velocities is visible at 3–4 km upstream in 2011 and 2015. This location is consistent with a pronounced change in ice thinning rates visible on the TSX/TDX 2011–2014 dh/dt profile and on the map (Fig. S7), which indicates that the ice is currently floating downstream. On the other hand, hydrostatic height anomalies along OIB tracks 4

and 5 in Figure 6 reveal that the ice is grounded further upstream. Additionally the 2011 front was located ~ 2 km upstream of the grounding line. Combining all of the information, we estimate the recent grounding line to be located at a position ~3–4 km upstream. However, the modeled bedrock data suggests that the grounding line is not situated on the upslope but on the downslope side of a subglacial hill. We hence cannot rule out that the hill may be shifted to the north in the modelled bedrock data, or that – if the modelled bedrock topography is correct - the recent grounding line is located ~ 1 km further to the south. At the lower part of Fleming Glacier, the velocity data and a pronounced change in the elevation change rates extracted along Profile 4 (Fig. 6) indicate that the current GL is located some hundred meters behind a hill chain, visible in the modelled bedrock data ~4 km upstream. A comparable velocity pattern in 2008 suggests that at this time the glacier was already grounded at a similar location. It is likely that the glacier is currently grounded on the hills, but the correct representation of the topography may be distorted in the modeled bedrock data.

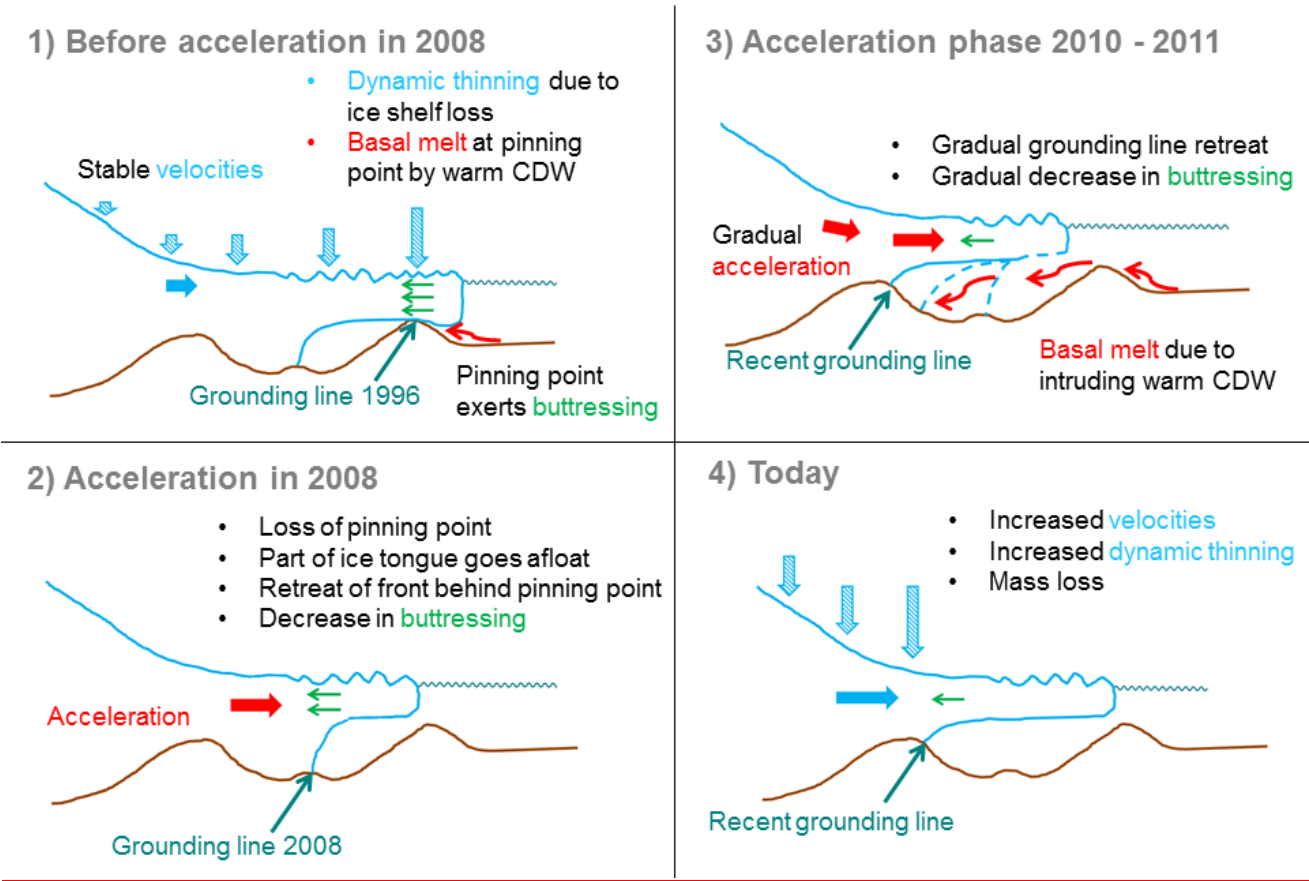
S7 Recent grounding line and TSX/TDX 2011–2014 elevation change

Figure S7: Glacier surface elevation change on Fleming Glacier between 2011 and 2014 derived from TSX/TDX bistatic and monostatic acquisitions with final solution of the likely recent grounding line location (pink line). Brown line: grounding line in 1996 from Rignot et al. (2005) and Rignot et al. (2011). Background: Mosaic of two Landsat-8 „Natural Color” images, acquired on 2015-09-16 ©USGS.



S8 Evolution of the grounding line before 2008–today

Figure S8: Schematic drawing of the ungrounding history of Fleming Glacier from before 2008 to today. Interpretation is based on the data presented in this study and in the literature. Figures are not drawn to scale.



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