

Response to the Interactive comment on

“Basal drag of Fleming Glacier, Antarctica, Part B: implications of evolution from 2008 to 2015”

by Chen Zhao et al.

Anonymous Referee #1

Received and published: 5 Feb 2018

We are grateful to Reviewer 1 for the positive and constructive suggestions to improve our paper. We have addressed the comments below. The line numbers in the responses are based on the revised manuscript without change track.

Please note that Mathieu Morlighem created the ice thickness data for the Fleming Glacier system using the mass conservation method, which is very important for most experiments done in this study. We do value his contribution to this paper, so we add him as the co-author in the revised text.

In the revised companion paper (Zhao et al., companion paper), we implemented a new sensitivity test to the enhancement factor (E). It reveals that the optimal value of $E = 1.0$ should be chosen as the enhancement factor in the CONTROL experiment. Accordingly, we re-ran all the simulations in this study with $E = 1.0$, and the high basal shear stress band near the ice front in 2008 has decreased into high basal shear spots, which are suspected of being artefacts of the inversion process and are discussed below. We modified the text and figures accordingly. All other result and interpretations are not qualitatively changed from the original manuscript.

General comments

This paper presents some interesting results suggesting that glacier-bed interactions have an important role in the dramatic speedup of Fleming Glacier, Antarctica from 2008-2015. Recent work by Walker and Gardner posited that abnormally warm ocean temperatures in Marguerite Bay over this time period caused the observed changes in the glaciers that fed the former Wordie Ice Shelf. Friedl et al. 2017 used a combination of several remote sensing data sets to show that large areas near the terminus of Fleming Glacier ungrounded between 2008-2015. These data sets showed that Fleming Glacier lies on a retrograde bed slope, and thus is susceptible to runaway retreat via marine ice sheet instability. Zhao et al. argue that both of these explanations leave out a key factor, the interaction of the glacier and its bed. The authors used inverse methods to estimate the basal shear stress under Fleming Glacier in 2008 and 2015. This analysis revealed a band of high basal shear stress near the terminus in 2008 that is no longer present in 2015. They argue that the retreat of the glacier off of this region of high basal friction may also be a factor in the subsequent speedup.

The large changes that Fleming Glacier exhibited make it a valuable test case for understanding glacier change in Antarctica. The authors' results suggest that glacier-bed interactions are an important factor, in addition to ocean melting and geometric instability, in understanding the recent behavior of the glaciers that fed the former Wordie Ice Shelf. While I recommend publication, the work could be improved on a number of fronts. Drawing conclusions about the physics of glacier-bed interactions from the results of inverse methods can be difficult because, as the authors

acknowledge, any one feature could be an artefact. Many of the arguments made in the paper are speculative and I think this should be made clearer. Suppose that the high-basal shear band they claim to find at the terminus of Fleming Glacier in 2008 were merely an artefact – what would that mean for the physics?

This is a great suggestion. As mentioned above, in the modified companion paper (Zhao et al., companion paper), we speculate that the high basal shear spots near the ice front may be artefacts. However, the possibility of the high basal friction spots being real features, which might be caused by pinning points near the 1996 grounding line position is not excluded. Based on the inferred basal shear stress (Fig. 3a) and height above buoyancy (Fig. 5a), the 1996 grounding line position may have not retreated prior to Jan 2008, and Friedl et al. (2018) also suggested that the grounding line position may have retreated behind the 1996 position after Jan-Apr 2008.

The discussion in the manuscript has been modified to respond to the reviewer's comment adding (Line 361-365): "If the high basal resistance spots are artefacts, ungrounding of this region in early 2008 is less viable as an explanation for an abrupt increase in ice flow speed, since the loss of backstress would be more gradual. In this case, positive feedbacks, such as the marine ice sheet instability or the basal melt feedback, are even more likely to explain the FG's recent behavior."

In general, with regard to inverse methods, small features can more easily arise as inversion artefacts than larger features. Small basal shear stress features may be locally balanced by extensional/compressional stresses in the ice without needing to balance the gravitational driving stress. For features with larger horizontal scales basal shear stress must approximately balance the driving stress and these features are less likely to be artefacts. All features discussed in the paper arising from the inversion process, aside from the sticky spots near the 2008 front, are large enough that we are confident they are robust features of the inversion and not artefacts.

Currently one possibility for rapid retreat of the grounding line is that there were some sticky spots near the front, and rapid retreat occurred when the ice ungrounded from these sticky spots.

Specific comments

First, I think the abstract could be improved by (1) cutting many of the details that are covered in the discussion section and (2) giving a clearer statement about what this paper adds to the existing knowledge. The main precedents that the authors draw from are Walker and Gardner 2017 and Friedl et al. 2017. What do these two papers conclude, and how do the authors' conclusions agree with or depart from them? For example, both the present work and Friedl et al. 2017 argue that the speedup and thinning of Fleming Glacier is a consequence of ongoing marine ice sheet instability. However, the authors argue that subglacial hydrological effects may have also initiated the retreat off of a stable bedrock high, while Friedl et al. point solely to ocean warming. To my knowledge, this work and the companion paper are the first to use inverse methods to estimate the basal shear stress of this particular site at high resolution, as opposed to low-resolution estimations for all of Antarctica. This information, which ideally would be front-and-center in the abstract, is partly obscured by details that will be addressed in the discussion section anyway. In any case, this problem is more one of presentation and not of actual content.

Thanks for the reviewer's suggestion. We added this point in the first paragraph of Section 4.1 (Line 225-229) "Although low-resolution estimation of basal shear stress

has been carried out for the whole Antarctic Ice Sheet (Fürst et al., 2015; Morlighem et al., 2013; Sergienko et al., 2014), this is the first application of inverse methods to estimate the basal friction pattern of the Fleming system at a high resolution and use the full-Stokes equations.” and modified the conclusion and abstract correspondingly.

At several points, the authors pose the question of whether either ocean warming or basal processes are the "dominant" causes of the observed changes (see lines 84, 411). The question of which process is dominant assumes that the two are additive, but if instead the relationship is causative, this question ceases to be meaningful.

The reviewer seems to think that when we say “dominant” we mean “only”. Of course, it is possible to have a small perturbation caused by one process (ocean-warming driven basal melting) and massively enhanced by another process (basal process). In this case we would describe the latter as “dominant” because it has caused the biggest change, even if there would have been no change without the former.

In the discussion section, the authors suggest that hydrological effects could destabilize the high-friction band, resulting in speedup, thinning, and ungrounding. In this scenario, hydrology-induced speedup and ungrounding create conditions where the ocean can then melt the ice shelf from underneath. One could also imagine a scenario in which ocean melting comes first and hydrological effects second. For example, ocean melting could push the glacier terminus off of a highly resistive bedrock bump, and the glacier begins to speed up and thin. The reduced overburden pressure then changes the overall hydraulic potential. The authors’ hypothesis that hydrology might have initiated the recent changes is still significant and worth considering. Nonetheless, the paper’s intent might be clearer by changing questions about which process is dominant to questions about which one came first. Finally, the authors suggest that coupled ice sheet-ocean modeling could help determine which case is more likely. This point could be expanded on further. For example, a coupled model using pre-2006 values of ocean heat flux that does exhibit a hydrology-induced destabilization would show that oceanic forcing is not necessary to explain observations.

We don’t see a need to choose to only consider which process comes first or which process is dominant. The two questions are complementary rather than contradictory. When we discuss which process is dominant we do not mean to exclude the relevance of which came first. It was not our intention to propose that the changes were initiated by the subglacial hydrologic system – we don’t have a mechanism in mind for that. We don’t see how an increase in subglacial melting can happen without an external trigger, except through increased insulation due to ice thickening such as occurs in surging glaciers. But we doubt this is happening here. We suspect the ice shelf collapse triggers a positive feedback at the bed of the fast flow region, and that once the shelf has gone, the melt rates due to the ocean warming do not make much difference. Subglacial melting probably has to be happening all the time under the fast flowing region in any case. Ocean melting/ice shelf collapse provide a triggering mechanism to the ungrounding process, and then the positive feedback between the basal sliding and subglacial water pressure at the bed kicks in. We have clarified the nature and role of this positive feedback mechanism in the Sect. 4.2 and Sect. 5.

We don’t think there is a need to expand further about designing coupled experiments as that is well outside the scope of this paper.

The text gives conflicting statements about the authors' degree of confidence in the veracity of their conclusions. For example, in line 314 the authors assert that the disappearance of the high friction band near the calving front is a "likely" trigger for the subsequent retreat, but at other points they equivocate about whether this feature is real or merely an artefact. A lack of complete certainty about this resistive band is entirely reasonable but the paper would be improved if it were more consistent in what kind of assertions are made.

Thanks for pointing this out. Based on the modified companion paper (Zhao et al., companion paper), we speculate that the high basal shear spots in 2008 (rather than the band of high shear seen in the previous version) may be artefacts but we do not rule out the possibility of high friction spots as a real feature caused by the pinning points at the 1996 grounding line. For consistency we modified the text in the manuscript accordingly (Line 359-361).

For the example mentioned by the reviewer, we modified "likely" to "possible" (Line 383). Under this speculation, if the sticky spots were totally artefacts, the reduction in basal drag would be likely due to the positive feedbacks between the basal sliding and basal subglacial water.

A numerical experiment could shed some more light on whether the resistive band near the terminus is real or not. The methods section describes inferring the basal friction using the 2008 ice thickness and the 2015 velocity to examine whether the result is sensitive to the geometry. In this vein, the authors could compute a velocity using the 2015 basal friction and the 2008 thickness. How well does this computed velocity agree with observations, weighted by the error variances? Is the misfit worse than that of the velocity computed using both the 2008 thickness and basal friction? If so, by how much? The presence of a resistive band at the terminus would be doubtful if a basal friction field without this feature can explain the 2008 data just as well as a basal friction field with this feature.

The basal friction field without the sticky spots cannot explain the 2008 data. Although we are not sure whether the high basal drag spots in 2008 are real or not, we are sure that the basal drag of high velocity regions in 2008 should not be as small as that in 2015. However, we still tried this experiment as the reviewer suggested. Results show that the simulated surface velocity was nearly 2.5 times the observed surface velocity of 2008 near the ice front. So we cannot use the suggested experiment to say that the sticky spots are an artefact.

Technical corrections

17-21: Flip the order of the sentences starting with "To explore the mechanism underlying these changes..." and "Recent observational studies..."

Modified.

23-28: Giving too much justification in the abstract obscures your overall point, this could be moved to the discussion.

We agree to remove the sentence about the grounding line position in 2008, but we think the comparison results between 2008 and 2015 should appear in the Abstract.

66-69: "As a marine-type glacier system..." Rephrase or break up into 2 sentences.

The whole sentence has been modified into "As a marine-type glacier system residing

on a retrograde bed with bedrock elevation as much as ~800 m below sea level (Fig. 1c), the Fleming system is hence potentially vulnerable to marine ice sheet instability (Mercer, 1978; Thomas and Bentley, 1978; Weertman, 1974). The acceleration and greater dynamic thinning of the FG over 2008-2015 suggests the possible onset of unstable rapid grounding line retreat (Walker and Gardner, 2017; Zhao et al., 2017), which has been confirmed by Friedl et al. (2018). ” (Line 74-79).

73-74: "None of these past studies have modelled the glacier system and hence these hypotheses are untested." This suggests that modelling is the only way to really test these hypothesis. It's better to just say that the precise nature of the feedbacks hasn't been established and that you will test them using models.

Thanks for pointing this out. We modified this sentence into “An alternative hypothesis is that the recent changes arise from feedbacks in the dynamics of the evolving glacier, possibly involving the subglacial hydrology. The examination of changes in basal shear stress distributions between 2008 and 2015 in this modelling study provides a first step in exploring possible feedback hypotheses. ” (Line 83-87).

88-90: "Changes in basal shear stress..." Rephrase this sentence.

Reviewer 3 has suggested deleting this sentence, since it is not helpful here. We agree with Reviewer 3, so we delete this sentence.

162-165: "To explore their relative impacts..." While this experiment is a good sanity check, the result isn't essential to making your point and this could be relegated to a supplement.

Thanks for the suggestion. We have moved this part into the Sect. S1 in the supplementary material.

294-296: Make this "The change in area", and "additional evidence supporting the hypothesis of rapid grounding line retreat".

We modified it into “This change in area” and “additional evidence supporting the hypothesis of rapid grounding line retreat” (Line 354-356).

313-315: Could the basal resistance band at the front be an artefact of neglecting backstress at the terminus from mélange or sea ice?

We have discussed this in Sect. 4.4 of the revised companion paper (Zhao et al., companion paper). The ice mélange back force ($\sim 1.1e7 \text{ N m}^{-1}$) used to prevent the rotation of an iceberg at the calving front (Krug et al., 2015) could account for the equivalent of up to ~2.3 m sea level in terms of ice front boundary condition. The experiment with the sea level increased by 10 m shows that the high basal shear spots are decreasing but have not disappeared. The situation at the front is complicated. Sea level, bedrock/ice thickness uncertainty, mélange backstress, ice front positions, these things can all impact on our inversion near the ice front.

324-327: Overly long sentence, break up into 2 sentences.

Modified into “For a glacier lying on a retrograde slope in a deep trough, the grounding line may be vulnerable to rapid retreat without any further change in external forcing, once its geometry crosses a critical threshold, which is the marine ice sheet instability hypothesis (e.g., Mercer (1978); Thomas and Bentley (1978); Weertman (1974)). A similar theory has been proposed on the prospective rapid retreat of Jakobshavn Isbræ in West Greenland without any trigger after detaching from a pinning point (Steiger et al., 2017).” (Line 394-399).

326-327: "...as in the rapid retreat of Jakobshavn Isbrae in West Greenland (Steiger et al., 2017)." There were other factors in the retreat of Jakobshavn, see Motyka et al. 2011 and Holland et al. 2008.

Yes, we agree with the reviewer. We should have made it clear that we were talking about the future behavior of the Jakobshavn here. Steiger et al., 2017 found that after decades of stability and with constant external forcing, the grounding lines of Jakobshavn may retreat rapidly without any trigger due to losing the pinning-points. To make it clearer, we modified it into "A similar theory has been proposed on the prospective rapid retreat of Jakobshavn Isbr e in West Greenland without any trigger after detaching from a pinning point (Steiger et al., 2017)."

336-340: Run-on sentence, break up into 2 or 3 sentences.

Modified into "If the system remains out of balance and continues to thin, the grounding line could eventually move across this bed obstacle. If this occurs, the grounding line is then likely to retreat rapidly down the retrograde face of the FG upstream basin, likely to be accompanied by further glacier speed up and dynamic thinning." (Line 409-413)

400-402: "...hard to say how much forcing would be needed to push the grounding line into it." Rephrase.

Modified into "More thinning would be needed to destabilise the upstream basin, and it is hard to estimate how much forcing would be needed to push the grounding line into the upstream basin boundary." (Line 497-499).

414: Change "simulate" to "estimate".

Modified.

References

- Friedl, P., Seehaus, T. C., Wendt, A., Braun, M. H., and H oppner, K.: Recent dynamic changes on Fleming Glacier after the disintegration of Wordie Ice Shelf, Antarctic Peninsula, *The Cryosphere*, 12, 1-19, 2018.
- Krug, J., Durand, G., Gagliardini, O., and Weiss, J.: Modelling the impact of submarine frontal melting and ice m lange on glacier dynamics, *The Cryosphere*, 9, 989-1003, 2015.
- Mercer, J. H.: West Antarctic ice sheet and CO2 greenhouse effect: a threat of disaster, *Nature*, 271, 321, 1978.
- Steiger, N., Nisancioglu, K. H.,  kesson, H., de Fleurian, B., and Nick, F. M.: Non-linear retreat of Jakobshavn Isbr e since the Little Ice Age controlled by geometry, *The Cryosphere Discuss.*, 2017, 1-27, 2017.
- Thomas, R. H. and Bentley, C. R.: A model for Holocene retreat of the West Antarctic Ice Sheet, *Quaternary Research*, 10, 150-170, 1978.
- Walker, C. C. and Gardner, A. S.: Rapid drawdown of Antarctica's Wordie Ice Shelf glaciers in response to ENSO/Southern Annular Mode-driven warming in the Southern Ocean, *Earth and Planetary Science Letters*, 476, 100-110, 2017.
- Weertman, J.: Stability of the Junction of an Ice Sheet and an Ice Shelf, *Journal of Glaciology*, 13, 3-11, 1974.
- Zhao, C., Gladstone, R., Zwinger, T., Warner, R., and King, M. A.: Basal friction of Fleming Glacier, Antarctica, Part A: sensitivity of inversion to temperature and bedrock uncertainty, *The Cryosphere*, companion paper. companion paper.
- Zhao, C., King, M. A., Watson, C. S., Barletta, V. R., Bordoni, A., Dell, M., and Whitehouse, P. L.: Rapid ice unloading in the Fleming Glacier region, southern

Antarctic Peninsula, and its effect on bedrock uplift rates, *Earth and Planetary Science Letters*, 473, 164-176, 2017.

Response to the Interactive comment on

“Basal drag of Fleming Glacier, Antarctica, Part B: implications of evolution from 2008 to 2015”

by Chen Zhao et al.

Anonymous Referee #2

Received and published: 6 Mar 2018

We are grateful to Reviewer 2 for the positive and constructive suggestions to improve our paper. We have addressed the comments below. The line numbers in the responses are based on the revised manuscript without change track.

Please note that Mathieu Morlighem created the ice thickness data for the Fleming Glacier system using the mass conservation method, which is very important for most experiments done in this study. We do value his contribution to this paper, so we add him as the co-author in the revised text.

In the revised companion paper (Zhao et al., companion paper), we implemented a new sensitivity test to the enhancement factor (E). It reveals that the optimal value of $E = 1.0$ should be chosen as the enhancement factor in the CONTROL experiment. Accordingly, we re-ran all the simulations in this study with $E = 1.0$, and the high basal shear stress band near the ice front in 2008 has decreased into high basal shear spots, which are suspected of being artefacts of the inversion process and are discussed below. We modified the text and figures accordingly. All other result and interpretations are not qualitatively changed from the original manuscript.

General comments

This paper, using diagnostic inverse modeling of basal conditions, discusses the possible causes of the retreat of Fleming glacier observed between 2008 to 2015. In particular, the potential acceleration induced by the production of water by frictional heating at the base of the glacier is discussed. This paper is well written, even if some sentences are too long and some figures can be improved. I have made below some suggestions that I believe could improve the manuscript.

Specific comments

line 62: nearly twice or more than twice?

“More than twice” is more suitable here. Modified.

line 95: I don't really see where in Gladstone et al. (2017) inverse methods are used?

The reference is deleted here.

line 123: define what is `bed_zc`

`bed_zc`, has been defined using Eq. (1). To clarify it better, we modified the sentence into “The bedrock data, `bed_zc` (Fig. 2b), ...” (Line 136)

line 127: S_{2008} is not the "surface DEM in 2008" but the "surface elevation in 2008".

We modified it into “where S_{2008} is the surface elevation in 2008 combined from two DEM products as discussed in Zhao et al. (companion paper),...” (Line 140-141).

line 134: (on the same line) The "2008 velocity" should be "The 2008 velocity

dataset"

Modified.

line 155: the assumption that all the ice is grounded is for the inverse method? May be you can specify already here that floating ice will be deduced as the place where basal stress is lower than a threshold? It is not clear all along the manuscript if there is still a floating part or not on Fleming glacier and it would help if it could be mentioned more clearly in the introduction.

Yes, the assumption that all the ice is grounded is for the inverse method. The floating ice will be deduced where basal shear stress is lower than a threshold. To clarify this, we added a sentence "This assumption might be incorrect for the main branch of the FG, and we evaluate it based on the deduced floating area where the inferred basal shear stress is lower than a threshold, which is discussed in Sect. 4.1." (Line 172-175).

In the introduction, we declared that the ice front position in Apr 2008 (dark blue line in Figs. 1b and 1c, Wendt et al. (2010)) has almost coincided with the 1996 grounding line position (Line 62). For this study, we assume that all the ice is grounded and the ice front position is same as the 1996 ice front position, which is added in Line 171-172.

line 175: it should be mentioned that Eq. (4) is valid under the assumption of $N = 0$

Thanks for the suggestion. We added one sentence after this equation (Line 208-210). "Here we assume that the water pressure in the subglacial hydrologic system is given by the ice overburden pressure, which is equivalent to assuming that the effective pressure at the bed, N , is zero (Shreve, 1972)"

line 186: here it should be mentioned that Eq. (6) is derived under the assumption of a perfect connectivity of the basal hydrology system with the ocean

Thanks for the suggestion. We did say that we used a simpler hydrostatic balance. In order not to get tangled up with the interior hydraulic modeling, we add a sentence to qualify this "This expression for Z_* assumes a perfect connectivity of the basal hydrology system with the ocean. This is appropriate for the present study where we are exploring the degree of grounding of the fast flowing regions of the FG over the downstream basin." (Line 217-220).

line 192: C is not a vector (not in bold)

Modified.

line 380: The increase of the amount of melt water should be quantified by integrating the frictional heating over the bedrock. But it should be also discussed that more melt doesn't necessarily induce an acceleration of the glacier as the basal hydrology system is evolving dynamically to adjust this surplus of water. The link of basal sliding with basal water should be clarified, and specifically it should be mentioned that the important variable is not the amount of water but its pressure. And this later quantity is not evaluated in the present work.

The amount of melt water has been quantified based on the Eq. S1 in the Sect. S2 and shown in Fig. S4 in the supplementary material. We present the distribution of the basal melt water along with a 2015-2008 difference plot rather than presenting the integrated total. This approach demonstrates the patterns and regions of important differences, which would not be apparent in an integrated quantity. Also, the

integrated basal melt would be sensitive to the region of integration. We mentioned this in Line 468-471.

We have clarified the positive feedback mechanism in Sect. 4.2 (Line 295-301). “Since the reduction of effective pressure is the key process to enhance sliding, this positive feedback is dependent on a positive feedback of melt water generation to water pressure. This dependence can break down when there is sufficient basal water to generate efficient drainage channels (Schoof, 2010). However, such efficient channelization in the subglacial hydrologic system is typically associated with seasonal surface meltwater pulses reaching the bed (Dunse et al., 2012), a process that is not expected to occur for Fleming Glacier (Rignot et al., 2005).”

For the subglacial water pressure, it is not possible to evaluate this quantity without a hydrology model, which is beyond the scope of this study.

line 430: Can the buttressing exerted by the pinning band in 2008 be quantified in a more rigorous way? A complementary experiment would be to remove this band of high friction (by setting no friction there) and to see how the velocity field is modified upstream. This would directly quantify the increase of velocity induced by an instantaneous loss of the pinning band. The difference between this velocity field and the 2015 one would indicate places where a decrease of basal shear stress is necessary to explain the 2015 velocity field.

We integrated the basal shear stress ($\sim 3.42 \times 10^{11}$ N) for the frontal sticky spots in 2008 (where the $T_{aob} > 0.01$ MPa shown in Fig. S3). We have clarified this in Line 232.

We have tried some sensitivity tests to different ice front positions and ice front ocean-pressure boundary conditions in the companion paper (Zhao et al., companion paper). Those experiments have a similar effect to modifying basal shear stress near the ice front. The results show that those changes didn't impact on the velocity very far upstream. So this unpinning on its own is unlikely to have caused the speed up, but it could be a trigger for basal feedbacks to kick in.

line 528: Schaëfer is not spelled correctly

Modified.

Caption Fig. 1: inset (c) should be located in (b) and in (c) the front position in 2008 and 2016 should be added to visualise a potential ice-shelf?

Modified and added.

Fig. 3: the grounding line in 2014 seems to have a different form than the one of Friedl et al. (2017) in their Fig. 6?

Fig. 3 is generated with Paraview. To add the grounding line of 2014 in Paraview, we have to generate the mesh with the grounding line of 2014. A typical element size in this region is ~ 200 - 300 m. The only difference between the grounding line in Fig. 3 and the original shapefile is mapping it to nodes on the Elmer mesh, therefore the differences are always less than one element size. The mesh size and the refinement affected the location of grounding line. So the difference is never more than 300 m (an element's width), and it would not affect the analysis in this paper.

References

Dunse, T., Schuler, T. V., Hagen, J. O., and Reijmer, C. H.: Seasonal speed-up of two outlet glaciers of Austfonna, Svalbard, inferred from continuous GPS measurements, *The Cryosphere*, 6, 453-466, 2012.

Rignot, E., Casassa, G., Gogineni, S., Kanagaratnam, P., Krabill, W., Pritchard, H., Rivera, A., Thomas, R., Turner, J., and Vaughan, D.: Recent ice loss from the Fleming and other glaciers, Wordie Bay, West Antarctic Peninsula, *Geophysical Research Letters*, 32, 2005.

Schoof, C.: Ice-sheet acceleration driven by melt supply variability, *Nature*, 468, 803-806, 2010.

Shreve, R.: Movement of water in glaciers, *Journal of Glaciology*, 11, 205-214, 1972.

Wendt, J., Rivera, A., Wendt, A., Bown, F., Zamora, R., Casassa, G., and Bravo, C.: Recent ice-surface-elevation changes of Fleming Glacier in response to the removal of the Wordie Ice Shelf, Antarctic Peninsula, *Annals of Glaciology*, 51, 97-102, 2010.

Zhao, C., Gladstone, R., Zwinger, T., Warner, R., and King, M. A.: Basal friction of Fleming Glacier, Antarctica, Part A: sensitivity of inversion to temperature and bedrock uncertainty, *The Cryosphere*, companion paper. companion paper.

Response to the Interactive comment on

“Basal drag of Fleming Glacier, Antarctica, Part B: implications of evolution from 2008 to 2015”

by Chen Zhao et al.

Anonymous Referee #3

Received and published: 12 Mar 2018

We are grateful to Reviewer 3 for the positive and constructive suggestions to improve our paper. We have addressed the comments below. The line numbers in the responses are based on the revised manuscript without change track.

Please note that Mathieu Morlighem created the ice thickness data for the Fleming Glacier system using the mass conservation method, which is very important for most experiments done in this study. We do value his contribution to this paper, so we add him as the co-author in the revised text.

In the revised companion paper (Zhao et al., companion paper), we implemented a new sensitivity test to the enhancement factor (E). It reveals that the optimal value of $E = 1.0$ should be chosen as the enhancement factor in the CONTROL experiment. Accordingly, we re-ran all the simulations in this study with $E = 1.0$, and the high basal shear stress band near the ice front in 2008 has decreased into high basal shear spots, which are suspected of being artefacts of the inversion process and are discussed below. We modified the text and figures accordingly. All other result and interpretations are not qualitatively changed from the original manuscript.

GENERAL COMMENTS

Main question in the abstract: Is the observed acceleration of the flow and thinning of the glacier due to increased ocean warming and/or marine ice sheet instability?

Method: Infer basal shear stress from observations and calculate a steady state temperature field using a Stokes ice sheet model for 2008 and 2015.

Results: Reduction in magnitude and increase in area of low basal shear stress near the 1996 grounding line and reduction in height above flotation between 2008 and 2015 suggest the grounding line has retreated for Fleming Glacier, southern branch of Fleming Glacier and Prospect Glacier.

A band of higher basal shear stress parallel to the 1996 grounding line at 2008 suggests that Fleming Glacier was still grounded at that time. Subglacial water may be generated from high basal frictional heating upstream of Fleming Glacier. Frictional heating has increased between 2008 and 2015 over a rise between two deep bedrock basins.

As mentioned above, in the revised companion paper, we implemented a new sensitivity test to enhancement factor (E). It reveals that the optimal value of 1.0 should be chosen as the enhancement factor in the CONTROL experiment. So we redid all the simulations in this study with $E=1.0$, and the high basal shear stress band near the ice front changed into high basal shear spots in 2008, which are suspected to be artefacts. We did not rule out, however, the possibility that the ice front was still grounded on some pinning points. We discuss this point in the first sentence of the Discussion section (Line 359-361).

Comments: I don't think the main question can be answered from instantaneous time slices of the ice flow. The authors need to do forward experiments with various ocean forcing such as different basal melt rates or vertical melting at the calving front. Alternatively, the authors need to pose a different question. The band of high basal shear stress may not be physical realistic. The model error reported in their companion paper is relatively high in this area.

Clearly forward modelling of the Fleming system to study the recent ungrounding transition is a natural next step. That would, as the reviewer acknowledges, require an extensive exploration of forcing influences. In the present work we have clearly shown the differences in basal shear stress distributions for the Fleming system between 2008 and 2015, reflecting different surface elevations and the recent acceleration in ice flow. This has provided insights into the recent ungrounding – and suggested possible feedback processes that may have contributed to the recent changes. We consider this scope has provided sufficient worthwhile material for the present paper. Experiments with future coupled ice sheet-ocean models would also be valuable. We have mentioned this in the Conclusion section (Line 535-537).

In the modified companion paper (Zhao et al., companion paper), the misfit between the simulated and observed surface velocity at the ice front of the FG is very small. The difference between the relaxed and observed surface is < 15 m after three cycles in the CONTROL experiment. It means the modified model with the enhancement factor of 1.0 models the ice front well.

Interesting idea: The authors propose that basal water generated from high basal frictional heating upstream draining towards the front, triggered grounding line retreat of Fleming Glacier. This mechanism is an alternative to the usual ocean forcing explanation. Mass loss could significantly increase, due to marine instability, if the grounding line retreated over a bedrock rise into the second deeper basin. The highest frictional basal heating in 2015 is located over the rise, which may be a potential trigger for the grounding line retreat.

Manuscript in general: The font is too small and the text is not double spaced, which made reviewing the paper tricky. Picking out the references was particularly difficult give the font size and text spacing. Some of figures are too small.

Apologies if the manuscript was not in the format the Reviewer expected. We are happy to comply with whatever formatting requests are made by the Copernicus staff in this regard.

SPECIFIC COMMENTS

Ocean forcing: It seems reasonable to suggest that increased melting at the vertical face of the front of FGL due to incursions of CDW may have affected the pressure boundary condition at the front sufficiently to remove the high band of basal shear stress. However, I don't think your results shed any new light on what has been suggested in the other references you use about ocean basal melting. Forward time-dependent modelling experiments are needed to test these theories and here's an example for Larsen B of how you can extend the work you have done for this paper. Vieli et al 2007 Causes of pre-collapse changes of the Larsen B ice shelf: Numerical modelling and assimilation of satellite observations. Earth and Planetary Science Letters. <https://doi.org/10.1016/j.epsl.2007.04.050>

As we mentioned above, we agree that transient experiments will be valuable, but are beyond the scope of the current study. We aim to carry out both transient ice dynamic

simulations and coupled ice-ocean simulations, and hope we will be able to bring such studies to fruition over the coming years.

Grounding line retreat: The results for 2015 of low basal shear stress and low height above buoyancy confirm the findings of Friedl et al 2017 that Fleming Glacier's grounding line has retreated. The results for PGL are different to FGL: Driving stress appears to be much higher for PGL in 2015.

The revised ratio of driving stress τ_{d2015} to τ_{d2008} (Fig. 3f) shows that the driving stress of PG in 2015 was much lower (not higher) than 2008. We have clarified that the cases for the southern FG and PG are different from the main branch of FG. We did not account in the model for the remaining ice shelf for those two glaciers because we do not have the ice thickness data for the ice shelf. We modified our analysis on those two glaciers (Line 275-284). We think the northern section of the southern FG has been ~ 2 km behind the 1996 grounding line position based on the ice front position shown in Fig. 1c. However, it is hard to decide whether the southern section of the southern FG or the PGL have also retreated from 2008 (Fig. 3a) to 2015 (Fig. 3b), since we did not account for the normal stress of the remaining small ice shelf at the front of the southern FG (Fig. 1c) in the inverse modelling. Note that the hypsometric model used generate the DEM in 2015 is based on the observed elevation change rates (Zhao et al., 2017). However, the observations are mainly focused in the FG region (Fig. 2a), so the DEM2015 of PG could be an artefact. That might explain why the driving stress was lower in 2015 (Fig. 3f).

Temperature homologous near the 1996 grounding line (for PG) appears much lower in 2015 suggesting that the glacier may have become frozen to the bed?

Note that we have replaced the term “temperature homologous” with “temperature relative to pressure melting point” in the entire text.

The temperature near the ice front/grounding line of PG is indeed colder in the 2015 steady-state calculation. The main difference in the modeled temperature between 2008 and 2015 is due to a reduction in friction heat. This is in turn due to reduced basal shear stress, which occurs in the inversion as a result of the reduced driving stress compared to 2008. This may be due to the lack of observational hypsometric data – the imposed surface lowering (which causes the driving stress reduction) in 2015 is based mainly on data from FG.

A contributing factor could be the steady state temperature assumption, which is almost certainly worse for 2015 than it is for 2008, because the recent acceleration means that the glacier is further from steady state in 2015 than in 2008.

Also, the current modelling approach does not represent the capacity of the subglacial hydrologic system to redistribute heat at the bed. In reality the flow of basal melt water from upstream to downstream will bring more latent heat to the base of the ice sheet near the grounding line.

Is the band of high basal stress at the front of FGL physically realistic? The authors attempt to address this question in the paragraph beginning on line 209. Part A shows that the misfit between the modelled and observed speed is high, where the modeled speed is too fast, and the surface slope is also higher here than over the region of low shear stress. The driving stress is not obviously high given the relatively high surface slope. What concerns me is your model appears unable to model the front.

As mentioned above, the revised companion paper of this study (Zhao et al., companion paper) shows high basal stress spots rather than a band (as previously) at the front of FGL. This may, as the reviewer suggests, be an artefact, owing to various uncertainties. We also do not rule out the possibility that the ice front was still grounded on some pinning points. We clarified this in Line 359-361.

In the revised companion paper (Zhao et al., companion paper), the misfit between the simulated and observed surface velocity at the ice front of the FG has been very small. The difference between the relaxed and observed surface is < 15 m after three cycles in the CONTROL experiment. It means the modified model with the enhancement factor of 1.0 models the ice front in 2008 better now than in the version of the companion paper to which the reviewer refers. It is also worth noting that our 2015 simulations have not had any difficulties modelling the ice front. This suggests that the problem is in the boundary conditions rather than the model itself, which was the motivation for the ice front position and pressure sensitivity experiments in the companion paper. These experiments indicated that the inversion is only sensitive to such ice front uncertainties within a short distance of the front.

What about rheology of the ice near the front? Perhaps the standard A is not appropriate here. Part A shows a large vertical shear at the front where the basal speed is much smaller than the surface. Is the ice stiffer at the front? Vieli et al 2006 Numerical modelling and data assimilation of the Larsen B ice shelf, Antarctic Peninsula, *Phil. Trans. R. Soc. A*, 364, 1815–1839, doi:10.1098/rsta.2006.1800 solved the inversion problem for effective viscosity. Modelling a front is difficult!

Based on the sensitivity test to various values of enhancement factors (0.5, 1.0, 2.0, 4.0) in the revised companion paper, we found that the value of 1.0 is the optimal value for the overall Fleming system. Various studies of anisotropic ice properties and enhancement factors (e.g. Graham et al. (2018); Ma et al. (2010)) suggest that ice near the ice front could well be stiffer than ice deforming under simple shear near the bedrock in the interior of the ice sheet, however, we have only a uniform enhancement factor E as a control parameter in the present study.

About solving the inversion problem for effective viscosity: it is simple to invert for ice rheology in an ice shelf model, as suggested. Here for the grounded glacier - certainly largely grounded in 2008 - the velocity mismatch can be addressed by adjusting the ice stiffness and the basal drag. Simultaneous inversions for stiffness and basal friction coefficient are possible but beyond the model tools we have available.

What about the direction of the flow? Is there a difference in the modelled flow direction and the observed direction? Is there a change in flow direction between 2008 and 2015 as the ice moves over the sticky band and becomes ungrounded. Also, could ice mélange at the front FGL affect the boundary condition?

The inversion scheme we used (following Gagliardini et al. (2013)) only compares the mismatch in modelled and observed speeds, not directions. To a simple visual inspection the velocity directions in 2008 and 2015 are very similar if not identical. A direct overlay of streamlines may allow minor deviations to be identified, but we have not identified an urgent need to such analysis to be carried out.

About the ice mélange at the ice front, we explored the effect of an extra normal force at the ice front (to simulate the potential effect of ice mélange) in the ice front boundary condition experiments of the revised companion paper (Zhao et al.,

companion paper). We calculated that ice mélange back force ($\sim 1.1 \times 10^7 \text{ N m}^{-1}$) used to prevent the rotation of iceberg at the calving front (Krug et al., 2015) could account for the equivalent of up to $\sim 2.3 \text{ m}$ sea level in terms of ice front boundary condition, which was included in the experiments with different sea levels.

Basal frictional heating and subglacial water: Could the region of low basal shear stress near the front simply be due to subglacial water from upstream pooling in the bedrock basin, e.g. FGL in 2008? Could the region be partially grounded?

The reviewer may be right. Figs. 4d and 4e show that there is a plateau in the hydraulic potential in the downstream basin. In 2015, we think the downstream basin is mainly ungrounded, based on the inferred basal shear stress (Fig. 3b) and the height above buoyancy (Fig. 5b). There could therefore be some pooling of water in 2008, and it could be partially grounded, but a big cavity is not possible given the geometry. Our inferred basal shear stress (Fig. 3a) and the height above buoyancy (Fig. 5a) show that the downstream basin of the FG in 2008 should still remain grounded.

In the discussion section, we have clarified the issue of basal hydrology, and the potential of water to pool at certain locations (Line 471-476). “The plateaus in hydraulic potential in both downstream and upstream basins of the FG suggest the possibility that basal water may accumulate in those regions, or at least show a low throughput. The downstream plateau appears to be fed by a large frictional heat source over the ridge between the downstream and upstream basins in addition to flow from further inland, while the upstream plateau appears to be fed by an extensive upstream region of basal melting.”

The temperature homologous is high, which prevents the water from refreezing. I think the role of subglacial water could be explained more in the literature review. Paragraph beginning 85:

We modified this sentence (Line 98-102) into “A positive feedback between basal sliding and basal water pressure (through friction heating) upstream of the grounding line could be another possible factor in the glacier acceleration and grounding line retreat (Bartholomaus et al., 2008; Iken and Bindshadler, 1986; Schoof, 2010). The possibility of such a feedback, is not ruled out by Friedl et al. (2018), and is discussed further in Sect. 4.2 and Sect. 5.”

You don't explain what the feedback mechanism is. As I understand it, Schoof (2010) talks about the importance of variability of basal water on flow dynamics, with flow accelerating due to a short-lived increase in basal water, but then the flow slows if the basal water stays high. Is that happening here? You have high basal frictional melting in 2015, which you say is speeding up the flow, but figure 3 in Friedl et al 2017 shows that the ice speed of FGL decreased between 2011-2015.

We have added a description about the positive feedback in Sect. 4.2 (Line 288-301). As we have clarified, “Since the reduction of effective pressure is the key process to enhance sliding, this positive feedback is dependent on a positive feedback of melt water generation to water pressure. This dependence can break down when there is sufficient basal water to generate efficient drainage channels (Schoof, 2010). However, such efficient channelization in the subglacial hydrologic system is typically associated with seasonal surface meltwater pulses reaching the bed (Dunse et al., 2012), a process that is not expected to occur for Fleming Glacier (Rignot et al., 2005).”

In the published Fig. 2 and Fig. 3 in Friedl et al. (2018), it shows that the ice speed of

FGL remains stable with a very small median velocity increase (0.06 m d^{-1}) from 2011 to 2016, not decrease. Besides, the speed up in 2015 is relative to the surface velocity 2008. We do not dispute the observed acceleration phrase occurred in Mar 2010-early 2011 found by Friedl et al. (2018).

Figure 1 shows that the ice front for PGL and sFGL has calved between 2008 and 2016, with some advance in the southern part of the bay. Could calving event(s) explain the speed up and lowering of the surface of the streams? Also, from figure 1, PGL has an ice shelf, are you applying the normal stress, hydrostatic pressure boundary condition at the 1996 grounding line or at the real front?

Yes, we agree with the reviewer. The calving events may explain the speed up of the PGL and sFGL. The surface lowering for those two regions have not been confirmed by Zhao et al. (2017) owing to the lack of elevation observations. But inversions will not give a clear answer to this - transient experiments would be needed. A good experiment to do would be to carry out an inversion with an advanced ice shelf, then two transient experiments: one simply carrying on from the inversion, and one in which the shelf is removed. The difference between these transient experiments would be informative as to the impact of calving on flow speeds and surface lowering. Such experiments would be interesting, and we hope to have a chance to carry out such experiments, but that they are beyond the scope of the current study.

As we responded above, we did not account for the remaining ice shelf for those two glaciers because we don't have the ice thickness data for the ice shelf. We modified our analysis on those two glaciers (Line 275-284).

Bedrock plots: The way the bedrock is plotted is a bit inconsistent and unclear. Figure 1 c clearly shows where bedrock is above or below sea level with a white colour band around 0, but the bed elevation colour in figure 4 c and d looks like most of the bedrock is either below or at sea level and figure 2 b is too small. It would be useful to see where the retrograde and prograde slope are.

We modified Fig. 2b to have the same color scale as Fig. 1c. For Figs. 4d, 4e (original Figs. 4c, 4d), we think it is better to use a different color scale to show the retrograde and prograde slopes. Here we plotted the bed elevation with meters above sea level. The regions of retrograde slope are now easily identifiable by eye in Figs. 4d, 4e, but plotting these regions is not trivial to automate.

Figures 3,4,5: useful to have a third column of figures showing the difference between the first two columns.

Thanks for the suggestion. For Fig. 3, we computed the ratio of τ_{b2015} over τ_{b2008} (Fig. 3e), and also the ratio of τ_{d2015} over τ_{d2008} (Fig. 3f) to represent the difference. For Fig. 4 and Fig. 5, we plotted the difference between 2015 and 2008 (2015 minus 2008).

Line 158: The Linear sliding law is fine for the inverse problem because τ remains unchanged, if a higher power of u is used, only the coefficient C would change. However, for a forward run a linear law may be inappropriate.

We agree. There is, however, no transient simulation involved in this study except in the brief surface relaxation step. For the steady-state temperature simulation, the Stokes solver is turned off and the velocity field is fixed. It means that the basal shear stress is fixed for the temperature simulation. So we don't think it is inappropriate to run our simulations with a linear sliding law. For transient simulations we intend to

convert C to a distribution that, for whatever sliding law is used, gives the same initial basal shear stress distribution as we obtain from the inversion.

Please define basal frictional heating in your method section.

We have added the equation as Eq. 4.

Figure 2 is too small and/or too detailed. A difference plot of surface elevation may be more informative.

Modified.

Figure 3: The patterns in c and d seem to be influenced by the computational mesh. Have you investigated mesh resolution by halving or doubling element sizes?

The sensitivity tests to horizontal (1 km, 500 m, 250 m, 125 m) and vertical (10 layers and 20 layers) mesh resolution have been carried out in our companion paper (Zhao et al., companion paper). The resolution of 250 m is fine enough for inverse modeling in this study. Also note that the features in Figs. 3c and 3d (now 3d and 3e after revision) are much coarser than the element size – these features are resolved.

The ratio of basal shear stress to driving stress: I'm not sure what figure 3 c and d are showing or why a low value means the ice may be close to flotation.

A balance between τ_b and τ_d is indicative of a heavily grounded regime – longitudinal stresses are low and the dominant force balance is between the gravitational driving stress and the basal resistance. This assumption is similar to the assumption behind the derivation of the “shallow ice approximation”, which has been used extensively for long time scale grounded ice sheet simulations. The opposite assumption, the “shallow shelf approximation” (SSA), is a balance between membrane stresses and driving stresses, with basal shear stress being vanishingly small. This is the typical stress balance in a floating ice shelf. So a low value of this stress ratio means we're closer to the SSA regime, i.e. an ice shelf regime.

Figure 4 c and d: Would showing the potential gradient be more informative? I can't see labels on the contours of hydraulic potential.

We agree that the direction of flow is important, but a vector plot of the potential gradient would be rather complicated. The direction of flow, perpendicular to the contours of hydraulic potential, should be clear from the figure, with very few local minima. The magnitude of the gradient can be clearly seen in the proximity of contours of hydraulic potential. Where the contours are close together is where the gradient is steep. Labelling the contours increases the complexity of the plot, and in fact the main purpose of the whole calculation is to demonstrate the pattern of basal water flow rather than to estimate specific values. We have actually tried several options for how to present this Figure, including a scalar field plot of hydraulic potential gradient, but we are still confident that our current combination of bedrock and hydraulic potential contours gives the clearest picture of the pattern of basal water flow.

Figure 5: Height above buoyancy appears to be negative south of the main stream of FGL (and south of PGL). Is the bedrock above sea level there?

No. Note that the height above buoyancy is never negative where the bedrock is above sea level (Eq. 7). Negative values simply indicate that the estimated thickness of ice present should be afloat, given the bedrock is so far below sea level. As we discuss in the text, uncertainties in ice thickness and bedrock data affect the

calculation of the height above buoyancy. The bedrock of the southern branch of the FG is below sea level while a little part of the south PG is below sea level. To clarify the question of bedrock values, we modified the Fig. 1c to show the bedrock below sea level only.

Discussion section: Is a maximum melt rate of 1 m/a enough to generate a plume of high enough velocity to entrain incursions of CDW to enhance basal melting beneath the floating ice? You can calculate the flux of subglacial water for each year by Taobub/Lii x area that feeds the grounding line based on the hydraulic potential (or its gradient).

To address this question, we need a 3D ocean model. The buoyant plume is a function of many things, of which subglacial outflow is only one. We'd need to know a lot about the CDW pathways in the area, whether CDW is coming into contact with the grounding line, volume fluxes, heat fluxes, the regional oceanography. So yes, calculating total subglacial outflow is relatively straightforward but simulating the local ocean circulation and plume behavior is way beyond the scope of the current study, and the subglacial outflow is a fairly meaningless number without this oceanographic context.

Line 395: Could you explain the positive feedbacks. Estimating the time scale for the ice to unground from the rise between basins leaving the ice stream vulnerable of marine instability in the upstream basin is good, but I'm not sure you can say height above buoyancy is a measure of potential mass loss.

We have added a description of the positive feedbacks in Sect. 4.2 (Line 288-301).

We did not say the height above buoyancy could indicate potential mass loss. It indicates potential vulnerability. If Z^* is close to zero, then the system is close to ungrounding, and would only require a small perturbation to unground. The height above buoyancy would be relevant to sea level rise in a grounding line retreat situation. The accelerated ice flux across the grounding line is probably a separate issue.

TECHNICAL CORRECTIONS

Line 46: abbreviation 'GL' is not defined.

We don't use the abbreviation "GL" for the "grounding line", so we modified "GL" into "grounding line".

Line 88: Not sure the sentence is helpful. Might be better to delete it.

Deleted.

Section 2.2: Is Hmc part of a dataset from Morlighem or have you combined two dataset yourself?

Morlighem has been added as the co-author of both the companion paper and this paper. He generated Hmc for the companion paper (Zhao et al., companion paper). Hmc includes three regions: for fast flowing region, he computed the ice thickness data for fast-flowing regions using the Ice Sheet System Model's mass conservation method (Morlighem et al., 2011; Morlighem et al., 2013), based on ice thickness measurements from the Center for Remote Sensing of Ice Sheets (CReSIS), using ice surface velocities in 2008 from Rignot et al. (2011b), surface accumulation from RACMO 2.3 (van Wessem et al., 2016) and 2002-2008 ice thinning rates from Zhao et al. (2017); for slow flowing region, he adopted data from bedmap2; for the

transition region, he smoothed the data. It has been clearly described in the revised companion paper.

Lines 132, 151: Part 1 or Part A

Thanks for pointing this out. It should be “Part A”.

Line 145: Is the basal frictional heating calculated from output from the inverse problem and used as an input into the heat equation?

Yes. The basal frictional heating shown in Figs. 4a and 4b, is calculated directly from the output of the inversion process – as the product of inferred basal shear stress and basal velocity – see Eq. 4, which we added at the request of this reviewer (see above). The basal frictional heating is an integral part of the steady state temperature simulation. That calculation does use the velocities and friction coefficients from the inversion, so the frictional heating from the inversion is indeed included.

184: I don’t think N needs a numbered equation because it isn’t used.

We would like to make this change if the Copernicus proofreaders request it.

Line 222: northern and eastern. It might be helpful to add an arrow indicating North on one of the figures.

Added.

Lines 367, 420: Friedl et al 2017 gave an estimated grounding line for 2014.

Thanks for pointing this out. We have modified the relevant text to grounding line in 2014.

Figure 1: sFGL is not marked on the figure.

Added.

Figure 3: It is difficult to work out where the plotted regions exist in relation to figures 1 and 2. Orientation is given in figure 5 but would be more useful on figure 3.

Thanks for the reviewer’s suggestions. We have added an inset map in Fig. 3a to show the plotted region. We also added the north direction in Fig. 3a.

Figure 3: Cannot see cyan contour on printed paper.

We modified the color of all the velocity contours into white color.

Figure 4: I can’t distinguish between red and magenta contours.

We changed both colors in Fig. 3a, 3b, 3d, 3e.

Line 529: Case is wrong for Schafer.

Modified.

References

Bartholomaus, T. C., Anderson, R. S., and Anderson, S. P.: Response of glacier basal motion to transient water storage, *Nature Geosci*, 1, 33-37, 2008.

Dunse, T., Schuler, T. V., Hagen, J. O., and Reijmer, C. H.: Seasonal speed-up of two outlet glaciers of Austfonna, Svalbard, inferred from continuous GPS measurements, *The Cryosphere*, 6, 453-466, 2012.

Friedl, P., Seehaus, T. C., Wendt, A., Braun, M. H., and Höppner, K.: Recent dynamic changes on Fleming Glacier after the disintegration of Wordie Ice Shelf, Antarctic Peninsula, *The Cryosphere*, 12, 1-19, 2018.

Gagliardini, O., Zwinger, T., Gillet-Chaulet, F., Durand, G., Favier, L., de Fleurian, B., Greve, R., Malinen, M., Martín, C., Råback, P., Ruokolainen, J., Sacchettini, M., Schäfer, M., Seddik, H., and Thies, J.: Capabilities and performance of Elmer/Ice, a new-generation ice sheet model, *Geosci. Model Dev.*, 6, 1299-1318, 2013.

Graham, F. S., Morlighem, M., Warner, R. C., and Treverrow, A.: Implementing an empirical scalar constitutive relation for ice with flow-induced polycrystalline anisotropy in large-scale ice sheet models, *The Cryosphere*, 12, 1047-1067, 2018.

Iken, A. and Bindschadler, R. A.: Combined measurements of Subglacial Water Pressure and Surface Velocity of Findelengletscher, Switzerland: Conclusions about Drainage System and Sliding Mechanism, *Journal of Glaciology*, 32, 101-119, 1986.

Krug, J., Durand, G., Gagliardini, O., and Weiss, J.: Modelling the impact of submarine frontal melting and ice mélange on glacier dynamics, *The Cryosphere*, 9, 989-1003, 2015.

Ma, Y., Gagliardini, O., Ritz, C., Gillet-Chaulet, F., Durand, G., and Montagnat, M.: Enhancement factors for grounded ice and ice shelves inferred from an anisotropic ice-flow model, *Journal of Glaciology*, 56, 805-812, 2010.

Morlighem, M., Rignot, E., Seroussi, H., Larour, E., Ben Dhia, H., and Aubry, D.: A mass conservation approach for mapping glacier ice thickness, *Geophysical Research Letters*, 38, n/a-n/a, 2011.

Morlighem, M., Seroussi, H., Larour, E., and Rignot, E.: Inversion of basal friction in Antarctica using exact and incomplete adjoints of a higher-order model, *Journal of Geophysical Research: Earth Surface*, 118, 1746-1753, 2013.

Rignot, E., Casassa, G., Gogineni, S., Kanagaratnam, P., Krabill, W., Pritchard, H., Rivera, A., Thomas, R., Turner, J., and Vaughan, D.: Recent ice loss from the Fleming and other glaciers, Wordie Bay, West Antarctic Peninsula, *Geophysical Research Letters*, 32, 2005.

Rignot, E., Mouginot, J., and Scheuchl, B.: Ice Flow of the Antarctic Ice Sheet, *Science*, 333, 1427-1430, 2011b.

Schoof, C.: Ice-sheet acceleration driven by melt supply variability, *Nature*, 468, 803-806, 2010.

van Wessem, J. M., Ligtenberg, S. R. M., Reijmer, C. H., van de Berg, W. J., van den Broeke, M. R., Barrand, N. E., Thomas, E. R., Turner, J., Wuite, J., Scambos, T. A., and van Meijgaard, E.: The modelled surface mass balance of the Antarctic Peninsula at 5.5 km horizontal resolution, *The Cryosphere*, 10, 271-285, 2016.

Zhao, C., Gladstone, R., Zwinger, T., Warner, R., and King, M. A.: Basal friction of Fleming Glacier, Antarctica, Part A: sensitivity of inversion to temperature and bedrock uncertainty, *The Cryosphere*, companion paper. companion paper.

Zhao, C., King, M. A., Watson, C. S., Barletta, V. R., Bordoni, A., Dell, M., and Whitehouse, P. L.: Rapid ice unloading in the Fleming Glacier region, southern Antarctic Peninsula, and its effect on bedrock uplift rates, *Earth and Planetary Science Letters*, 473, 164-176, 2017.

Basal ~~drag friction~~ of Fleming Glacier, Antarctica, Part B: ~~implications of~~ evolution from 2008 to 2015

Chen Zhao^{1,3}, Rupert M. Gladstone², Roland C. Warner³, Matt A. King¹, Thomas Zwinger⁴, [Mathieu Morlighem⁵](#)

5 ¹ [School of Technology, Environments and Design School of Land and Food](#),
University of Tasmania, Hobart, Australia

² Arctic Centre, University of Lapland, Rovaniemi, Finland

³ Antarctic Climate and Ecosystems Cooperative Research Centre, University of Tasmania, Hobart, Australia

10 ⁴ CSC-IT Center for Science Ltd., Espoo, Finland

⁵ [Department of Earth System Science, University of California Irvine, CA 92697-3100, USA](#)

Abstract

15 The Wordie Ice Shelf-Fleming Glacier system in the southern Antarctic Peninsula has experienced a long-term retreat and disintegration of its ice shelf in the past 50 years. ~~Increases in Upstream the~~ glacier ~~acceleration-velocity~~ and dynamic thinning have been observed over the past two decades, especially after 2008 when only a ~~little constraining~~ small ice shelf remained at the Fleming Glacier front. It is important to know whether the substantial ~~further~~ speed up and ~~greater~~ surface draw-down of the glacier since 2008 is a
20 direct response to ~~increasing~~ ocean forcing, or driven by ~~the~~ feedbacks within ~~an unstable marine-based~~ the grounded ~~marine-based~~ glacier system, or both. ~~Recent observational studies have suggested the 2008-2015 velocity change was due to the ungrounding of the Fleming Glacier front.~~ To explore the mechanisms underlying the ~~recent~~ changes, we use a ~~full-Stokes ice sheet (full stress)~~ model to simulate the basal shear stress ~~distribution~~ of the Fleming system in 2008 and 2015. ~~Recent observational studies have suggested the 2008-2015 velocity change was due to the ungrounding of the Fleming Glacier front.~~ ~~This study is part of the first high resolution modelling campaign of this system. Our modelling shows that the fast flowing region of the Fleming Glacier shows a very low basal shear stress in 2008 but with a band of higher basal shear stress along the ice front. It indicates that the ungrounding process might have not started in 2008, which is consistent with the height above buoyancy calculation in 2008.~~ Comparison of ~~our~~ inversions for basal shear stresses for 2008 and 2015 suggests the migration of the grounding line ~~by~~ ~9 km upstream ~~by 2015 from from~~ the ~~2008 ice front/grounding line positions, which virtually coincided with the~~ 1996 ~~grounding line position, a~~ ~~This shift migration which~~ is consistent with the change in floating area deduced
35 from the ~~calculated~~ height above buoyancy in 2015. ~~The southern branch of the Fleming Glacier and the Prospect Glacier apparently have retreated by ~1-3 km from 2008 to 2015.~~ The retrograde ~~submarine~~ bed underneath ~~the lowest part of~~ the Fleming Glacier ~~may have~~ promoted ~~retreat migration~~ of the grounding line, ~~Grounding line retreat which we suggest may also be triggered enhanced by a feedback mechanism upstream of the grounding line by which increased basal lubrication due to increasing subglacial drainage as a response to the increased basal water supply through greater frictional heating enhances sliding and thinning at the ice bedrock interface further upstream in the fast flowing region.~~ Improved knowledge of bed topography near the grounding line and further transient simulations ~~with oceanic forcing is are~~ required to ~~accurately~~ predict the future ~~grounding line~~ movement of
45 the Fleming Glacier system ~~grounding line precisely~~ and ~~better subsequently~~ understand ~~better its~~ the ice dynamics and ~~the its~~ future contribution to sea level.

1 Introduction

In the past few decades, glaciers in West Antarctica and the Antarctic Peninsula (AP) have experienced rapid regional atmospheric and oceanic warming, leading to significant retreat and disintegration of ice shelves and rapid acceleration of mass discharge and dynamic thinning of their feeding glaciers (Cook et al., 2016; Gardner et al., 2018; Wouters et al., 2015). Most of the West Antarctic Ice Sheet and the glaciated margins of the AP (Fig. 1a) rest on a bed below sea level sloping down towards the ice sheet interior, and the grounding lines of outlet glaciers located on such reverse bed slopes may be vulnerable to rapid retreat depending on the bedrock and ice shelf geometry (e.g., Gudmundsson (2013); Gudmundsson et al. (2012); Schoof (2007)). Once perturbed past a critical threshold, such as grounding-line retreat over a bedrock hump into a region of retrograde slope, the ~~GL~~ grounding line will may continue to retreat inward until the next stable state without any additional external forcing (e.g., Mercer (1978); Thomas and Bentley (1978); Weertman (1974)). This marine ice sheet instability has been invoked to explain the recent widespread and rapid grounding line retreat of glaciers in the Amundsen Sea sector, ~~possibly driven~~ likely triggered by increased basal melting reducing the buttressing influence of ice shelves (Rignot et al., 2014). Rapid grounding line retreat and accelerated flow in these unstable systems leads to significant increases in ice ~~discharge flux~~ and increased contribution from these marine ice sheets to sea-level rise.

The former Wordie Ice Shelf (WIS; Fig. 1b) in the western coast of AP started its initial recession in 1960s with a substantial break-up occurring around 1989, followed by continuous steady retreat (Cook and Vaughan, 2010; Vaughan and Doake, 1996; Wendt et al., 2010; Zhao et al., 2017). The former ice shelf is fed by three tributaries as shown in Fig. 1b. The Fleming Glacier (FGL; Fig. 1b), as the main tributary glacier, splits into two branches: the main branch to the north and the southern branch (hereafter “southern FGL”). The floating part in front of the main FGL ~~nearly~~ almost entirely disappeared sometime between 1997 and 2000 (Fig. 1b), and the ice front position in Apr 2008 (dark blue line in Figs. 1b and 1c, Wendt et al. (2010)) almost coincides with the latest known grounding line position in 1996 (Rignot et al., 2011a). The main branch of the FGL has thinned at a rate of -6.25 ± 0.20 m yr⁻¹ near the front from 2008 to 2015, ~~nearly more than~~ twice the thinning rate during 2002-2008 (-2.77 ± 0.89 m yr⁻¹) (Zhao et al., 2017). This is consistent with the recent findings that the largest velocity changes across the whole Antarctic Ice Sheet over 2008-2015 occurred at FGL (500 m yr⁻¹ increase close to the 1996 grounding line) (Walker and Gardner, 2017). Time series of surface velocities along the centerline of the FG (extending ~16 km upstream from the 1996 grounding line) (Friedl et al., 2018) indicate that two rapid acceleration phases occurred: in Jan-Apr 2008 and from Mar 2010 to early 2011, followed by a relatively stable period from 2011 to 2016. During the first acceleration phase in Jan-Apr 2008, the front of the FG retreated behind the 1996 grounding line position for the first time (Friedl et al., 2018).

As a marine-type glacier system residing on a retrograde bed with bedrock elevation as much as ~800 m below sea level (Fig. 1c), ~~and the Fleming system is~~ hence potentially vulnerable to marine ice sheet instability (Mercer, 1978; Thomas and Bentley, 1978; Weertman, 1974), ~~the~~ The acceleration and greater dynamic thinning of the FG over 2008-2015 ~~may~~ indicatesuggests the possible onset of unstable rapid grounding line retreat (Walker and Gardner, 2017; Zhao et al., 2017), which has been confirmed by Friedl et al. (2018). The speedup of the FGL before 2008 was originally assumed to be a continuing direct response to the ~~loss-collapse~~ of ~~buttressing due to the~~ Wordie ice shelf ~~Shelf collapse~~ (Rignot et al., 2005; Wendt et al., 2010). Recent studies have suggested that the recent further glacier speedup and grounding line retreat could be a direct response to oceanic forcing (Friedl et al., 2018; Walker and Gardner, 2017). An alternative hypothesis is that the recent changes arise from feedbacks in the dynamics of the evolving glacier, possibly involving the subglacial hydrology. The examination of changes in basal shear stress distributions between 2008 and 2015 in this modelling study provides a first step in exploring possible feedback hypotheses.

100 ~~None of the past studies have modelled the glacier system and hence these hypotheses are untested. In this paper w~~We explore the potential for these hypotheses in Sect. 5.

By analyzing the detailed history of surface velocities, ~~and~~ rates of elevation change, ~~and ice front positions~~ from 1994 to 2016, Friedl et al. (2018) ~~showed-suggested~~ that the initial ungrounding of the FGL from the 1996 grounding line position (Rignot et al., 2011a) occurred ~~during the first acceleration phase between~~ Jan and Apr 2008 and ~~expanded~~ further ~~expanded~~ upstream by ~6-9 km ~~from 2011 to~~ by 2014, which explained the speedup ~~and thinning~~ of the FGL since 2008, ~~and~~. They speculated this was mainly ~~the result of unpinning~~ caused by ~~the~~ increased basal melting ~~at the grounding line~~ due to the upwelling of ~~the~~ warm Circumpolar Deep Water (CDW). However, this study ~~by~~ Friedl et al. (2018) lacked direct measurements of basal melting and did not perform relevant numerical modelling ~~of the evolution of a sub-ice ocean cavity or coupling to a cavity ocean circulation model~~, so it is still uncertain whether the enhanced basal melting triggered by ocean warming is the dominant reason for the ungrounding process. ~~A positive feedback between basal sliding and basal water pressure (through friction heating) upstream of~~

115 ~~Subglacial melting occurring at the ice-bed interface away from~~ the grounding line could be another possible factor in the glacier acceleration ~~owing to a positive feedback between the basal sliding and subglacial melt water volume and grounding line retreat~~ (Bartholomaeus et al., 2008; Iken and Bindshadler, 1986; Schoof, 2010). ~~The possibility of such a feedback, is not ruled out by~~ Friedl et al. (2018), ~~and is discussed further in Sect. 4.2 and Sect. 5.~~

120 ~~Changes in basal shear stress connected with changes in glacier flow could reveal the possible movement of the grounding line and also indicate possible influences on the changing dynamics.~~In this study, we employed the Elmer/Ice code (<http://elmerice.elmerfem.org/>) (Gagliardini et al., 2013), a ~~new-generation~~ three-dimensional (3D) full-Stokes ice sheet model, to solve the Stokes equations over the whole WIS-FGL catchment. Our implementation of the model solves the ice flow equations and the steady-state heat equation (Gagliardini et al., 2013; Gladstone et al., 2014). We also infer the basal shear stress using ~~control-an~~ inverse methods (e.g., Gillet-Chaulet et al. (2016); Gong et al. (2017)).

In ~~the~~ first part of this study (Zhao et al., companion paper), we explored the sensitivity of the inversion for basal shear stress to: ~~enhancement of ice deformation rates, bedrock elevation data, the ice front boundary condition, and~~ initial ~~model~~ assumptions about englacial temperatures, ~~bed elevation data, and ice front boundary condition~~. In ~~this second part of this study~~ (the current paper), we adopt the three-cycle spin-up scheme of Zhao et al. (companion paper) to derive the distributions of basal shear stress in 2008 and 2015. We present the observational data in Sect. 2 and our methods in Sect. 3. We compare the resulting basal shear distributions for ~~the~~ 2008 and 2015 and their connections with driving stress and basal friction heating in Sect. 4.1 and Sect. 4.2. The height above buoyancy for the two epochs is computed in Sect. 4.3 as an independent guide to grounding line changes. Through comparison of basal shear stress and height above buoyancy between 2008 and 2015, we analyze the stability of the grounding line in this period and discuss ongoing marine ice sheet instability and direct oceanic forcing as possible reasons for the ~~sharp~~ speed-up of the FGL in Sect. 5.

2 Observational Data

2.1 Surface elevation data in 2008 and 2015

145 The surface elevation data ~~set~~ for 2008 (DEM2008; Fig. 2a) from Zhao et al. (companion paper) ~~plays a central role was used~~ here. To estimate the surface topography in 2015 (DEM2015; Fig. 2a), we generated the average surface-lowering rate during 2008-2015 for the fast flow regions (surface velocity in 2008 ≥ 20 m yr⁻¹) by using the hypsometric model for elevation change described in Zhao et al. (2017) ~~during-for~~ the same period. The DEM2015 was then generated from ~~the~~ DEM2008 ~~by~~ applying ~~these~~ ice thinning rates from

2008 to 2015. For the area with velocities $< 20 \text{ m yr}^{-1}$, we assume the DEM in 2015 remains the same as that in 2008.

2.2 Bed elevation data

The bed topography plays a significant role in simulation of basal sliding and ice flow distribution for fast-flowing glaciers (Zhao et al., companion paper), and also in interpreting the grounding line movement precisely (De Rydt et al., 2013; Durand et al., 2011; Rignot et al., 2014). Zhao et al. (companion paper) ~~discussed-investigated~~ the sensitivity of the basal shear stress distribution to three bedrock topography datasets, ~~and~~ The bedrock dataset, bed_zc (Fig. 2b), with higher accuracy and resolution, was suggested as the most suitable ~~bedrock data~~ for modelling the WIS-FGL system. Here-Recall that bed_zc is computed by:

$$\text{bed_zc} = S_{2008} - H_{\text{mc}} \quad (1)$$

where S_{2008} is the surface ~~DEM elevation~~ in 2008 combined from two DEM products as discussed in Zhao et al. (companion paper), and H_{mc} is the ice thickness data with a resolution of 450 m combined from the ice thickness data computed using a Center for Remote Sensing of Ice Sheets (CREGIS) ice thickness measurements using a mass conservation method for the regions of faster flow (Morlighem et al., 2011; Morlighem et al., 2013), and ice thickness from Bedmap2 for other regions (Fretwell et al., 2013). A complete description is given by Zhao et al. (companion paper).

2.3 Surface velocity data in 2008 and 2015

We use the same velocity data for 2008 as in Part ~~1-A~~ of this study (Zhao et al., companion paper), which is from the InSAR-based Antarctic ice velocity dataset (MEASUREs (version 1.0) produced ~~from the fall 2007 and/or 2008~~ by Rignot et al. (2011c) from fall 2007 and/or 2008 measurements over the study area). The 2008 velocity dataset has a resolution of 900 m and the uncertainties over the study region ranges from 4 m yr^{-1} to 8 m yr^{-1} ~~over the study area~~. For 2015, we adopt the velocity data extracted from Landsat 8 imagery with a resolution of 240 m and errors ranging from 5 m yr^{-1} to 20 m yr^{-1} (Gardner et al., 2018). The vVelocity dataset ~~in~~ for 2015 has a full coverage over the WIS-FGL domain, while the velocity in 2008 has no data in the gray area in Fig. 1b.

2.4 Other datasets

The steady state temperature field is simulated from an initial temperature field, with a linearly interpolated initial temperature between upper and lower ice surfaces, which leads to robust ~~does not affect the final~~ inversion results as demonstrated by Zhao et al. (companion paper).- The surface temperature is constrained by yearly averaged surface temperature over 1979-2014 computed from RACMO2.3/ANT27 (van Wessem et al., 2014) and the basal temperature is initialized to pressure melting temperature. The temperature simulations utilize the spatial distribution of bottom heat flux boundary condition includes the geothermal heat flux ~~from estimated by~~ Fox Maule et al. (2005) and the simulated basal frictional heating.

Our DEM is an ellipsoidal WGS84 system and hence a height of 0 m does not refer to sea level. An observed sea level height of 15 m (WGS84 ellipsoidal height) in Marguerite Bay (Zhao et al., companion paper) was taken to compute the sea pressure on the ice front.

3 Method

The modelling method using Elmer/Ice presented in Part ~~1-A~~ of this study (Zhao et al., companion paper) is adopted here, including the mesh generation, mesh refinement, model parameter choices and ~~applied~~ boundary conditions. The simulations for both 2008 and 2015 the two epochs retain the same assumptions about the ice-covered domain, namely a

195 common spatial extent with fixed ice front location, and the assumption that all the ice is
grounded. The ice front position is assumed to coincide with the 1996 grounding line position
(Rignot et al., 2011a). This assumption might be incorrect for the main branch of the FG, and
we evaluate it based on the deduced floating area where the inferred basal shear stress is
lower than a threshold, which is discussed in Sect. 4.1. It is very clear from satellite imagery
200 that in 2008 a small ice shelf is still present in front of the southern FG and the Prospect
Glacier (hereafter PG) (Fig. 1c). In 2015 the ice shelf in front of the southern FG has
disappeared, while the floating part of the PG has retreated in the east and re-advanced in the
west (Fig. 1c). However, we don't include the floating parts of the southern FG and PG in
either epoch in this study, owing to the lack of the ice shelf thickness data.

205 We follow the three-cycle spin-up scheme (Zhao et al., companion paper) and simulate the
basal shear stress τ_b in 2008 and 2015 with the linear sliding law:

$$\tau_b = -Cu_b \quad (2)$$

Here C is a-the basal dragfriction coefficient, a variational parameter in the inversion
procedure, and u_b is the basal sliding velocity.

210 There are two key differences between the data used for the 2008 and 2015 inversions:
increased surface velocity and changed ice geometry, namely a thinner glacier in 2015
compared to 2008 due to dynamic thinning. To explore their relative impacts, we carry out an
additional inversion with the geometry from 2008 but the surface velocity from 2015 (see
Appendix A Sect. S1 in the supplementary material). We found that both geometry variations
and velocity changes are important to the inverted basal stress condition.

215 To explore the relationship between the basal shear stress and local gravitational driving
stress τ_d , the gravitational driving stress is also computed for both epochs:

$$\tau_d = \rho_i g H |\vec{\nabla} z_s| \quad (3)$$

220 where ρ_i is the ice density, g is the gravitational constant, H is the ice thickness, and $|\vec{\nabla} z_s|$ is
the gradient of the ice surface elevation. Considering the snow and firn on the ice surface, we
apply a relatively low ice density of 900 kg m^{-3} following Berthier et al. (2012).

Hoffman and Price (2014) also found a positive feedback between the basal melt and basal
sliding through the frictional heating on for an idealized mountain glacier using coupled
subglacial hydrology and ice dynamics models. To explore possible effects of changes of
basal frictional heating between 2008 and 2015, we compute the friction heating (q_f)
225 generated at the bed:

$$q_f = \tau_b u_b \quad (4)$$

~~Subglacial water has the capacity to modulate ice velocity and mass balance for outlet~~
~~glaciers.~~ To explore the possible flow path of subglacial water beneath the FGL, we calculate
hydraulic potential at the bed, ~~and since the its negative gradient of this governs~~
230 ~~determines~~ subglacial flow direction. The hydraulic potential, (Φ), expressed in equivalent metres of
water, is given by:

$$\Phi = (z_s - z_b) \frac{\rho_i}{\rho_{fw}} + z_b \quad (45)$$

235 where ρ_{fw} is the fresh water density (1000 kg m^{-3}), and z_s and z_b are the surface and bed
elevations, respectively. Here we assume that the water pressure in the subglacial hydrologic
system is given by the ice overburden pressure, which is equivalent to assuming that the
effective pressure at the bed, N , is zero. (Shreve, 1972).

240 Height above buoyancy (Z_*) is ~~an good~~ indicator of how ~~heavily grounded~~ close to floatation a marine-based glacier is, which is relevant to the glacier's evolution and additionally helps ~~interpret the~~ identify likely floating regions ~~based on simulated basal shear stress in this study~~. Z_* is related to the effective pressure N at the bed by the relationship:

$$N = \rho_i g Z_* \quad (56)$$

245 In this study, we use a simpler hydrostatic balance based on sea level with the relationship:

$$Z_* = \begin{cases} H, & \text{if } z_b \geq z_{sl} \\ H + (z_b - z_{sl}) \frac{\rho_w}{\rho_i}, & \text{if } z_b < z_{sl} \end{cases} \quad (67)$$

250 where ρ_w is the density of ocean water and z_{sl} is the sea level. This expression for Z_* assumes a perfect connectivity of the basal hydrology system with the ocean. This is appropriate for the present study where we are exploring the degree of grounding of the fast flowing regions of the FG over the downstream basin.

4 Results

4.1 Comparison of basal shear stress and driving stress in 2008 and 2015

255 We obtain the spatial distributions for basal shear stress, τ_b (Figs. 3a, 3b), and basal velocity of the WIS-FGL system for 2008 and 2015 using ~~the an~~ inversion method to determine the basal dragfriction coefficient, C , with the geometry and velocity data described above. Although low-resolution estimation of basal shear stress has been carried out for the whole Antarctic Ice Sheet (Fürst et al., 2015; Morlighem et al., 2013; Sergienko et al., 2014), this is the first application of inverse methods to estimate the basal friction pattern of the Fleming system at a high resolution and use the full-Stokes equations.

260 In 2008 the main FGL shows ~~a some sticky spots band~~ of high basal shear stress approximately 2 km wide close to the ice front (Fig. 3a). The backstress exerted by these sticky spots with $\tau_b > 0.01$ MPa (shown in Fig. S3) is $\sim 3.42 \times 10^{11}$ N, while immediately upstream a region of low basal stress covers most of the downstream bedrock basin, returning to more typical values (~ 0.05 - 0.53 MPa) ~ 9 km from the ice front. In contrast, the basal friction at the front of the southern FG is low, with more typical values ~ 2 km upstream. By 2015, the high dragfriction band-spots near the FGL ice front has have disappeared while in the downstream basin the region of already low basal drag shear stress already seen in 2008 is more extensive and even even lower in value in 2015 (Fig. 3b). which This is consistent with the observed speed-up from 2008 to 2015. Further upstream in thise basin, including and over the ridge between the downstream and upstream basins, the basal shear stress does not change much between the two epochs (Fig. 3c).

275 To explore the ice dynamics evolution from 2008 to 2015, we present the ratio of basal shear stress τ_b to driving stress τ_d (hereafter referred as “RBD”) in Figs. ~~3e3d, 3d3e~~, which can provide insight into the dynamical regime (Morlighem et al., 2013; Sergienko et al., 2014). In particular, it provides an indication whether the driving stress is locally balanced by the basal shear or whether there is a significant role for membrane stresses and a regional momentum balance. We assume designate the region with $\tau_b < 0.01$ MPa or RBD < 0.1 to be theas a “low dragfriction” area, considering the uncertainties of the model input, and the very low inferred basal drag is assumed to correspond to potentially indicative of floatation, i.e. ungrounded ice.

280 ~~It is hard to determine whether t~~ The high basal shear stress band-spots inferred by the inversion detected at the front of the main branch of the FGL in 2008 (Fig. 3a) is may be a real feature or at least in part an artifacts error due to uncertainties from the ice thickness, local bed topography, local sea level, ice mélange backstress, and the ice front position (as

285 discussed in Zhao et al. (companion paper)). Sensitivity to such uncertainties was explored in
Zhao et al. (companion paper), and the adjustments of ice front boundary condition with a
higher sea level of 25 m or an advanced ice front position showed as decrease in the basal
shear stress friction coefficients around near the ice front, but did not completely remove
290 has not shown any sign of disappearance of these high basal drag friction bands. This implies
that the front of the FG in 2008 might still be partly grounded on the 1996 grounding line due
to the presence of real pinning points. ~~Improved bed topography data and accurate ice front~~
~~position are necessary to interpret the precise grounding line position in 2008.~~

As expected, the gravitational driving stress of this system shows no significant changes from
2008 (Fig. 3e) to 2015, except for the front of PG (Fig. 3f). In 2015, the boundaries of the
295 zone in the main FGL with $\tau_{b2015} < 0.01$ MPa (magenta-blue lines in Fig. 3b and Fig. 4) or
RBD₂₀₁₅ < 0.1 (red lines in Fig. 3d and Fig. 4) have some similarity to ~~are partly consistent~~
with the deduced grounding line position of the FGL in 2014 from Friedl et al. (2018) (white
dots in Figs. 3 and 4). The differences with that study are around the northern southern and
eastern parts, but the magenta-blue and red boundaries (Figs. 4c, 4d) in the northern part fit
300 the bedrock ridges in the present study (Figs. S2b), while the white points fit the
corresponding bedrock topography data in used by Friedl et al. (2018). This result comparison
confirms the significant role of bedrock topography in determining the grounding line
position. Around the eastern part of the region within which velocities > 1500 m yr⁻¹ (cyan
contour in Fig. 3b), the low basal drag friction area in this study extends ~1-3 km further
305 upstream than the extracted estimated grounding line in 2015-2014 (Friedl et al., 2018). An
unexplained rib-like basal resistance pattern ($\tau_{b,2015} > 0.1$ MPa) is found approaching the
Fleming front parallel to the yellow velocity contour (Fig. 3b). This feature, which is not
present in 2008 (Fig. 3a), is located within the boundary area from topographic low to high
along the southern margin of the downstream FGL (Fig. 4d).

310 Comparison of basal shear stress between 2008 (Fig. 3a) and 2015 (Fig. 3b,c) shows a
significant decrease from 2008 to 2015 in fast flowing regions (velocity > 1500 m yr⁻¹) at the
front of the FGL. A similar pattern occurred at front of the PGL and the southern FGL. For
the northern section of the southern FGL, the grounding line has retreated by ~2 km in 2008
from the last known grounding line position in 1996 (Rignot et al., 2011a) (Fig. 3a), which is
315 reasonable considering that the northern section of the ice front has retreated ~2 km behind
the 1996 grounding line position (Fig. 1c). However, it is not clear whether the southern
section of the southern FG has also retreated in 2008 as indicated in Fig. 3a, and whether the
floating area has expanded ~3 km further inland in 2015 based on the decreased basal shear
stress from 2008 (Fig. 3a) to 2015 (Fig. 3b). Similarly, it is also hard to estimate the possible
320 grounding line positions of the PG based from the inferred basal shear stress in both epochs.
That is because we did not account for the normal stress of the remnant small ice shelf at the
front of the southern FG and the PG (Fig. 1c) in the inverse modelling. The surface lowering
in DEM2015 for the PG could also be an artefact since no observations were available for the
PG when building the hypsometric model that generates the DEM2015 (see inset map in Fig.
325 2a; Zhao et al. (2017)) and continued retreating by ~3 km upstream in 2015 (Fig. 3b). For the
PGL, the grounding line in 2008 largely coincides with that in 1996 (Fig. 3a) but retreats by
~3 km until 2015 (Fig. 3b). We attribute this decreased basal friction to the ice ungrounding
process from 2008 to 2015.

4.2 Basal melting and subglacial hydrology

330 Increases in ~~s~~Subglacial water pressure could ~~be a~~ contribute ~~or~~ to lower basal shear stress and
higher basal sliding at the base of the FGL-FG, potentially through the positive hydrology
feedback mentioned earlier. That feedback mechanism can be summarized simply: a general
acceleration of glacier flow (for example due to a backstress reduction from ice shelf collapse
or unpinning from a sticky spot) can lead to increased basal sliding in regions where the basal
335 shear stress almost remains unchanged (for example in the FG trunk above the downstream
basin (Figs. 3a-c). This increases friction heating and basal melt water generation, which - as

suggested by Hoffman and Price (2014) - may increase the effective basal water pressure downstream, thereby increasing sliding speeds (Gladstone et al., 2014; Hoffman and Price, 2014). Since the reduction of effective pressure is the key process to enhance sliding, this positive feedback is dependent on a positive feedback of melt water generation to water pressure. This dependence can break down when there is sufficient basal water to generate efficient drainage channels (Schoof, 2010). However, such efficient channelization in the subglacial hydrologic system is typically associated with seasonal surface meltwater pulses reaching the bed (Dunse et al., 2012), a process that is not expected to occur for Fleming Glacier (Rignot et al., 2005).

Basal melt water arises from two main sources in polar regions: either surface melt water draining into the subglacial hydrologic system via crevasses or moulins or in-situ melting at the bed (Banwell et al., 2016; Dunse et al., 2015; Hoffman and Price, 2014). Hoffman and Price (2014) also found a positive feedback between the basal melt and basal sliding through the frictional heating on an idealized mountain glacier using coupled subglacial hydrology and ice dynamics models. However, the amount of surface melt water in the WIS-FGL region is not thought to be sufficient to percolate to the base (Rignot et al., 2005), so we take basal melting due to the friction heat and geothermal heat flux as the only source of subglacial water. The geothermal heat flux at-in the fast flowing regions of our study area (Fox Maule et al., 2005) is two orders of magnitude smaller than the friction heating at the base, leaving friction heating as the dominant factor in generating basal melt water.

To explore the potential subglacial water sources and the likely flow directions, we plot the frictional heating (Figs. 4a, 4b), the contours of hydraulic potential (Φ) (Figs. 4d, 4e), and the basal homologous temperature (temperature relative to the pressure melting point) (Figs. 4g, 4h) for both epochs. Friction heating due to sliding at the bed (Figs. 4a, 4b) provides a basal melt water source where ice is at pressure melting temperature, which is the case for the fast flow regions of the FGL (see the basal homologous temperature relative to the pressure melting point in Figs. 4g, 4h), and while the gradient of the hydraulic potential (Figs. 4d, 4e) indicates likely water flow paths at the ice-bed interface. The frictional heat generated at the base is high where both basal shear stress and basal sliding velocities are high. The modeled friction heating in both 2008 and 2015 (Figs. 4a, 4b) extends as far and high as in the upstream basin under the FGL, indicating high basal melt rates in this region (a heat flux of 1 W m^{-2} could melt ice at the rate of 0.1 m yr^{-1} in regions at the pressure melting temperature). The highest friction heating is generated over the bedrock rise between the FGL upstream and downstream basins, where the most melt water will be generated-produced and will-this will be routed towards the downstream basin given the gradient of hydraulic potential in this region (Figs. 4g, 4h). Hence it is a major source of basal water for the downstream basin. This could explain the low basal dragfriction downstream, while the increase in heating between 2008 and 2015 (Fig. 4c) could further enhance the basal sliding in the fast-flowing regions, contributing to the observed accelerations. Both the hydraulic potential and frictional heating could help to understand the mechanism behind the rapid acceleration and surface draw-down of the FGL, which is further discussed in Sect. 5.

4.3 Height above buoyancy Z_*

We compute the height above buoyancy, Z_* , for 2008 and 2015 for the FGL based on Eq. (67) with a sea level of 15 m (Figs. 5a, 5b). To allow for the over- or under-estimation of Z_* owing to uncertainties from the topography data, ice thickness, ice density and the sea level applied above, we suggest that the areas where $Z_* - |Z_*| < 20 \text{ m}$ might be floating, and accordingly while include-including areas where $Z_* > -20 \text{ m}$ in Fig. 5.

In 2008 A-a low height above buoyancy Z_* in 2008 (Fig. 5a) is only found near the 1996 grounding line position in the downstream basin, which reveals-indicates that ungrounding of the main FGL may not have started or only just commenced in 2008. In 2015, the area close

390 to flotation with $Z_* < 20$ m (taken as an upper limit) has expanded, reaching about 9 km upstream ~~in 2015~~ (magenta lines in Fig. 5b), which broadly coincides with the estimated grounding line in 2014~~5~~ (Friedl et al., 2018) except for an almost encircled patch with slightly higher Z_* (20-30 m). The implications of the different Z_* from 2008 and 2015 are a small FGL grounding line retreat from 1996 to 2008 but significant retreat from 2008 to 2015. Uncertainty in the predicted grounding line in 2015 is significant, but a new position ~9 km upstream is likely.

395 In addition to the main branch of the FGL, its southern branch and the PGL also show an expansion of the region~~reduction~~ in which Z_* is close to zero, which ~~suggests~~ indicates possible grounding line retreat. However, the DEM2015 used to compute Z_* has large uncertainties in the southern branch of FG and PG, since the surface lowering in DEM2015 for those regions could be artefacts due to the lack of observations as mentioned above (see
400 inset map in Fig. 2a; Zhao et al. (2017)). Therefore, it is hard to determine the current grounding line locations for those two glaciers. Based on the area with $Z_* < 20$ m, the southern FGL has retreated by ~1.5 km between 1996 and 2008 (Fig. 5a) and a further ~1.5 km by 2015, with an associated increase in floating area (Fig. 5b). The PGL does not show obvious sign of retreat between 1996 and 2008 but migrates for ~1 km upstream by 2015.

405 Changes in Z_* from 2008 to 2015 suggest the creation of an ungrounded area consistent with the area of very low modelled basal shear stress shown in Figs. 3a and 3b. The This change in area close to floating, defined by $Z_* < 20$ m, constitutes additional evidence supporting the hypothesis of rapid grounding line retreat over 2008 to 2015 and the likely grounding line positions of the FG in both epochs.

410 5 Discussions

The A sticky spots of band of high basal shear stress near the terminus of the FGL in 2008 might be artefacts, but the possibility that this high friction area is a real feature due to some pinning points is not excluded. If the high basal resistance spots are artefacts, ungrounding of this region in early 2008 is less viable as an explanation for an abrupt increase in ice flow speed, since the loss of backstress would be more gradual. In this case, positive feedbacks, such as the marine ice sheet instability or the basal melt feedback, are even more likely to explain the FG's recent behavior. If the sticky spots are real features, the implication
415 is suggests that the ice front might have been at least partly still grounded at that time in early 2008, an. This interpretation is consistent with the relatively high bedrock topography near the ice front compared to upstream (Fig. 1c). Friedl et al. (2018) ~~deduced~~ proposed that the likely grounding line position of the FGL in after Jan-Apr 2008 must have been located upstream of at a possible small hill from the bedrock topography (~2.5 km upstream of the 1996 grounding line) as from their interpretation of rapid abrupt surface acceleration detected around March the same period April 2008. This is also confirmed by the fact that the glacier front had retreated behind the 1996 grounding line during the acceleration phase (Friedl et al., 2018). However, The acceleration phase in March April 2008 occurred later than the timing of the DEM2008 data used in this study (acquired in January 2008 for fast flowing regions). Therefore it is quite possible that these grounding line retreat had not retreated by occurred after January 2008, when our DEM2008 was acquired. The analysis of height above
420 buoyancy for ~~the~~ DEM2008 and inferred basal shear stress in 2008 supports the main FGL being grounded close to the ice front and hence near the 1996 grounding line location. Considering Given the uncertainties of grounding line position in 1996 (several kilometers) (Rignot et al., 2011a) and uncertainty about interpreting the frontal high basal drag friction band area in this study, the exact grounding line position in January 2008 is somewhat
430 uncertain, as is the extent of any retreat associated with the significant acceleration during March April 2008. Improved bed topography/ice thickness data and accurate historic ice front position are necessary to interpret the precise grounding line position in 2008. Detailed

bathymetry of the relevant location might become available if the ice front of the FG retreats in future.

440 The disappearance of ~~a the inferred~~ high basal ~~resistance-shear region~~band (a ~~likely possible~~ physical pinning ~~bandpoints~~) near the FGL front between 2008 and 2015 is a ~~likely possible~~ trigger for the sudden acceleration and increased surface lowering of the FG ~~during this period~~. The increased flux of ice, combined with the ~~changed~~ glacier geometry, suggests the substantial grounding line retreat, which agrees with two recent studies (Friedl et al., 2018; Walker and Gardner, 2017). The timing of ~~these the~~ acceleration, ~~whichs~~ occurred in Jan-Apr 2008 (Friedl et al., 2018), suggests that the loss of this basal resistance occurred shortly after the first epoch we analyzed (Jan 2008). Given the low basal ~~dragfriction~~ already present over most of the downstream basin (a possible cavity proposed by Friedl et al. (2018)), one would expect the loss of the localized ~~dragfriction~~ near the ice front to promptly result in an increase in velocity over the entire low-~~dragfriction~~ region. This is consistent with the near uniform increase in velocity ~~reported in in early Apr 2008 -for a region 4-10 km upstream of the 1996 grounding line reported~~ by Friedl et al. (2018) ~~for a region 4-10 km upstream of the 1996 grounding line.~~

455 For a glacier lying on a retrograde slope in a deep trough, the grounding line may be vulnerable to rapid retreat without any further change in external forcing, once its geometry crosses a critical threshold, which is the marine ice sheet instability hypothesis (e.g., Mercer (1978); Thomas and Bentley (1978); Weertman (1974)). A similar theory has been proposed on the prospective rapid retreat of Jakobshavn Isbræ in West Greenland without any trigger after detaching from a pinning point (Steiger et al., 2017). The FG grounding line in early 2008 may have experienced a retreat after moving across the geometric pinning ~~bandpoints~~ near the front, and then retreated further to the position ~~in 2015~~ about 9 km upstream in the FGL downstream basin by 2015. This has been proven by Friedl et al. (2018), and they also suggested that a further stage of grounding line retreat of the FG may have happened between Mar 2010 and early 2011. A similar ungrounding process has been detected in the Thwaites, 465 Smith and Pine Island Glaciers from 1996 to 2011 (Rignot et al., 2014).

The current grounding line of the FG (Friedl et al., 2018) ~~appears to be on the prograde slope of the bedrock high between the FGL downstream and upstream basins. With the establishment of an ocean cavity under the new ice shelf we can expect that ocean-warming driven basal melting will further modify the thickness of the recently ungrounded ice. If the system remains out of balance and continues to thin, the grounding line could eventually move across this bed obstacle, and if this occurs, the grounding line is then likely to retreat rapidly down the retrograde face of the FGL upstream basin, likely to be accompanied by further glacier speed up and dynamic thinning, unless the ice shelf buttressing of an increasingly long and confined fjord-like Fleming ice shelf increases sufficiently to restore its stability.~~

480 Walker and Gardner (2017) attribute the ~~sharp-significant~~ increase in observed ice velocity and drop in surface elevation from 2008 to 2015 to increased calving front melting caused by incursion of relatively warm Circumpolar Deep Water (CDW). The CDW flows onto the continental shelf within the Bellingshausen Sea, penetrating into ~~the~~ Marguerite Bay, driven by changes in regional wind patterns resulting from global atmospheric circulation changes (Walker and Gardner, 2017). Friedl et al. (2018) also explain the unpinning from the 1996 grounding line position in 2008 and further landward migration of the grounding line in 2010-2011 with the same mechanism, namely the increased ~~front and/or~~ basal melting due to ocean warming. This explanation appears consistent with the finding that the acceleration, retreat, and thinning of outlet glaciers in the Amundsen Sea Embayment (ASE) are triggered by ~~the ungrounding process due to~~ the inflow of warm CDW onto its continental shelf and into sub-ice-shelf cavities (Turner et al., 2017). However, the floating parts of the FGL remained negligible in 2008 as indicated in Sect. 4.3 based on the height above buoyancy in 2008 (Fig. 5a). The speedup and ungrounding occurring in the ASE glaciers was a direct response to 490 significant loss of buttressing caused by ice shelf thinning and grounding-line retreat (Turner

et al., 2017). When the CDW incursions started in the ASE, the floating parts of ASE glacier systems were much larger than the residual ice shelf of the Fleming system in 2008. After the recent changes the newly floating region of the FGL has an area of ~60 km², based on the estimated 2014~~5~~ grounding line from Friedl et al. (2018) and the 2016 ice front position in this study~~-, which is consistent with o~~ Our height above buoyancy analysis for 2015 (Fig. 5b) ~~also indicates substantial grounding line retreat since 2008~~. So, significant buttressing reduction is not likely to have occurred on the FGL during the rapid acceleration of 2008, but further changes to the FGL after 2015 may resemble ASE glacier and ice shelf systems more closely. No direct measurements are available to confirm the direct effect of the frontal or basal melting on the FGL grounding zone over this period, nor have previous studies attempted to quantify the amount of melting required to drive significant FGL grounding line retreat. The ocean-driven basal melting at the ice shelf front or base may have contributed to grounding line retreat, or the reduction of the frontal high basal shear zone, but establishing this as the main cause would require further quantification of the cause-effect link.

Ongoing thinning as a result of backstress reduction following the collapse of the WIS is another possible cause for the recent ungrounding. The WIS evolved from an embayment-wide ice shelf in 1966 to smaller individual remnant ice shelves in 1997 (Fig. 1b) (Cook and Vaughan, 2010; Wendt et al., 2010). The floating part of the FG in particular was in the form of an ice tongue in 1997 (Cook and Vaughan, 2010), and as such would likely have imposed much lower backstress on the grounded part. Point measurements indicate that the FG accelerated by 40-50% between 1974 and 1996 (Doake, 1975; Rignot et al., 2005). If this acceleration was a response to loss of buttressing, the FG system may have been out of equilibrium, and losing mass, since before 1996. If the increased velocity in response to shelf collapse was maintained over time, maintaining persistent thinning, eventual ungrounding of the bedrock high where the 1996 grounding line was located would occur independently of ocean-induced increased shelf melt. The recent accelerations and enhanced thinning (Friedl et al., 2018; Gardner et al., 2018; Walker and Gardner, 2017) may indicate an ongoing response to the WIS collapse, amplified by positive feedbacks within the FG system.

Rapid sliding at the base is dependent on the presence of a sub-glacial hydrologic system~~Ongoing presence of subglacial water could contribute to a radical destabilization of marine ice sheet systems.~~ Evidence suggests that increased basal water supply could accelerate basal motion ~~and surface lowering~~ of both mountain glaciers (Bartholomaeus et al., 2008) and ice sheets (Hoffman et al., 2011), presumably by changing the subglacial water pressure or bed contact, and further contribute to grounding line retreat of marine-based glaciers. Jenkins (2011) has also suggested that subglacial water emerging at the grounding line can enhance local ice shelf basal melt rates by driving buoyancy driven plumes in the ocean cavity. The rapid sliding and high friction heating in the upstream FGL (Figs. 4a, 4b), together with the direction of the hydraulic potential gradient (Figs. 4d, 4e), has provided evidence for an extensive active hydrologic system beneath the FG, which might already have been enhanced by the previous significant WIS collapse that occurred before 2008.

High basal frictionally generated heating in the fast flowing regions~~upstream basin~~ of the FGL is the main source of meltwater flowing into the FGL downstream basin. It is also clear that the frictional heating in 2015 (~~Fig. 4b~~) was greater than in 2008 in the upstream basin (Fig. 4a~~4c~~), with the increase in basal meltwater production peaking over the bedrock rise between the downstream and upstream basins indicating more basal melt water generation in 2015 (see Sect. S2 and Fig. S4). The plateaus in hydraulic potential in both downstream and the upstream basins of the FGL suggests the possibility that basal water may accumulate in ~~this~~ those regions, or at least show a low throughput. The downstream plateau appears to be fed by a large frictional heat source over the ridge between the downstream and upstream basins in addition to flow from further inland, while This the upstream plateau appears to be fed by an extensive upstream region of basal melting with a large frictional heat source. There might be some pooling of water in those plateaus in 2008, but the inferred basal shear stress (Fig. 3a) and the height above buoyancy (Fig. 5a) indicate that those regions should still

545 ~~remain grounded. Outflow from this plateau region, a~~ According to our hydraulic potential calculations, outflow from the upstream plateau region is likely to be predominantly in the direction of the downstream basin, but future outflow across the shallow saddle in hydraulic potential towards the Southern-southern branch of the FG cannot be ruled out, since the evolution of the potential responds to the changing elevation, as can be seen by-in comparing the contours in Figs. 4e and 4d4f.

550 The further sharp-abrupt speed-up events that occurred in 2010-2011 reported by Friedl et al. (2018) could have several potential causes in addition to the previously proposed mechanism of a direct response to ocean-induced melting (Walker and Gardner, 2017). One possibility is an outburst of subglacial water from the upstream basin after ~~subglacial water~~ building up over years to decades in response to increased sliding and friction heating and progressive
555 lowering of the ice surface. Another possibility is local unpinning near the retreating grounding line: ungrounding from pinning points may cause a step reduction in basal resistance. -This unpinning could be a feature of ongoing thinning in response to WIS collapse, as discussed above. Another possibility could be positive feedbacks in the subglacial hydrologic system – rapid change may result from the direct feedback between changes in
560 sliding speed, friction heat and basal water production.

The height above buoyancy is an indicator for the vulnerability of marine-based grounded ice to dynamic thinning and acceleration. The area with $Z_* < 20$ m in 2015 has shown that the downstream basin is currently ungrounding and this may continue until the grounding line finds a stable position on the prograde slope separating the two major basins. More thinning
565 would be needed to destabilise the upstream basin, and it is hard to say-estimate how much forcing would be needed to push the grounding line into the upstream basin boundary~~into it~~. If the retrograde slope of the upstream basin is reached, further rapid and extensive grounding line retreat would be expected. A clear decrease can be seen in Z_* from 2008 (red in Fig. 5a) to 2015 (dark red in Fig. 5b) in the upstream basin (around the ~~2015-2008~~ velocity contour of
570 1000 m yr^{-1}), indicating the potential vulnerability of the FGL to continued ice mass loss. The surface lowering rate between 2008 and 2015 in this region is ~~$\sim 4.6-6$~~ $\sim 4.6-6 \text{ m yr}^{-1}$ (Zhao et al., 2017). If this thinning trend-rate continues linearly with time, the ice in regions with Z_* of 200-300 m would be expected to unground in ~~$\sim 3045-50-65$~~ years. This could be-take a longer or shorter period since-if the future thinning rate cannot be expected to remain constant is not
575 linear with time.

In the absence of precise and accurate knowledge of bed topography and ice shelf/stream basal processes, the dominant cause of the recent FG ungrounding cannot be determined. Further research is necessary to better understand the dominant mechanisms.

6 Conclusions

580 We used a full-Stokes ice dynamics model~~solver~~ (Elmer/Ice) at high spatial resolution to simulate-estimate the basal shear stress, temperature and frictional heating of the Wordie Ice Shelf-Fleming Glacier system in 2008 and 2015. Both increased surface velocity and surface lowering during this period are important for the calculation of basal shear stress.

585 Decreased basal dragfriction from 2008 to 2015 in the Fleming Glacier downstream basin indicates significant grounding line retreat, consistent with change in the suggested floating area based on the geometry in 2015 and the deduced grounding line in 2014~~5~~ from Friedl et al. (2018). Grounding line retreat also occurred on the southern branch of the FGL and the PGL. Our height above buoyancy calculations also indicate the FGL downstream basin was close to flotation in 2015 and is vulnerable to continued ice thinning and acceleration.

590 Pronounced basal melting driven by oceanic warming in ~~the~~ Marguerite Bay may have contributed to-triggered the ungrounding of the Fleming Glacier front in early 2008, as previously suggested by Walker and Gardner (2017) and Friedl et al. (2018), but ongoing

595 ~~thinning following the collapse of Wordie Ice Shelf may also provide an explanation. In either case, feedbacks in the subglacial hydrologic system may provide the dominant trigger mechanism for rapid increases in basal sliding and ongoing ungrounding process. The derived basal shear stress distributions suggest a major influence was could have been the loss ungrounding of a narrow some sticky spots band of higher basal shear near the ice front of the main Fleming Glacier, as basal friction under most of the region considered afloat by 2015 was already low in 2008 (a possible subglacial cavity).~~

600 The marine-based portion of the Fleming Glacier extends far inland. It is not clear whether grounding line retreat into the Fleming Glacier upstream basin will occur without further forcing. Transient simulations with improved knowledge of bed topography are necessary to predict the movement of the grounding line and how long it will take to achieve a new stable state. Coupled ice sheet ocean modelling ~~will~~ may be required to explore the evolution of the ~~new ice shelf~~ melting and impact of buttressing from the remaining and new ice shelf on the grounded glacier. Future studies of the dynamic evolution of the Fleming Glacier system will enhance our understanding of its vulnerability to marine ice sheet instability and provide projections of its future behavior.

Appendix A: Sensitivity to velocity changes

610 ~~Figure A1 shows the results from the inversion for basal shear stress in 2008 (Fig. A1a), 2015 (Fig. A1b), and from another additional inversion with the geometry from 2008 but using surface velocity from 2015 (Fig. A1c). The basal shear stress of this hybrid simulation shows patterns and magnitudes between those of the standard 2008 and 2015 simulations. This suggests that changes in both ice geometry and velocities have comparable impact on the inferred basal shear stress distribution, with the implication that an inversion study based on a change in either velocity or geometry alone would underestimate the change in basal drag.~~

615

Author Contribution

Chen Zhao collected the datasets, ran the simulation, and drafted the paper. All authors contributed to the refinement of the experiments, the interpretation of the results and the final manuscript.

620

Acknowledgements

Chen Zhao is a recipient of an Australian Government Research Training Program Scholarship and Quantitative Antarctic Science Program Top-up Scholarship. Rupert Gladstone is funded by the European Union Seventh Framework Program (FP7/2007-2013) under grant agreement number 299035 and by Academy of Finland grant number 286587. Matt A. King is a recipient of an Australian Research Council Future Fellowship (project number FT110100207) and is supported by the Australian Research Council Special Research Initiative for Antarctic Gateway Partnership (Project ID SR140300001). Thomas Zwinger's contribution has been covered by the Academy of Finland grant number 286587. This work was supported by the Australian Government's Business Cooperative Research Centres Programme through the Antarctic Climate and Ecosystems Cooperative Research Centre (ACE CRC). This research was undertaken with the assistance of resources and services from the National Computational Infrastructure (NCI), which is supported by the Australian Government. We thank Alex S. Gardner for providing the velocity dataset for 2015 ~~and Mathieu Morlighem for the ice thickness data~~. We thank E. Rignot, J. Mouginot, and B. Scheuchl for making their SAR velocities publically available. We thank Yongmei Gong for advice on the analysis of hydraulic potential. SPOT 5 images and DEMs were provided by the International Polar Year SPIRIT project (Korona et al., 2009), funded by the French Space

635

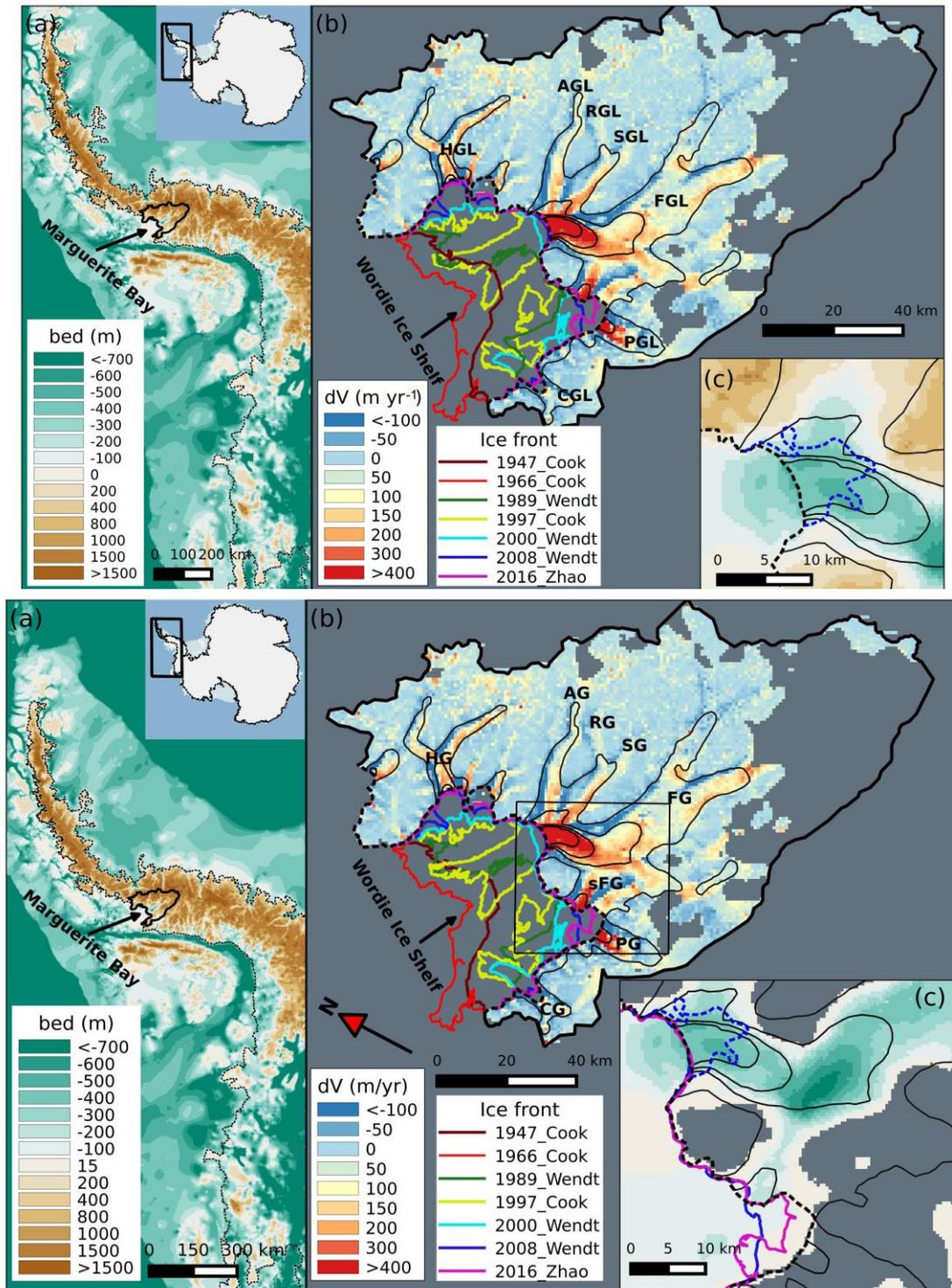
Agency (CNES). This work is based on data services provided by the UNAVCO Facility with
640 support from the National Science Foundation (NSF) and National Aeronautics and Space
Administration (NASA) under NSF Cooperative Agreement No. EAR-0735156. The ASTER
L1T data product was retrieved from https://lpdaac.usgs.gov/data_access/data_pool,
maintained by the NASA EOSDIS Land Processes Distributed Active Archive Center (LP
645 DAAC) at the USGS/Earth Resources Observation and Science (EROS) Center, Sioux Falls,
South Dakota.

References

- Banwell, A., Hewitt, I., Willis, I., and Arnold, N.: Moulin density controls drainage development beneath the Greenland ice sheet, *Journal of Geophysical Research: Earth Surface*, 121, 2248-2269, 2016.
- 650 Bartholomaus, T. C., Anderson, R. S., and Anderson, S. P.: Response of glacier basal motion to transient water storage, *Nature Geosci*, 1, 33-37, 2008.
- Berthier, E., Scambos, T. A., and Shuman, C. A.: Mass loss of Larsen B tributary glaciers (Antarctic Peninsula) unabated since 2002, *Geophysical Research Letters*, 39, 2012.
- 655 Cook, A. J., Holland, P. R., Meredith, M. P., Murray, T., Luckman, A., and Vaughan, D. G.: Ocean forcing of glacier retreat in the western Antarctic Peninsula, *Science*, 353, 283-286, 2016.
- Cook, A. J. and Vaughan, D. G.: Overview of areal changes of the ice shelves on the Antarctic Peninsula over the past 50 years, *The Cryosphere*, 4, 77-98, 2010.
- De Rydt, J., Gudmundsson, G. H., Corr, H. F. J., and Christoffersen, P.: Surface undulations of Antarctic ice streams tightly controlled by bedrock topography, *The Cryosphere*, 7, 407-417, 2013.
- 660 Doake, C. S. M.: Bottom sliding of a glacier measured from the surface, *Nature*, 257, 780-782, 1975.
- Dunse, T., Schellenberger, T., Hagen, J. O., Kääh, A., Schuler, T. V., and Reijmer, C. H.: Glacier-surge mechanisms promoted by a hydro-thermodynamic feedback to summer melt, *The Cryosphere*, 9, 197-215, 2015.
- Dunse, T., Schuler, T. V., Hagen, J. O., and Reijmer, C. H.: Seasonal speed-up of two outlet glaciers of Austfonna, Svalbard, inferred from continuous GPS measurements, *The Cryosphere*, 6, 453-466, 2012.
- 665 Durand, G., Gagliardini, O., Favier, L., Zwinger, T., and le Meur, E.: Impact of bedrock description on modeling ice sheet dynamics, *Geophysical Research Letters*, 38, n/a-n/a, 2011.
- Fox Maule, C., Purucker, M. E., Olsen, N., and Mosegaard, K.: Heat Flux Anomalies in Antarctica Revealed by Satellite Magnetic Data, *Science*, 309, 464-467, 2005.
- 670 Fretwell, P., Pritchard, H. D., Vaughan, D. G., Bamber, J. L., Barrand, N. E., Bell, R., Bianchi, C., Bingham, R. G., Blankenship, D. D., Casassa, G., Catania, G., Callens, D., Conway, H., Cook, A. J., Corr, H. F. J., Damaske, D., Damm, V., Ferraccioli, F., Forsberg, R., Fujita, S., Gim, Y., Gogineni, P., Griggs, J. A., Hindmarsh, R. C. A., Holmlund, P., Holt, J. W., Jacobel, R. W., Jenkins, A., Jokat, W., Jordan, T., King, E. C., Kohler, J., Krabill, W., Riger-Kusk, M., Langley, K. A., Leitchenkov, G.,
- 675 Leuschen, C., Luyendyk, B. P., Matsuoka, K., Mouginot, J., Nitsche, F. O., Nogi, Y., Nost, O. A., Popov, S. V., Rignot, E., Rippon, D. M., Rivera, A., Roberts, J., Ross, N., Siegert, M. J., Smith, A. M., Steinhage, D., Studinger, M., Sun, B., Tinto, B. K., Welch, B. C., Wilson, D., Young, D. A., Xiangbin, C., and Zirizzotti, A.: Bedmap2: improved ice bed, surface and thickness datasets for Antarctica, *The Cryosphere*, 7, 375-393, 2013.
- 680 Friedl, P., Seehaus, T. C., Wendt, A., Braun, M. H., and Höppner, K.: Recent dynamic changes on Fleming Glacier after the disintegration of Wordie Ice Shelf, Antarctic Peninsula, *The Cryosphere*, 12, 1-19, 2018.
- Fürst, J. J., Durand, G., Gillet-Chaulet, F., Merino, N., Tavard, L., Mouginot, J., Gourmelen, N., and Gagliardini, O.: Assimilation of Antarctic velocity observations provides evidence for uncharted pinning points, *The Cryosphere*, 9, 1427-1443, 2015.
- 685 Gagliardini, O., Zwinger, T., Gillet-Chaulet, F., Durand, G., Favier, L., de Fleurian, B., Greve, R., Malinen, M., Martín, C., Råback, P., Ruokolainen, J., Sacchetti, M., Schäfer, M., Seddik, H., and Thies, J.: Capabilities and performance of Elmer/Ice, a new-generation ice sheet model, *Geosci. Model Dev.*, 6, 1299-1318, 2013.
- 690 Gardner, A. S., Moholdt, G., Scambos, T., Fahnestock, M., Ligtenberg, S., van den Broeke, M., and Nilsson, J.: Increased West Antarctic and unchanged East Antarctic ice discharge over the last 7 years, *The Cryosphere*, 12, 521-547, 2018.
- Gillet-Chaulet, F., Durand, G., Gagliardini, O., Mosbeux, C., Mouginot, J., Rémy, F., and Ritz, C.: Assimilation of surface velocities acquired between 1996 and 2010 to constrain the form of the basal

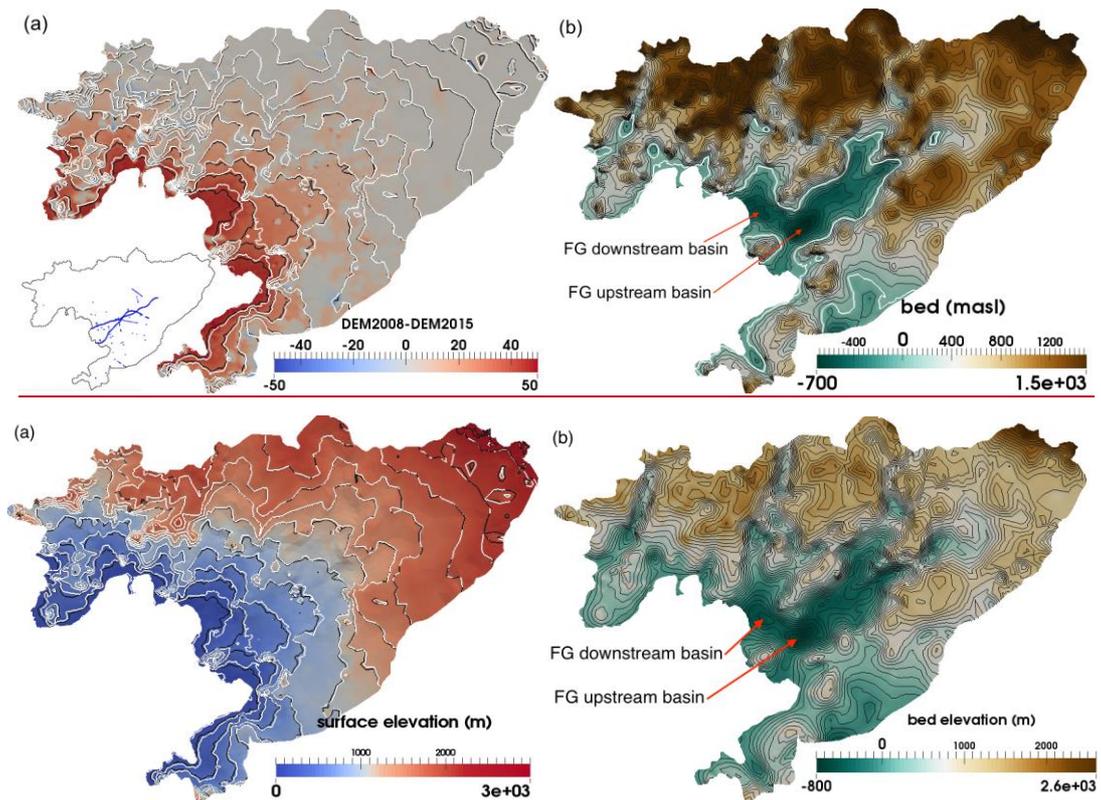
- 695 friction law under Pine Island Glacier, *Geophysical Research Letters*, doi: 10.1002/2016GL069937, 2016. n/a-n/a, 2016.
Gladstone, R., Schäfer, M., Zwinger, T., Gong, Y., Strozzi, T., Mottram, R., Boberg, F., and Moore, J. C.: Importance of basal processes in simulations of a surging Svalbard outlet glacier, *The Cryosphere*, 8, 1393-1405, 2014.
- 700 Gong, Y., Zwinger, T., Cornford, S., Gladstone, R., Schäfer, M., and Moore, J. C.: Importance of basal boundary conditions in transient simulations: case study of a surging marine-terminating glacier on Austfonna, Svalbard, *Journal of Glaciology*, 63, 12, 2017.
Gudmundsson, G. H.: Ice-shelf buttressing and the stability of marine ice sheets, *The Cryosphere*, 7, 647-655, 2013.
- 705 Gudmundsson, G. H., Krug, J., Durand, G., Favier, L., and Gagliardini, O.: The stability of grounding lines on retrograde slopes, *The Cryosphere*, 6, 1497-1505, 2012.
Hoffman, M. and Price, S.: Feedbacks between coupled subglacial hydrology and glacier dynamics, *Journal of Geophysical Research: Earth Surface*, 119, 414-436, 2014.
Hoffman, M. J., Catania, G. A., Neumann, T. A., Andrews, L. C., and Rumrill, J. A.: Links between acceleration, melting, and supraglacial lake drainage of the western Greenland Ice Sheet, *Journal of Geophysical Research: Earth Surface*, 116, n/a-n/a, 2011.
- 710 Iken, A. and Bindshadler, R. A.: Combined measurements of Subglacial Water Pressure and Surface Velocity of Findelengletscher, Switzerland: Conclusions about Drainage System and Sliding Mechanism, *Journal of Glaciology*, 32, 101-119, 1986.
- 715 Jenkins, A.: Convection-Driven Melting near the Grounding Lines of Ice Shelves and Tidewater Glaciers, *Journal of Physical Oceanography*, 41, 2279-2294, 2011.
Korona, J., Berthier, E., Bernard, M., Rémy, F., and Thouvenot, E.: SPIRIT. SPOT 5 stereoscopic survey of Polar Ice: Reference Images and Topographies during the fourth International Polar Year (2007–2009), *ISPRS Journal of Photogrammetry and Remote Sensing*, 64, 204-212, 2009.
- 720 Mercer, J. H.: West Antarctic ice sheet and CO₂ greenhouse effect: a threat of disaster, *Nature*, 271, 321, 1978.
Morlighem, M., Rignot, E., Seroussi, H., Larour, E., Ben Dhia, H., and Aubry, D.: A mass conservation approach for mapping glacier ice thickness, *Geophysical Research Letters*, 38, n/a-n/a, 2011.
Morlighem, M., Seroussi, H., Larour, E., and Rignot, E.: Inversion of basal friction in Antarctica using exact and incomplete adjoints of a higher-order model, *Journal of Geophysical Research: Earth Surface*, 118, 1746-1753, 2013.
- 725 Rignot, E., Casassa, G., Gogineni, S., Kanagaratnam, P., Krabill, W., Pritchard, H., Rivera, A., Thomas, R., Turner, J., and Vaughan, D.: Recent ice loss from the Fleming and other glaciers, Wordie Bay, West Antarctic Peninsula, *Geophysical Research Letters*, 32, 2005.
- 730 Rignot, E., Mouginot, J., Morlighem, M., Seroussi, H., and Scheuchl, B.: Widespread, rapid grounding line retreat of Pine Island, Thwaites, Smith, and Kohler glaciers, West Antarctica, from 1992 to 2011, *Geophysical Research Letters*, 41, 3502-3509, 2014.
Rignot, E., Mouginot, J., and Scheuchl, B.: Antarctic grounding line mapping from differential satellite radar interferometry, *Geophysical Research Letters*, 38, L10504, 2011a.
- 735 Rignot, E., Mouginot, J., and Scheuchl, B.: MEASURES InSAR-Based Antarctica Ice Velocity Map [Version1.0], Boulder, Colorado USA: NASA DAAC at the National Snow and Ice Data Center., doi: 10.5067/MEASURES/CRYOSPHERE/nsidc-0484.001, 2011c. 2011c.
Schoof, C.: Ice sheet grounding line dynamics: Steady states, stability, and hysteresis, *Journal of Geophysical Research: Earth Surface*, 112, n/a-n/a, 2007.
- 740 Schoof, C.: Ice-sheet acceleration driven by melt supply variability, *Nature*, 468, 803-806, 2010.
Sergienko, O., Creyts, T. T., and Hindmarsh, R.: Similarity of organized patterns in driving and basal stresses of Antarctic and Greenland ice sheets beneath extensive areas of basal sliding, *Geophysical Research Letters*, 41, 3925-3932, 2014.
Shreve, R.: Movement of water in glaciers, *Journal of Glaciology*, 11, 205-214, 1972.
- 745 Steiger, N., Nisancioglu, K. H., Ákesson, H., de Fleurian, B., and Nick, F. M.: Non-linear retreat of Jakobshavn Isbræ since the Little Ice Age controlled by geometry, *The Cryosphere Discuss.*, 2017, 1-27, 2017.
Thomas, R. H. and Bentley, C. R.: A model for Holocene retreat of the West Antarctic Ice Sheet, *Quaternary Research*, 10, 150-170, 1978.
- 750 Turner, J., Orr, A., Gudmundsson, G. H., Jenkins, A., Bingham, R. G., Hillenbrand, C.-D., and Bracegirdle, T. J.: Atmosphere-ocean-ice interactions in the Amundsen Sea Embayment, West Antarctica, *Reviews of Geophysics*, 55, 235-276, 2017.
Vaughan, D. G. and Doake, C. S. M.: Recent atmospheric warming and retreat of ice shelves on the Antarctic Peninsula, 1996. 1996.

- 755 Walker, C. C. and Gardner, A. S.: Rapid drawdown of Antarctica's Wordie Ice Shelf glaciers in response to ENSO/Southern Annular Mode-driven warming in the Southern Ocean, *Earth and Planetary Science Letters*, 476, 100-110, 2017.
- Weertman, J.: Stability of the Junction of an Ice Sheet and an Ice Shelf, *Journal of Glaciology*, 13, 3-11, 1974.
- 760 Wendt, J., Rivera, A., Wendt, A., Bown, F., Zamora, R., Casassa, G., and Bravo, C.: Recent ice-surface-elevation changes of Fleming Glacier in response to the removal of the Wordie Ice Shelf, Antarctic Peninsula, *Annals of Glaciology*, 51, 97-102, 2010.
- Wouters, B., Martín-Español, A., Helm, V., Flament, T., van Wessem, J. M., Ligtenberg, S. R. M., van den Broeke, M. R., and Bamber, J. L.: Dynamic thinning of glaciers on the Southern Antarctic Peninsula, *Science*, 348, 899-903, 2015.
- 765 Zhao, C., Gladstone, R., Zwinger, T., Warner, R., and King, M. A.: Basal friction of Fleming Glacier, Antarctica, Part A: sensitivity of inversion to temperature and bedrock uncertainty, *The Cryosphere*, companion paper. companion paper.
- 770 Zhao, C., King, M. A., Watson, C. S., Barletta, V. R., Bordoni, A., Dell, M., and Whitehouse, P. L.: Rapid ice unloading in the Fleming Glacier region, southern Antarctic Peninsula, and its effect on bedrock uplift rates, *Earth and Planetary Science Letters*, 473, 164-176, 2017.



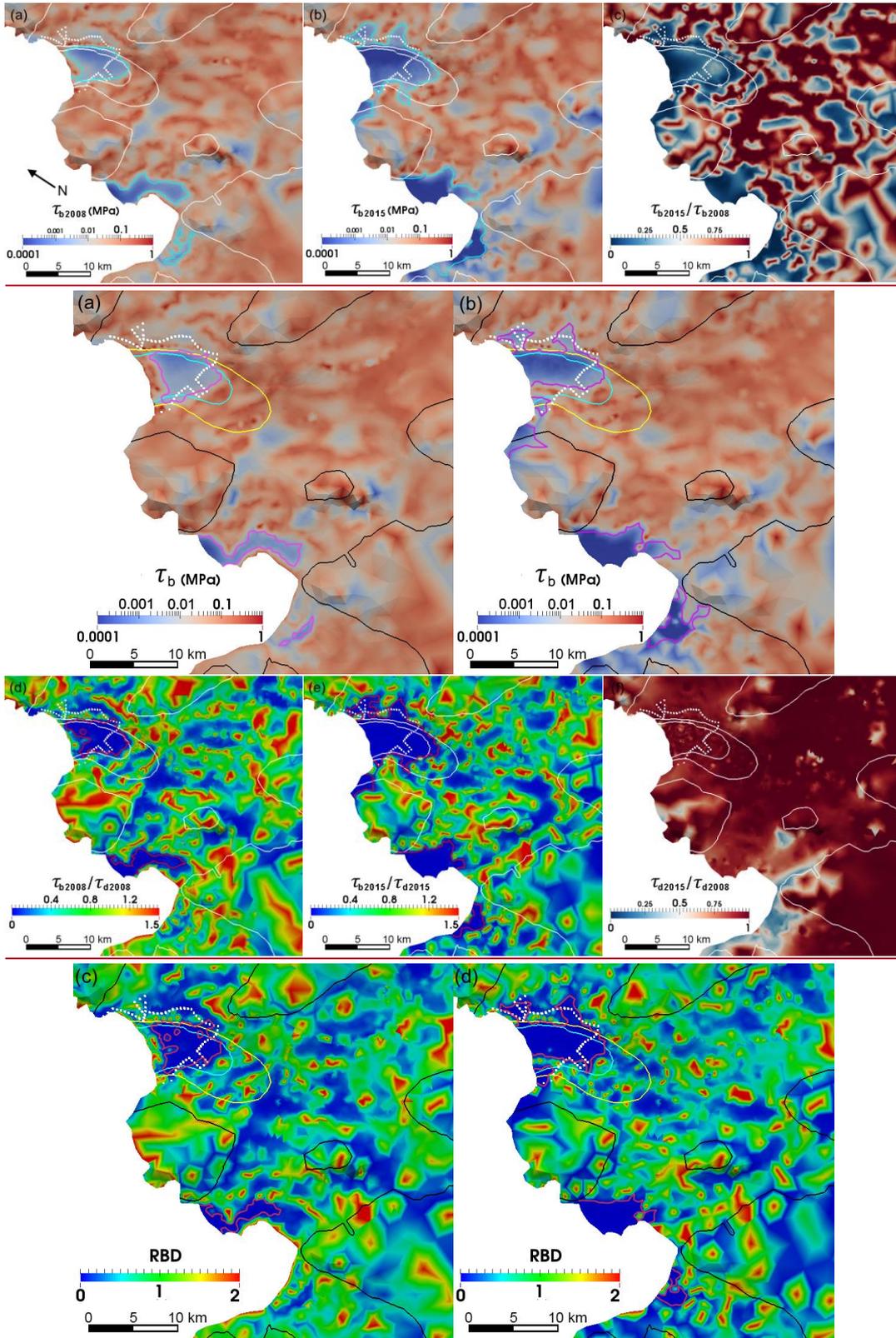
775 Figure 1. (a) The location of the study region in the Antarctica Peninsula (solid line polygon) with bedrock elevation data “bed_zc”, based on BEDMAP2 (Fretwell et al.,
 2013) but refined using a mass conservation method for the fast-flowing regions of the Fleming Glacier system (Zhao et al., companion paper). (b) Velocity changes of the
 780 Wordie Ice Shelf-Fleming Glacier system from 2008 (Rignot et al., 2011c) to 2015 (Gardner et al., 2018). Black contours representing the velocity in 2008 with a spacing of
 500 m yr⁻¹. The colored lines represent the ice front positions in 1947, 1966, 1989, 1997, 2000, 2008, and 2016 obtained from Cook and Vaughan (2010), Wendt et al. (2010), and

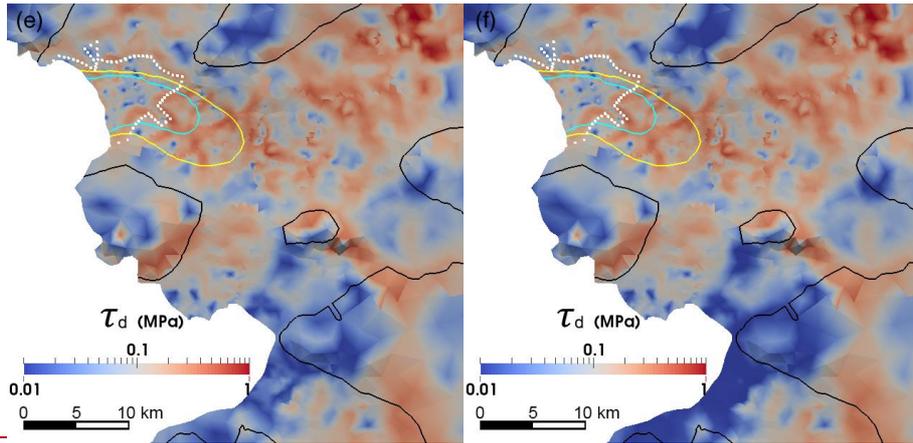
785 Zhao et al. (2017). The feeding glaciers for the Wordie Ice Shelf include three branches:
790 Hariot Glacier (HGL) in the north, Airy Glacier (AGL), Rotz Glacier (RGL), Seller Glacier
(SGL), Fleming Glacier (FGL), southern branch of the FGL (sFGL) in the middle, and
Prospect Glacier (PGL), and Carlson Glacier (CGL) in the south. The grey area inside the
catchment shows the region without velocity data. (c) Inset map of the Fleming Glacier
with ice front positions in 2008 and 2016, grounding line in 1996 (dashed black line) from
Rignot et al. (2011a) and deduced grounding line in 2014 (dashed blue line) from Friedl et
al. (2018). The background image is the bedrock from panel (a) and the black contours
are the same ones as in panel (b).



795 Figure 2. (a) Surface elevation ~~data difference between 2008 and 2015 (2008 minus 2015)~~
 in 2008 ~~(color scale)~~ with black and white contours (interval: 200 m) representing the
 surface elevation in 2008 and 2015, respectively. Inset map shows the location in the
research domain with blue points showing the available elevation data points used to extract
the hypsometric model of elevation change from 2008 to 2015 (Zhao et al., 2017). (b) bed
 800 elevation data “bed_zc” (metres above sea level, masl) with two basins “FGL downstream
 basin” and “FGL upstream basin” from Zhao et al. (companion paper). The black contours
 show the bed elevation with an interval of 100 m. The white contour represents the sea
level used in this study.

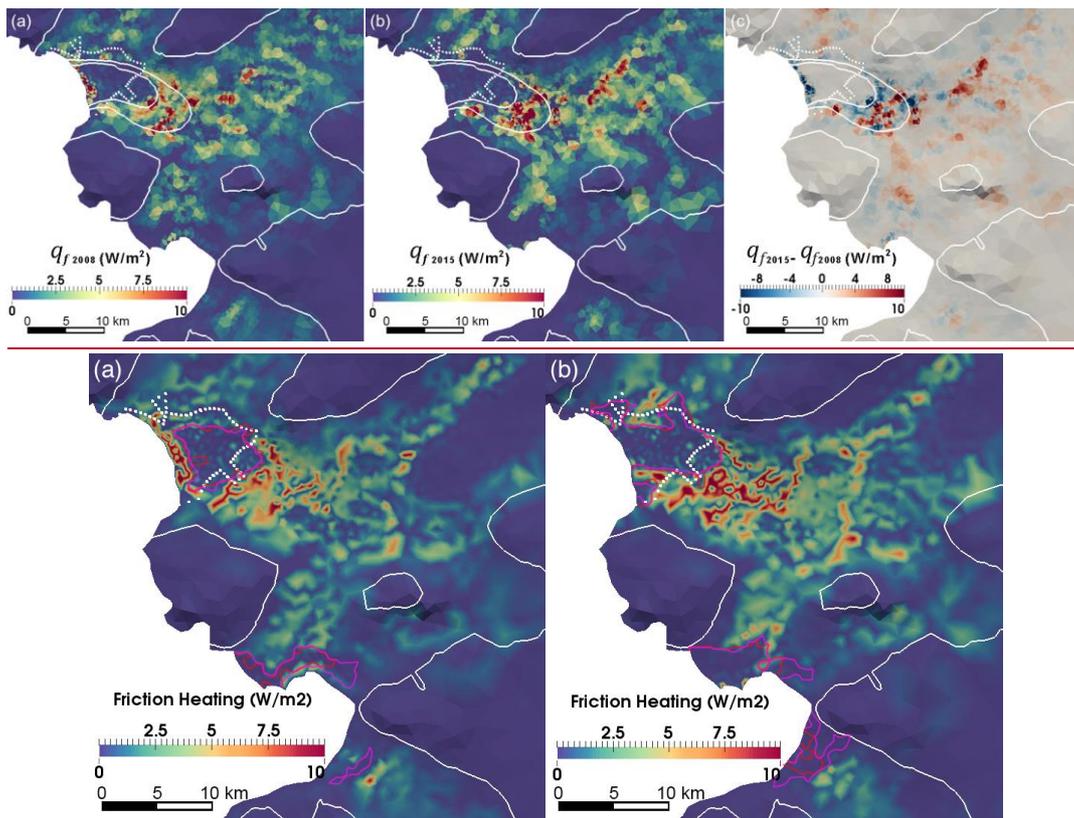
805

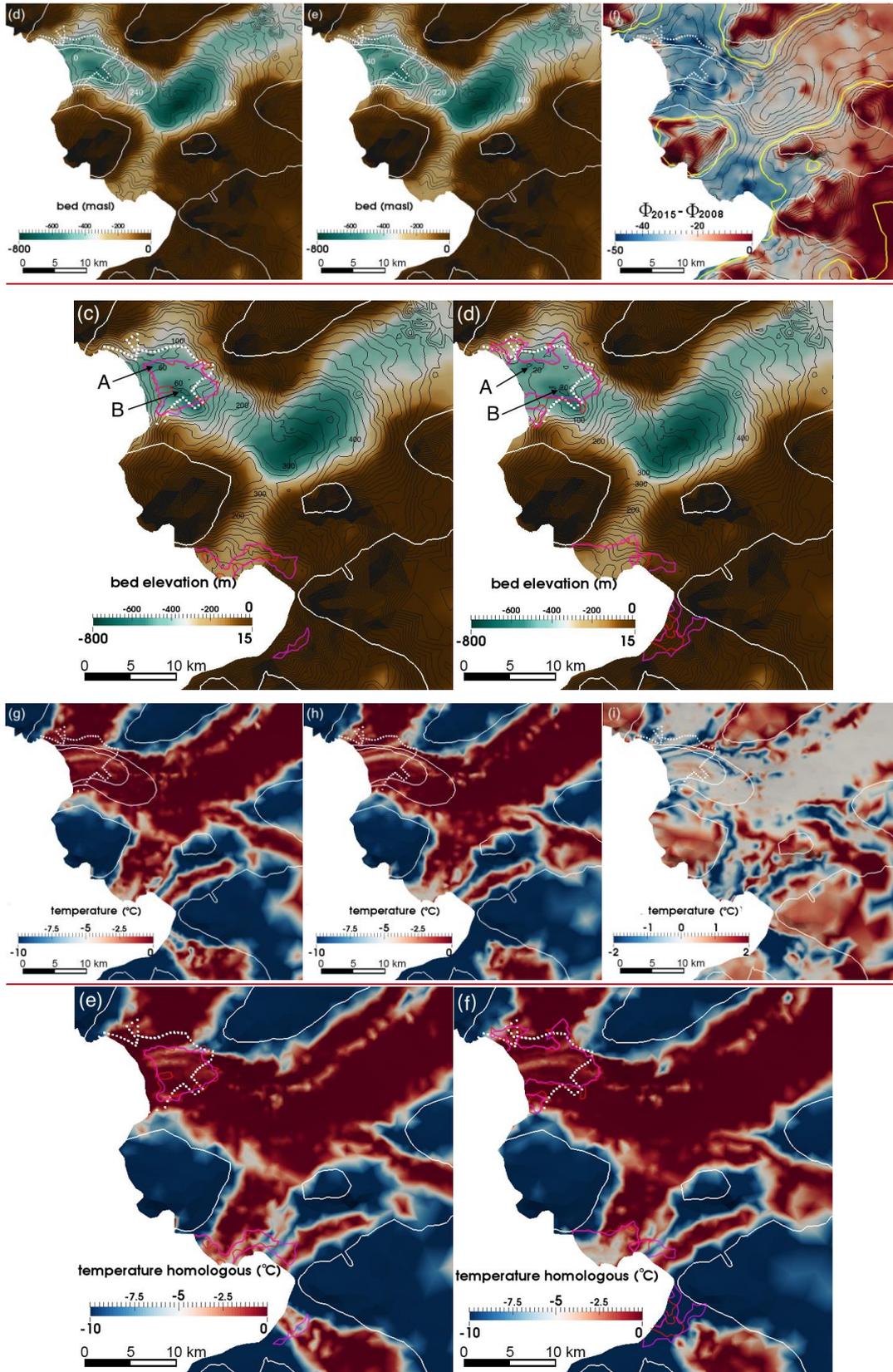




810 Figure 3. (a,b) Basal shear stress τ_b , (e,d,e) the ratio of τ_b to τ_d , and (e, f) the driving stress τ_d of the Fleming Glacier and the Prospect Glacier in 2008 (left) and 2015 (right). (c) the ratio of basal shear stress τ_{b2015} to τ_{b2008} , and (f) the ratio of driving stress τ_{d2015} to τ_{d2008} . The white dotted line represents the deduced grounding line in 2014 from Friedl et al. (2018). The cyan/magenta lines in (a) and (b) shows the boundaries of selected area with $\tau_b \leq 0.01$ MPa in each simulation contour. The red lines in (e) and (f) show the boundaries of selected area with RBD ≤ 0.1 contour in the current study. The black/white, yellow and cyan solid lines represent the 2008 surface speed contours of 100 m yr⁻¹, 1000 m yr⁻¹, and 1500 m yr⁻¹, respectively, to aid visual comparison across subplots. Give additional spatial connections between the figures.

820



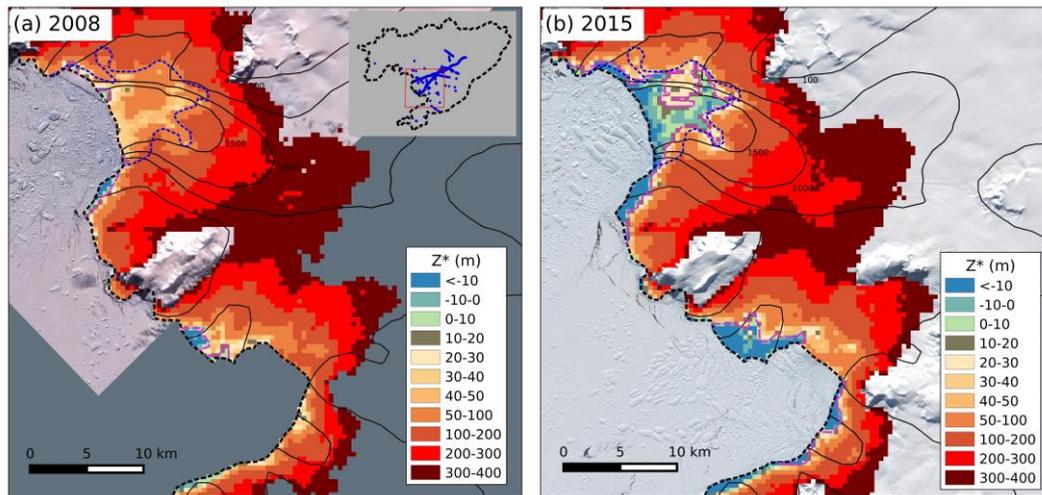


825

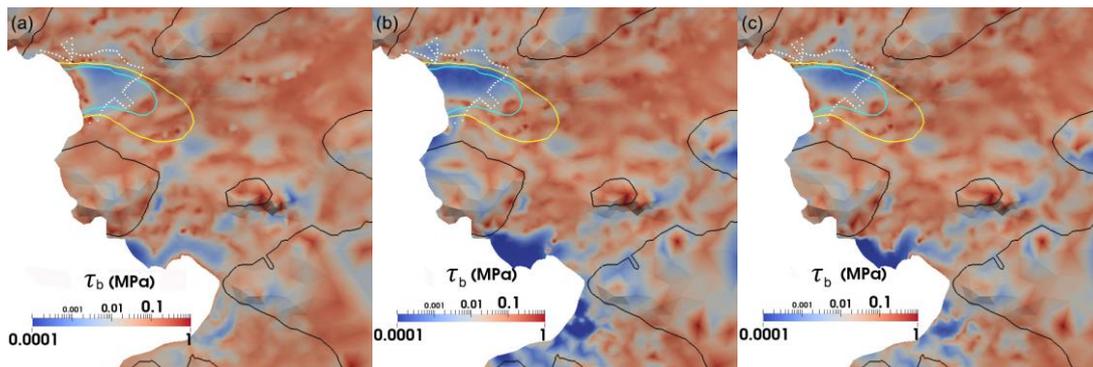
Figure 4. (a, b) The basal friction heating, (ed, de) the contours of hydraulic potential with a spacing of 20 m (black solid lines) with the bed elevation (metres above sea level) as the background, and (eg, fh) the simulated homologous temperature (temperature relative to the pressure melting point) at the base of the Fleming Glacier and the Prospect Glacier in 2008 (left) and 2015 (right/middle). The differences of (c) basal friction heating, (f) hydraulic

830

835 potential, and (i) simulated basal temperature between 2008 and 2015 (2015 minus 2008).
The black contours in (f) represent the bedrock elevation with a spacing of 100 m. The white
dotted line represents the deduced grounding line in 2014 from Friedl et al. (2018). The white
solid lines represents the 2008 surface speed contours of 100 m yr^{-1} , 1000 m yr^{-1} , and 1500 m
 yr^{-1} . ~~The magenta and red solid lines show the boundaries of area with $\tau_b < 0.01 \text{ MPa}$ and~~
~~area with $\text{RBD} < 0.1$, respectively. A and B indicate the location of two over-deepened~~
~~regions in the downstream basin.~~



840 Figure 5. The height above buoyancy Z_* in (a) 2008 and (b) 2015 of the Fleming Glacier and
 Prospect Glacier. The background images are from (a) ASTER L1T data in Feb 2nd, 2009, and
 (b) Landsat-8 in Jan 13th 2016, respectively. The black lines represent velocity contours in
 2008 (Rignot et al., 2011c) ~~and 2015~~. The dashed black and blue lines show the grounding
 line in 1996 (Rignot et al., 2011a) and 2014 (Friedl et al., 2018), respectively. The dashed
 magenta line shows the possible grounding line with $Z_* < 20$ m. Inset map shows the location
 845 in the research domain with blue points showing the available elevation data points used to
 extract the hypsometric model of elevation change from 2008 to 2015 (Zhao et al., 2017).



850 Figure A1. Basal shear stress, τ_b , for (a) 2008, (b) 2015, and (c) a simulation using
 topography from 2008 and velocity from 2015. The white dotted line represents the
 grounding line in 2014 estimated by Friedl et al. (2018). The black, yellow and cyan
 solid lines represent the 2008 surface speed contours of 100 m yr⁻¹, 1000 m yr⁻¹, and 1500 m yr⁻¹,
 respectively.