Response to reviewer comments

First of, we would like to thank both reviewers for excellent and insightful comments that have greatly improved the quality of the manuscript.
A temporary version of two new figures is at the end of this document.

Anonymous Reviewer #1

General Comment:
The sensitivity experiments comprise changes in lake size and different flow law ex- ponents. Lake discharge is also simulated for different lake sizes. Since the results appear to show that the most critical area where changes occur is at the boundary between the lake and the surrounding ice sheet, i.e., a boundary between free slip and high friction, I would have liked to see a sensitivity on the contrast between both. The used friction factor of C=1e13 is a rather high factor and three orders of magnitude larger than the friction one may expect under ice streams. It seems necessary in the general discussion of the observed features, such as the hummocky surface anomaly and the viscous heating at the contact between the lake and the surrounding ice sheet, that the sensitivity is enlarged to different factors of of slipperiness. This would not only make the paper richer, it would also explain features of other types of subglacial lakes encountered in Antarctica (and Greenland). A question that arises is whether at lower C-contrasts the hummocky feature is still prevalent, whether the viscous features play a similar role or not, and if internal layer anomalies show similar characteristics.

Answer: These questions have now been addressed with a new sensitivity study with a lower C-contrast and two different values for ice thickness (1500m as before and 3000m). The new simulations use $C = 10^{10}$, more appropriate for streaming conditions, and a smaller bed inclination ($\alpha = 0.1$). In contrast to simulations previously presented, most of the velocity obtained at the surface arises from sliding at the base. Consequently the transition in velocity from sliding outside and over the lake is much less sharp. Velocities only increase slightly over the lake ($\sim 7\%$) with very little additional strain heating in the transition zone and no temperate layer forming. The deflection of isochrone layers is also very small and hardly noticeable when compared to previous simulations. Two new subfigures will be included in the revised manuscript. One showing surface profiles for the new simulations and the other showing a comparison between steady state isochrone profiles for 3 different simulations with 1500m ice thickness (the new $C = 10^{10}$, and the n=3 and $\eta = 10^{14}$ cases already presented) The dip and ridge features still form but with a much smaller amplitude.

Changes: see new figure 5, page 28 and
see lines 19-28, page 12 in revised version
and lines 01-02, page 13 in revised version

"A weaker contrast in basal traction inside and outside the lake, or a less pronounced switch in flow mode (Leysinger Vieli et al., 2007; Parrenin and Hindmarsh, 2007), decreases the amplitude of the dip and ridge features as can be witnessed in Fig. 5a, where surface profiles for simulations with different sliding coefficients and ice thickness are compared (n = 3). Isochrones relative to the ice base are shown in Fig. 5b for two of the simulations presented in Fig. 4, (n = 3 (blue line) and = 1014 (red line)), in addition to isochrones for a simulation with a lower contrast in basal traction and a lower bed inclination than the other two (cyan line). The structure of internal layers depends heavily on both the contrast in basal traction and the viscosity formulation used. Low contrast results in small deflections of internal layers whereas a fixed viscosity (compared to flow with nonlinear viscosity and a similar surface velocity) results in larger deflections of internal layers due to the generally smaller ice flux at depth outside the lake compared to over it."

plus additional changes in the discussion

lines 6-11, page 14 in revised version

Detailed remarks:

P3860, L8-9: rephrase "A question is what effect this would have on internal ..."

Answer: Changed to "A question that arises is ...."

Changes: see line 08, page 02 in revised version

P3860, L33: past discharge events?

Answer: Changed to "past drainage events"

Changes: see line 23, page 02 in revised version

P3862, L14: The hydrostatic equilibrium of subglacial lakes may well be a function of their size. Since you use a full Stokes model, you can demonstrate whether this is the case for event the smallest lakes sizes in your sample.

Answer: We do use a full-Stokes model but in our model the lakes are already assumed to be in hydrostatic equilibrium. The ice sheets base is flat and the weight of ice everywhere fully supported at the base, including over the lake. This wasn't clear enough in the paper so the text in the manuscript has been changed to include the complete description of the basal stress boundary condition.

Changes: see lines 06-07, page 08 in revised version

P3868, L8: Are 15 layers really enough? Have you tested this with more? Moreover, in the discussion you base your analysis on just two layers with temperate ice (70m). I am afraid that the undersampling may have an influence on the results.

Answer: At present we have tested this with higher vertical resolution (2 times higher at the base, with 4 layers covering the first ($\sim 70m$)), but only solving for enthalpy, keeping the steady state surface and velocity field fixed. These simulations indicate that the thickness of temperate ice is somewhat overestimated in the manuscript at 70m (with just 2 temperate cells), but nevertheless that a temperate layer should exist (for the given surface and velocity field that is). We can be sure of that because even when keeping the temperature fixed everywhere at the pressure melting point it results in the 2nd cell above the base being temperate in the transition zone. This of course does not necessarily mean that there should be a temperate layer present if velocity, surface and enthalpy were all evolving simultaneously, but these are very computationally expensive simulations and I'm not sure that we will be able to do much better. The chosen sliding law has a big effect of course. We did however try to be a bit cautious when interpreting these results and only state in the manuscript that a temperate layer could potentially form (for a real lake) not that it necessarily would. This has been further addressed in section 4.1 where a discussion on the limitations of these results has been included and the wording in the abstract and the conclusions made a bit more cautious.

Changes: see lines 19-23, page 11 in revised version

"The vertical resolution is somewhat limited in the model but simulations with higher resolution do indicate that a temperate layer should form for the given velocity field, although its thickness might be slightly over- estimated. This is strongly affected by the chosen sliding law and sliding coefficients and should not be considered representative for subglacial lakes in general."

as well as modified lines in the abstract and conclusions

see lines 15-17, page 02 in revised version

and lines 22-24, page 17 in revised version

P3868, L25: What considerable distance? be precise. Is this so-called considerable distance a function of ice thickness/size of lake? Does slipperiness play a role?

Answer: What we mean here by a considerable distance is around 50km for the L_M size. And yes, it is a function of all of the above; ice thickness, lake size and slipperiness outside the lake. In general, the influence of membrane stresses, or longitudinal coupling increases with increased slipperiness, ice thickness and lake size (Cuffey and Paterson 2010, Hindmarsh 2006, Kamb and Echelmeyer 1986), but the effect of such a transition in basal friction as the one we model here also diminishes as the difference in friction outside and inside the lake decreases (the C-contrast) as can be witnessed by the new sensitivity studies that you proposed. The wording in the manuscript has been changed in order to further elucidate these facts and new citations added.

Changes: see lines 27-28, page 10 in revised version

P3869, L16: I suppose that this concentrated deformation is very much a function of the basal sliding law and coefficients used outside the lake. The value of C is also very high (3 orders of magnitude higher than that of an ice stream) which implies a sharp transition between the ice flow across the lake and the surrounding ice. See my general remark.

Answer: True. It is quite high. It was meant more as representative for areas at the onset of ice streaming, although we do state in one place in the paper (end of the introduction) that the aim was to model a lake underneath an ice stream. This was an oversight. Nevertheless, it is still maybe 1-2 orders of magnitude too large. Higher values than what you propose might still be justifiable though because the bed surface is completely flat, unlike any real bed surface which would be highly irregular, with obstacles of all sizes. Therefore, no resistance is provided by "large" obstacles (compared to the grid size) over which the ice would have to deform and which would add to the available energy for internal heating. Your point is a good one though, which is why we decided to do the sensitivity experiment you suggest. The results indicate that if you lower the C-contrast such that the majority of surface velocity is derived from basal sliding then the increase in surface velocity over the lake is relatively small ($\sim 7\%$) for the case presented, and that whatever strain heating that occurs in the transition zone is not enough to create a temperate layer. Two new figures are presented in the revised MS, one showing surface profiles for the simulations with lower C-contrast and the second showing steady state isochrone layers for low and high C-contrasts and the fixed viscosity case.

Changes: see answer to general remark

P3870, L8: Could the dip and ridge feature also be a function of bridging stresses in the full Stokes, besides a strain-induced flow-law effect?

Answer: It is possible that bridging stresses play a role, especially for small lakes, but it's not something that we can shed much light on without changing the model set-up as the lakes are assumed to be in hydrostatic equilibrium and the ice column fully supported everywhere by the normal stress at the bed.

Changes: no changes

P3870, L27: The travelling wave is surely an interesting phenomenon and would have had more value if this has also been detected in radargrams. I understand that it is probably difficult to detect when topography is rough. This would certainly be an added value to the paper.

Answer: Yes, we agree. Unfortunately we are not aware of any study where this has been detected and documented. In most cases we would imagine that previous drainage events would be difficult to detect and probably not immediately obvious from radargrams alone, as explained in the manuscript. Most likely a numerical model would be needed in order to isolate the effect of any past drainage on isochrones from the effect of varying rheology and basal topography.

Changes: addressed in chapter 4.3

P3871, L18: See previous remark on undersampling (2 vertical cells seems to me very narrow to base conclusions on).

Answer: see previous remark for answer

P3872, L13-14: remove both commas

Answer: commas removed

Changes: see line 1-2, page 15 in revised version

P3872, L24: change "everywhere the same" in "constant".

Answer: changed as suggested

Changes: see line 13, page 15 in revised version

P3873, L6: The sharp transition in sliding velocity is not tested. It is only an experiment with a sharp transition and should be compared to a less sharp transition to make this hard.

Answer: We have now compared two cases with different C-contrasts and included a new subfigure in the manuscript. For lower C-contrasts the amplitude of the dip and ridge features is much smaller and very little additional strain heating is introduced at the lake edges. No temperate layer forms. See answer to general comment.

Changes: see answer to general remark

P3874, L24: replace "so this", by "which"

Answer: changed as suggested

Changes: see line 13, page 17 in revised version

G.J.-M.C. Leysinger Vieli (Reviewer #2)

Specific Comment:
For the first experiment suite, a steady state experiment, the resulting surface velocities and topography as well as the internal layers are obtained for different sized lakes. The effect of a different viscosity on the surface topography and the vertical distribution of horizontal velocity is studied. However we do not see how a different viscosity affects the internal layers. If there is no change this should be mentioned in the text.

Answer: There is a considerable difference, especially between the case with a fixed viscosity versus the other two. The n=1 and n=3 cases result in very similar ratios of vertical to horizontal velocity (w/u) in the transition zone between little to full sliding. The fixed viscosity case results in much higher ratios, especially at depth, and consequently a much larger downward displacement of internal layers. This can also be witnessed in Fig. 4a where the surface of the fixed viscosity case has been displaced considerably more vertically than the other two, although no dip/ridge feature can be discerned. The temperature and pressure dependent viscosity concentrates velocity changes in the vertical to the lower layers of the ice sheet so a large part of the ice column has a horizontal velocity close to the surface one. A typical vertical velocity profile with fixed viscosity results in a much more gradual change in horizontal velocity with depth so that within the ice sheet, although the surface velocity for the 2 cases is similar, the velocity at depth for the fixed viscosity case will be lower along with the total mass flux. Regardless of viscosity, the vertical distribution of horizontal velocity over the lake is pretty much everywhere the same, uniform. So at depth for the fixed viscosity case, the ratio of vertical to horizontal velocity will be much greater for a particle of ice traversing the transition zone than in the n=3 case and isochrones therefore displaced further downward. Referring to your own work (Leysinger Vieli et al. 2007) and that of Parrenin and Hindmarsh (2007) we can consider the difference between flux shape functions for n=3 and a fixed viscosity and how the maximum step in isochrone elevation is proportional to the change in flux shape function in the transition from internal deformation to sliding. For the fixed viscosity case, the flux increases much more gradually with height and the maximum step in isochrone elevation will therefore be larger. The revised manuscript has been complemented with a new subfigure, showing a comparison between steady-state isochrone profiles for two cases previously presented (n=3 and $\eta = 10^{14}$) in addition to profiles from a new simulation with the same ice thickness but a smaller friction factor ($C = 10^{10}$), more appropriate for streaming conditions.

Changes: see new figure 5b and

see lines 19-28, page 12 in revised version

and lines 01-02, page 13 in revised version

"A weaker contrast in basal traction inside and outside the lake, or a less pronounced switch in flow mode (Leysinger Vieli et al., 2007; Parrenin and Hindmarsh, 2007), decreases the amplitude of the dip and ridge features as can be witnessed in Fig. 5a, where surface profiles for simulations with different sliding coefficients and ice thickness are compared (n = 3). Isochrones relative to the ice base are shown in Fig. 5b for two of the simulations presented in Fig. 4, (n = 3 (blue line) and = 1014 (red line)), in addition to isochrones for a simulation with a lower contrast in basal traction and a lower bed inclination than the other two (cyan line). The structure of internal layers depends heavily on both the contrast in basal traction and the viscosity formulation used. Low contrast results in small deflections of internal layers whereas a fixed viscosity (compared to flow with nonlinear viscosity and a similar surface velocity) results in larger deflections of internal layers due to the generally smaller ice flux at depth outside the lake compared to over it."

In the second experiment, a transient experiment, the authors look at how the internal layers evolve under a lake drainage happening over 10 years (compared to the ice velocity this is seen as an instantaneous drainage). However the lake surface is not a free surface in the vertical direction as it is fixed to the bed plane and only changes in the horizontal extend. What effect does this have on the internal layering?

Answer: If you were to isolate changes in the vertical, keeping the horizontal extent of the lake fixed we believe that the effect on isochrones should be much smaller than with changes in the horizontal. Such an experiment has been performed previously by Sergienko et al. (2007), although not looking specifically at isochrones. Their experiment resulted in little changes in horizontal velocity ($< 1\%$) and the ice surface reached its initial stage in about 20 years (there drainage occurred during 5 years). Its not unreasonable to believe that the vertical velocity (they used a 2D vertically integrated model) would also experience little changes as the horizontal one and therefore result in only minor disturbances to internal layers. As the lake boundary stays fixed in the horizontal, the points of strong vertical velocity (at the upstream and downstream ends of the lake) are also fixed and for there to be a strong change in isochrone pattern, there would have to be a strong temporally persistent change in the ratio between vertical and horizontal velocities (w/u) as a particle of ice traverses the transition zone, which is unlikely. This has now been addressed in more detail in section 3 (model experiments)

Changes: see lines 13-17, page 10 in revised version

"Sergienko et al. (2007) found that vertical movement of lake surfaces due to drainage resulted only in small changes to ice velocity and that the ice surface reached its initial stage after just a few years, indicating that it would not lead to a significant disturbance in isochrone structure compared to changes in the horizontal lake extent"

Furthermore it is not clear to me if the water that drains from the lake has been accounted for the ice downstream of the lake, which would affect the internal layering, e.g. by melting ice.

Answer: The water "draining" from the lake has not been accounted for no, and this would certainly affect the internal layering downstream of the lake by for instance, as you say, melting ice or potentially influencing ice velocities downstream. Channelized flow would induce some melting over a limited area but this would most likely be in the order of a few meters at most and therefore only have a small influence on the travelling wave, which in our case has an amplitude of around 100m. A more distributed drainage (Stearns et al. 2008) might potentially have a stronger effect with increased ice velocities downstream which would affect how fast layer disturbances would flatten and become undetectable. This has now been further clarified in section 3 (model experiments).

Changes: see lines 19-25, page 10 in revised version

"Water draining from subglacial lakes is likely to have an impact on ice dynamics and isochrone structures downstream of the lake, either through melting of ice or changes in basal water pressure and sliding, although no representation of subglacial hydrology is included here. The effect the draining water will have on isochrone layers will depend on the state of the drainage system and how the water is transported downstream (Stearns et al., 2008) but will likely be limited by the swiftness of the ice response compared to the time needed to considerably perturb isochrone layers."

Nevertheless the drainage experiment is interesting, but to me it is not entirely clear if the comparison with the work by Wolovick et al. (2014), where a slip boundary is moving WITH the ice can be compared to the downstream boundary of the draining lake that is moving UPSTREAM. I feel that there the interpretation of the observed result is going too far and is too speculative.

Answer: Yes, that could have been worded a bit more carefully. The intention was not to directly compare our work to the work of Wolovick et al. (2014), merely to note that slip boundaries or slippery patches moving with the ice (such as our upstream lake boundary) are capable of distorting isochrone layers to a much greater extent that stationary ones and that for maximum effect the boundary should be moving at a velocity comparable to the averaged ice column velocity. In our case the upstream lake boundary is moving too fast to cause any significant upward deflection of isochrones during the drainage. The paragraph has been reworded to avoid confusion.

Changes: see lines 14-18, page 16 in revised version

"Only slip bound- aries moving with the ice are capable of distorting layers to a greater extent than stationary boundaries. Boundaries moving in opposite directions to ice flow, like our downstream lake boundary, will have a smaller impact on internal layers than a stationary boundary as the relative horizontal velocity increases."

However, the discussion under which conditions the modelled signature in the internal layers - a travelling wave - might be observable with radar is good. The question however is how long is it visible? I find Figure 5 not yet that clear in showing the travelling wave. Maybe you could add a later time frame or show the last graph with the initial internal layers to better visualise the contrast.

Answer: We decided against showing the wave at a later stage because it becomes distorted and smoothed as it moves downstream. In order to properly simulate its advection downstream, a grid resolution comparable to the one used at the lake edges would have been needed downstream of the lake, which would have required much longer computation times, too long to be feasible. As it is now, the grid resolution is kept reasonably small for roughly two lake lengths in the downstream direction, but still far from the grid size used at the lake edges. The travelling wave is basically visible in the model until it reaches the downstream boundary but the distortion that arises is mostly due to the coarse resolution (compared to what would be needed). We can however display the internal layers at t=2000a as you suggest with the initial layer configuration and with an arrow pointing to it and this will be done in the revised version of the MS.

Changes: see modified version of figure 6 (formerly figure 5) where both the initial isochrone layers and a red arrow has been added to the final timeslice.

Technical corrections:

P3860, L2: odd wording: subsequent draw-down of isochrones and cold ice from the surface. An increase in velocity leads to a thinner ice body because its faster. The cold ice is still at the surface, but the temperature gradient does change. The maximum change in elevation of the isochrones is found at about a third of the ice thickness (see Leysinger Vieli et al., 2007) - this is where you would find the largest effect of temperature change - but its not surface ice. What you mean here is the Weertman effect. But this becomes only clear later in the text.

Answer: Somewhat confusing, yes. The ice originally at the surface stays at the surface of course as you rightfully point out but colder ice does get drawn down from above, but not directly from the surface, although the surface is also deflected downward at the upstream lake edge.

Changes: "from the surface" removed

see line 02, page 02 in revised version

P3860, L31: what is rapid?

Answer: This has been reworded to "We also conclude that rapid changes in the horizontal extent of subglacial lakes and slippery patches, compared to the average ice column velocity, create a travelling wave...."

Changes: see lines 20-24, page 02 in revised version

P3862, L14: Comma after hydrostatic equilibrium?

Answer: comma added

Changes: see line 07, page 04 in revised version

P3862, L19: Leysinger Vieli without hyphen, studied the effect of areas with basal sliding or melting on internal layer architecture, but not explicitly a subglacial lake.

Answer: Changed as suggested.

Changes: see line 11, page 04 in revised version

P3864, L3-5: could be written a bit clearer as you explain terms in an equation of an equation. Maybe its easier to explain it after each other?

Answer: reordered such that one is explained after the other

Changes: see lines 19-03, page 05-06 in revised version

P3864, Equation 8: I believe here is something missing e.g. $H > ....$

Answer: The conditions on which the conduction term depend on are explained in the paragraph following the equation. For clarity they have been added to the equation in the revised MS.

Changes: see equation 8 or line 17, page 06 in revised version

P3867, L21: Is the ice thickness realistic for Greenland or Antarctica? What effect does the ice thickness have on the result?

Answer: The ice thickness used in the model is relatively low for Antarctic lakes, that tend to be underneath around $\sim$ 3km thick ice (Siegert 2000), and high considering the only two identified subglacial lakes in Greenland (to date) that are both underneath around 700-800m of ice (Palmer et al. 2013). So it is somewhere in between. The study has now been complemented with a new parameter study where we look at how the deflection of internal layers and surface profiles are affected by a lower friction coefficient ($C = 10^{10}$), more appropriate for streaming conditions, and ice thickness. Increasing ice thickness leads to more strain heating at the base so the inclination of the bed plane was lowered to 0.1 deg for the new simulations. The dip and ridge features show a higher amplitude with thicker ice but this cannot be contributed solely to ice thickness as basal shear stresses increase as well. A new figure has been added to the MS showing surface profiles for the new parameter study and new paragraphs in the discussion have been added addressing the new results and related issues.

Changes: see new figure 5.a

P3869, L8-12: Refer to Figure 2b?

Answer: reference to figure 2b added

Changes: see line 13, page 11 in revised version

P3870, L10: In order to know what viscosity is used in the other experiments one needs to look it up in Table 1, but it is never referred to in the text.

Answer: references added

Changes: see lines 7-9, page 10

"All presented results are based on simulations with a fully nonlinear ice viscosity (n = 3), unless explicitly stated otherwise and all numeri- cal values for model constants are defined in Table 1."

in addition to several other references throughout the manuscript (figure captions, general text) where information about which viscosity case is being referred to has been added.

P3870, L26-28: Can you visualise the traveling wave better, e.g. at a later stage when the isochrones around the lake fully recovered?

Answer: see previous answer

Changes: see new version of figure 6 (formerly figure 5)

P3872, L9-11: Is this so? Why?

Answer: Well, maybe this was a somewhat stronger statement than the results actually allow for. The lake size experiment does result in a slight difference in downward displacement of internal layers, but it's only around 3% at most between the largest (L_S) size and the smallest (L_S) lake size. It could be an effect of the periodic boundary conditions. To avoid confusion and misconception, that sentence has been removed!

Changes: sentence removed (see lines 27, page 14 in revised version)

P3872, L15: Here you describe the Weertman effect.

Answer: True

P3873, L20: Both boundaries move but not in the same direction. It is rather a narrow- ing of the area.

Answer: True. The paragraph was reworded for clarity. The following text was added: "Only slip bound- aries moving with the ice are capable of distorting layers to a greater extent than stationary boundaries. Boundaries moving in opposite directions to ice flow, like our downstream lake boundary, will have a smaller impact in internal layers than a stationary boundary as the relative horizontal velocity increases."

Changes: see lines 14-18 page 16 in revised version

P3873, L23: affects instead of effects.

Answer: changed as suggested

Changes: see lines 11 page 16 in revised version

P3873, L24-23: Im not sure you can compare this to Wolovicks et al. (2014) work, as your downstream boundary is moving in opposite direction to the ice.

Answer: The paragraph has been reworded to highlight the differences between our moving boundaries and the ones in the work of Wolovick et al (2004). Such boundaries moving in opposite directions to ice flow have a smaller integrated effect on isochrones as the relative velocity increases and the points of vertical flow have less time in "action" compared to stationary boundaries.

Changes: see previous answer

see lines 14-18 page 16 in revised version

P3877, L24: Leysinger Vieli remove hyphen.

Answer: Fixed

Changes: see modified reference list

P3880: Refer to Table 1 in the text.

Answer: fixed

Changes: References added on:
see line 8,22 page 6 in revised version
and line 13, page 8 in revised version
and line 9, page 10 in revised version

P3883: Fig. 2b - not sure what colour-bar is for what. Are both colour-bars / colour- scales true everywhere or is one for the inlet figure only? Caption: Maybe along flow profile is a better expression than cross-sectional? I was always thinking at an across profile. Mention what the black lines are (its mentioned in the text but would be useful information in the caption too.) Mention in the caption the lake size you are showing.

Answer: Both color bars are true for Figure 2b including the inlet figure. The inlet figure is simply a zoom-in of the larger figure with the same colorscale. "Cross-sectional" replaced with "along flow profile" and a reference to what the black lines mean and what lake size (L_M) we're showing was added to the caption.

Changes: See modified Figure 2 caption: page 25 in revised version

P3884: Caption: again I find along flow profile over the lake easier to understand. Replace vertical bar with horizontal bar Not clear with what viscosity case this has been calculated (n=3?).

Answer: Replaced with "along flow profile" as suggested previously. "vertical bar" replaced with "horizontal bar". These were calculated with (n=3) as you correctly guess. That information was added to the caption.

Changes: See modified Figure 3 caption: page 26 in revised version

P3885: Caption: You are showing the vertical profile of the scaled horizontal velocity but in a way you are saying it but as it reads I understood it for c) only.

Answer: changed to "Vertical profiles of (b) the scaled horizontal velocity and (c) the logarithm of ice viscosity, over the center of the lake."

Changes: See modified Figure 4 caption: page 27 in revised version

P3886: Figure: The grey colour-bar is a bit odd - not clearly visible in the figure. Caption: what do you mean by yellow lines? They are not visible. With black lines you mean the grey scale colors? Mark the travelling wave - e.g. by an arrow, or show the initial isochrones together with the isochrones of the t=2000 case. Or could you show an even later stage?

Answer: The gray color-bar is not really essential. We kept it in for completeness but it could be removed. We are not really concerned with the actual age of ice in the model, just the geometry of the internal layers. This information has been added to the results section. The sentence mentioning the yellow lines was a reference to an older version of the figure and has been removed. Initial layers and an arrow has been added to the final time frame (t=2000a). With black lines we were referring to the gray-black color that denotes isochrones at different ages.

Changes: see modified figure 6 caption (formerly figure 5): page 29 in revised version

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Manuscript prepared for The Cryosphere Discuss.
with version 2014/09/16 7.15 Copernicus papers of the LaTeX class copernicus.cls.
Date: 14 December 2015

The influence of a model subglacial lake on ice dynamics and internal layering

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Abstract

As ice flows over a subglacial lake, the drop in bed resistance leads to an increase in ice velocities and a ~~subsequent~~ draw-down of isochrones and cold ice~~from the surface~~. The ice surface flattens as it adjusts to the lack of resisting forces at the base. The rapid transition in velocity induces changes in ~~temperature and ice viscosity , releasing deformation energy which raises~~ ice viscosity and releases deformation energy that can raise the temperature locally. Recent studies of Antarctic subglacial lakes indicate that many lakes experience very fast and possibly episodic drainage, during which the lake size is rapidly reduced as water flows out. A question that arises is what effect this would have on internal layers within the ice, and whether such past drainage events could be inferred from isochrone structures downstream.

Here, we study the effect of a subglacial lake on ~~the dynamics of a model ice stream~~ ice dynamics as well as the influence that such short timescale drainage would have on the internal layers of the ice. To this end, we use a Full–Stokes, polythermal ice flow model. An enthalpy gradient method is used to account for the evolution of temperature and water content within the ice.

We find that ~~the~~ a rapid transition between slow-moving ice outside the lake, and full sliding over the lake, ~~releases large~~ can release considerable amounts of deformational energy, ~~which has~~ with the potential to form a temperate layer at depth in the transition zone. In addition, we provide an explanation for a characteristic surface feature, commonly seen at the edges of subglacial lakes, a hummocky surface depression in the transition zone between little to full sliding. We also conclude that rapid changes in ~~lake geometry or basal friction~~ the horizontal extent of subglacial lakes and slippery patches, compared to the average ice column velocity, can create a travelling wave at depth within the isochrone structure that transfers downstream with the advection of ice, thus indicating the possibility of detecting past drainage events with ice penetrating radar.

1 Introduction

Nearly 400 subglacial lakes have been identified in Antarctica based on radar data and satellite measurements (Wright and Siegert, 2012), as well as at least two subglacial lakes in Greenland, all of which satisfy established criteria for identification of subglacial lakes (Palmer et al., 2013). Subglacial lakes are typically either located in the ice-sheet interior, close to ice divides where surface slopes are low and ice is thick, or in areas of higher ice velocity where internal ice deformation and sliding add to geothermal energy to produce melting at the base (Siegert et al., 1996; Dowdeswell and Siegert, 1999).

Subglacial lakes have been hypothesized to play a role in the initiation of fast flow by either (1) providing a steady stream of water downstream of their location, lubricating the base (Gray et al., 2005; Langley et al., 2014), (2) substantially influencing the thermal regime of the ice, gradually warming the basal ice from below, or (3) providing ice with enough thermal momentum to resist freezing on downstream of the lake (Bell et al., 2007).

The reason why so few lakes have been discovered in Greenland, despite considerable efforts and relatively dense surveying compared to Antarctica, is believed to be because of the generally warmer ice and higher surface slopes in Greenland, which favour rapid and more efficient drainage as well as increased vulnerability to drainage instabilities (Pattyn, 2008). In addition, the large supply of surface meltwater to the base of the Greenland ice sheet through hydrofracturing means that the drainage networks in place are probably already highly efficient, and thus capable of effectively draining subglacial water and preventing or limiting subglacial lake development (Palmer et al., 2013).

Apart from lake Vostok, the largest subglacial lake in the world as well as a few other ones, the typical subglacial lake is around 10 km in diameter (Siegert, 2000). Recent studies have presented compelling evidence of rapid transport of water stored in subglacial lakes indicating that lakes can either drain episodically, or transiently on relatively short time scales (Gray et al., 2005; Wingham et al., 2006; Fricker et al., 2007). Lakes appear to form a part of a connected hydrological network, with upstream lakes draining into downstream ones as subglacial water moves down the hydrological potential (Siegert et al., 2007). Both ice and water transport a substantial amount of sediment over time which is deposited in the lakes, thus reducing their size over long time scales. Sedimentation rates can vary from close to zero to several millimetres per year, with sediment layers estimated to be up to several hundreds of meters thick in some lakes (Christoffersen et al., 2008; Bentley et al., 2013). The dominant mechanism in transporting sediment to subglacial lakes is thought to be influx of sediment-laden water for open-system, or active lakes, and melt-out from the overlying ice for closed-system lakes, where water exchange happens purely through melting and freezing (Bentley et al., 2013). Subglacial lakes thus change size on a variety of time scales with different mechanisms, from fast drainage to slow sedimentation.

The presence of a subglacial lake is often signalled by a flattening of the ice surface, given that the lake is in hydrostatic equilibrium, and as a local speed-up of ice velocities. This draws down cold ice and deflects isochrone layers within the ice (Weertman, 1976).

Numerous studies have investigated the effect of spatially varying basal (or surface) conditions on ice dynamics and the internal layering of ice, such as melting or basal resistance. Previously, Leysinger Vieli et al. (2007) studied the effect of areas with basal sliding or melting on internal layer architecture, but not explicitly a subglacial lake~~on isochrones by prescribing the flow and then solving for the age of ice~~. The effect of a subglacial lake on ice dynamics was investigated by Sergienko et al. (2007), who used a 2-D vertically-integrated flow equation to study ice sheet response to transient changes in lake geometry and basal resistance, and by Pattyn (2008) who investigated the stability of a subglacial lake to drainage events. The interaction between lake circulation and ice dynamics and its effect on the basal mass balance has also been studied by Thoma et al. (2010).

The aim of this study is to investigate the influence of a subglacial lake on ~~the dynamics of a model ice stream~~ ice dynamics and the effect it has on the isochrone structure within the ice. As all stress components are important for such an interaction, we use a fully 3-D thermo-mechanically coupled Full Stokes ice sheet model, implemented in the commercial finite element software COMSOL. We employ an enthalpy-gradient method to account for the softening effect of ice temperature and water content on ice viscosity (Aschwanden et al., 2012) and we show how temporally varying basal conditions can lead to the appearance of flow bands, or arches and troughs, within the internal layering downstream of the original flow disturbance.

2 Model description

2.1 Ice flow

Ice is treated as an incompressible fluid with constant density, obeying conservation laws for mass and momentum:

$$\nabla \cdot \boldsymbol{u} = 0 \tag{1}$$

and

$$\nabla \cdot \boldsymbol{\sigma} = -\rho \boldsymbol{g}, \tag{2}$$

where $\boldsymbol{u}$ is the velocity vector, $\boldsymbol{g}$ the gravitational acceleration, ρ the ice density and $\boldsymbol{\sigma}$ the Cauchy stress tensor. The Cauchy stress tensor is given by:

$$\boldsymbol{\sigma} = \boldsymbol{\tau} - p\mathbf{I}, \tag{3}$$

where $\boldsymbol{\tau}$ and $p\mathbf{I}$ are the deviatoric and the isotropic parts, p is pressure, and $\mathbf{I}$ is the identity matrix. Inertial forces are assumed negligible and only body forces arising from gravity are taken into account.

Ice is assumed to follow Glen's flow law (Steinemann, 1954; Glen, 1955), in which deviatoric stresses are related to strain rates ($\dot{\varepsilon}$) by:

$$\boldsymbol{\tau} = 2\eta\dot{\boldsymbol{\varepsilon}} \tag{4}$$

and

$$\eta(T', W, \dot{\varepsilon}_{\mathrm{e}}) = \frac{1}{2}[A_{\mathrm{t}}(T', W)]^{-\frac{1}{n}} \dot{\varepsilon}_{\mathrm{e}}^{\frac{1-n}{n}}, \tag{5}$$

where η is the effective viscosity, $\dot{\varepsilon}_\mathrm{e}$ is the effective strain rate, ~~$T' = T + \gamma p$ is~~ (A_t) is a rate factor that depends on the homologous temperature ~~which corrects for the dependence of the pressure-melting point (T_m) on pressure, γ is the Clausius–Clapeyron constant,~~ (T') and the water content of the ice (W) and n is the power-law exponent in Glen's flow law. The ~~rate factor (A_t) depends on temperature (T) , pressureand the water content (W) of the ice as~~ homologous temperature $T' = T + \gamma p$ corrects for the dependence of the pressure-melting point (T_m) on pressure, where γ is the Clausius–Clapeyron constant. The formulation of the rate factor follows (Duval, 1977):

$$A_\mathrm{t}(T,p,W) = A(T,p) \times (1 + 1.8125W[\%]), \tag{6}$$

where

$$A(T,p) = A(T') = A_0 e^{-Q/RT'}, \tag{7}$$

and A_0 is a pre-exponential constant, Q is the activation energy and R is the universal gas constant, with numerical values listed in Table 1 (Greve and Blatter, 2009).

2.2 Enthalpy balance

An enthalpy-gradient method (Aschwanden et al., 2012) is employed, as opposed to the typically used cold-ice formulation, which is incapable of correctly reproducing the rheology of temperate layers within ice sheets. The enthalpy formulation allows for the possibility of including liquid water content within temperate ice, based on mixture theory, without explicitly tracking the cold/temperate transition surface (Aschwanden et al., 2012; Greve, 1997). In the enthalpy-gradient method, enthalpy replaces temperature as the thermodynamical state variable, such that:

$$\rho \left(\frac{\partial H}{\partial t} + \boldsymbol{u} \cdot \nabla H \right) = \boldsymbol{\nabla} \cdot \left(\left\{ \begin{array}{cc} K_\mathrm{i}(H) \nabla H & \text{if } H < H_{T_\mathrm{m}}(p) \\ k(H,p) \nabla T_\mathrm{m}(p) + K_0 \nabla H & \text{if } H \geq H_{T_\mathrm{m}}(p) \end{array} \right\} \right) + Q, \tag{8}$$

where $H = H(T, W, p)$ is the temperature and water content dependent specific enthalpy ~~,~~ ~~T_m the pressure melting point,~~ and $Q = 4\eta\dot{\varepsilon}_\mathrm{e}^2$ is the heat dissipation due to internal deformation. The conduction term in Eq. (8) depends on whether the ice is cold ($H < H_{T_\mathrm{m}}(p)$) or temperate ($H \geq H_{T_\mathrm{m}}(p)$). The conduction coefficient for cold is ice defined as $K_\mathrm{i} = k/c$, where k is the thermal conductivity and c is the heat capacity, both assumed to be constant (Table 1). $H_{T_\mathrm{m}}(p)$ is the specific enthalpy of the pressure dependent melting point of ice. The diffusivity for temperate ice is poorly constrained as little is known about the transport of microscopic water within temperate ice (Aschwanden et al., 2012; Hutter, 1982). In practice, we use the value $K_0 = 10^{-3} K_\mathrm{i}$, shown by Kleiner et al. (2015) to be sufficiently low to suppress transport of water by diffusion through the ice matrix, while still numerically stable.

2.3 Age equation

In addition to the balance equations above, we solve a separate equation for the ice age (χ) to determine the influence of the lake on the isochrone structure (Hindmarsh et al., 2009; Parrenin et al., 2006; Parrenin and Hindmarsh, 2007; Leysinger Vieli et al., 2007):

$$\frac{\partial \chi}{\partial t} + \boldsymbol{u} \cdot \nabla \chi = 1 + \left(d_\chi \nabla^2 \chi \right), \tag{9}$$

where χ is the age of ice and the second term on the right represents a diffusivity term needed for numerical stability, in which d_χ is the numerical diffusivity.

2.4 Boundary conditions

At the surface, stresses arising from atmospheric pressure and wind can be neglected as they are very small compared to the typical stresses in the ice sheet (Greve and Blatter, 2009), resulting in a traction-free boundary condition (BC),

$$\boldsymbol{\sigma} \cdot \boldsymbol{n} = 0, \tag{10}$$

where $\boldsymbol{n}$ is the normal vector pointing away from the ice. Accumulation and ablation (a_s) at the surface are assumed to be zero, giving the kinematic surface BC as:

$$\frac{\partial z_\mathsf{s}}{\partial t} + u\frac{\partial z_\mathsf{s}}{\partial x} + v\frac{\partial z_\mathsf{s}}{\partial y} - w = a_\mathsf{s} = 0, \tag{11}$$

where z_s is the surface elevation. We employ an inverse Weertman-type sliding law (Eq. 12), where the basal drag ($\boldsymbol{\tau}_\mathsf{b}$) is expressed as a function of the velocity of the ice ($\boldsymbol{u}_\mathsf{b}$) immediately above the ice/base interface, except over the lake surface where basal traction is set to zero (full slip). With basal sliding exponents $(p,q) = (1,0)$ appropriate for ice streaming conditions, the sliding relationship simplifies to a linear relationship between basal sliding and basal traction. The ice is assumed to be in hydrostatic equilibrium everywhere and the basal normal pressure (τ_n) taken as the ice overburden pressure. Ice accretion and melt at the base are assumed to be zero and along with the stress BC at the surface, a no-penetration condition is used to close the system:

$$\boldsymbol{u}_\mathsf{b} = -C^{-1}\frac{|\tau_\mathsf{b}|^{p-1}\boldsymbol{\tau}_\mathsf{b}}{N_b^q}\frac{|\tau_\mathsf{b}|^{p-1}\boldsymbol{\tau}_\mathsf{b}}{\tau_n^q} = -C^{-1}\boldsymbol{\tau}_\mathsf{b} \qquad \Rightarrow \qquad \boldsymbol{\tau}_\mathsf{b} = C\boldsymbol{u}_\mathsf{b}, \tag{12}$$

and

$$\boldsymbol{u}\cdot\boldsymbol{n} = 0, \tag{13}$$

where C is the sliding coefficient (Table 1). Periodic BC for the inlet/outlet are used, such that the velocity, pressure and specific enthalpy are the same on both ends. On the side boundaries of the domain, symmetry for velocity and thermal insulation are imposed:

$$\boldsymbol{u}_\mathsf{in} = \boldsymbol{u}_\mathsf{out} \qquad p_\mathsf{in} = p_\mathsf{out} \qquad H_\mathsf{in} = H_\mathsf{out}, \tag{14}$$

$$\boldsymbol{K} = [\mu(\nabla\boldsymbol{u} + (\nabla\boldsymbol{u})^T)]\boldsymbol{n}, \qquad \boldsymbol{K} - (\boldsymbol{K}\cdot\boldsymbol{n})\boldsymbol{n} = 0, \qquad \boldsymbol{u}\cdot\boldsymbol{n} = 0, \tag{15}$$

$$-\boldsymbol{n}\cdot\nabla H = 0. \tag{16}$$

At the surface, a value is set for specific enthalpy (H_s) corresponding to a surface temperature of $T_s = -30\,^\circ\mathrm{C}$ ($W_s = 0$, $p_s = 0$) such that:

$$H_s = H_i + c(T_s - T_m). \tag{17}$$

At the base, the geothermal flux (q_{geo}) is used for cold ice and a zero flux for temperate ice:

$$\boldsymbol{n} \cdot \left(\left\{ \begin{array}{c} -K_i \nabla H \\ -K_0 \nabla H \end{array} \right\} \right) = \left\{ \begin{array}{ll} q_{geo}, & \text{if } T < T_m \\ 0, & \text{if } T = T_m \end{array} \right\}, \tag{18}$$

where H_i is the specific enthalpy of pure ice at the melting temperature. To correctly determine the basal boundary condition for the enthalpy field equation, a switching between a Dirichlet and a Neumann condition is necessary (Aschwanden et al., 2012; Kleiner et al., 2015), which depends on basal temperature, water availability at the base and whether a temperate layer exists or not immediately above it. Here we opt for a simpler, more computationally efficient boundary condition where the geothermal flux is gradually decreased in the specific enthalpy range corresponding to [($T_m - 0.2\,^\circ\mathrm{C}$) ($T_m$)], with a smoothed Heaviside function with continuous derivatives.

For the age–depth relation (Eq. 9), periodic boundary conditions are used for the inlet and outlet. At the surface, the age is set to $\chi_s = 0$. Zero normal fluxes are used for the side boundaries and the lower boundary.

$$-\boldsymbol{n} \cdot (-\nabla \chi) = 0 \tag{19}$$

2.5 Computational domain

We define the computational domain as a rectangular box 350 km long and 100 km wide with a fixed bed slope (α). ~~The ice thickness is set to 1500and three~~ Three different lakes sizes are used, based on typical sizes of subglacial lakes found in Antarctica (Siegert et al., 1996; Siegert and Kwok, 2000). The lakes are all elliptical in shape, with major and minor axes defined in Table 2. The major axis is aligned with the direction of flow. The model experiments are all either steady state solutions or start from steady state solutions.

Throughout the domain, extruded triangular (prismatic) elements are used with a horizontal resolution down to $\sim 500\,\mathrm{m}$ at the lake edges. This relatively high resolution is needed in order to capture the effect of strain softening of ice around the lake edges where velocity gradients are large and to properly resolve the upper surface. The model uses 15 vertical layers, which become thinner towards the base, where the thinnest layer has a thickness of $\sim 35\,\mathrm{m}$.

3 Model experiments

The aim of the study is to show how ice dynamics and the thermal evolution are affected by the presence of a subglacial lake and to follow its effect on the internal layering through simple temporally dependent and steady state experiments.

All transient simulations are started from an initially steady-state configuration, where equations for mass, momentum, and enthalpy are solved jointly with a direct solver along with equations for surface or grid evolution. All presented results are based on simulations with a fully nonlinear ice viscosity ($n = 3$), unless explicitly stated otherwise and all numerical values for model constants are defined in Table 1.

The lake itself is modelled as a “slippery spot” (Pattyn et al., 2004; Sergienko et al., 2007). The lake surface is assumed to be fixed to the bed plane. In reality, any changes to lake size or volume would lead to vertical movement of the lake surface, which in itself would induce an expression at the surface of the ice. ~~For such a scenario, the strongest effect would be experienced by the internal layers closest to the bed~~ Sergienko et al. (2007) found that vertical movement of lake surfaces due to drainage resulted only in small changes to ice velocity and that the ice surface reached its initial stage after just a few years, indicating that it would not lead to a significant disturbance in isochrone structure compared to changes in the horizontal lake extent. Here we consider only planar changes in lake geometry, or changes in lake size fixed to the bed plane.

Water draining from subglacial lakes is likely to have an impact on ice dynamics and isochrone structures downstream of the lake, either through melting of ice or changes in basal water pressure and sliding, although no representation of subglacial hydrology is included here. The effect the draining water will have on isochrone layers will depend on the state of the drainage system and how the water is transported downstream (Stearns et al., 2008) but will likely be limited by the swiftness of the ice response compared to the time needed to considerably perturb isochrone layers.

The lack of basal friction over the lake results in increased velocities, not just over the lake but also a considerable distance both upstream and downstream of the lake ~~location~~that depends on both ice thickness, basal traction (Hindmarsh, 2006; Kamb and Echelmeyer, 1986) and the size of the lake. Figure 1a shows the surface velocity with black lines indicating streamlines and Fig. 1b shows the relative change in surface elevation caused by the presence of the lake, where the general slope has been subtracted from the surface. The surface responds to the change in basal conditions by becoming flatter and with more than a doubling of horizontal velocities over the lake. Streamlines contract and ice is brought in towards the lake from the sides.

Figure 2 shows different model output in vertical cross sections, through the center of the lake, in the direction of flow for the L_{M} lake size~~and different lake parameters~~. Over the lake, the ice basically moves as an ice shelf with more or less uniform horizontal velocity throughout the ice column (Fig. 2a). At the upstream end of the lake, a strong downward movement of ice causes cold ice from the upper layers to be drawn down towards the bottom, steepening the temperature gradient close to the base. Conversely, at the downstream end a strong upward flow restores internal layers to their prior depths, before the influence of the lake is felt (Fig. 2b).

The temperature and microscopic water content within the ice are shown in Fig. 2b. Black lines represent stream-lines which coincide, or line up, with isochrone layers for steady-state simulations (Hindmarsh et al., 2006).

The intense internal deformation close to the borders of the lake, where ice velocities change significantly over short distances, gives rise to a thin temperate layer of ice at both the upstream and downstream ends of the lake (Fig. 2b). The vertical resolution is somewhat limited in the model but simulations with higher resolution do indicate that a temperate layer should form for the given velocity field, although its thickness might be slightly overestimated. This is strongly affected by the chosen sliding law and sliding coefficients and should not be considered representative for subglacial lakes in general.

Figure 2c shows the internal deformation energy, with high values near the base where velocity gradients are high. Maximum values are reached near the edges of the lake, in the transition zone between low and full sliding.

Increases in temperature, pressure, water content and larger effective stress all have the effect of decreasing the viscosity, which is why the lowest viscosity values are obtained at the base of the ice sheet, in particular around the edges of the lake (Fig. 2d). Generally low viscosity is furthermore obtained throughout the ice column over the lake boundary, extending all the way up to the surface, effectively creating a vertical zone of softer ice.

Figure 3a shows profiles of horizontal surface velocity for the three different lake sizes considered. As expected, velocity peaks over the lake surfaces and two fringe peaks at the lake edges are discernible for the smallest lake size as well.

A characteristic surface dip feature is observed at the upstream end of many lakes in Antarctica, for instance the Recovery Lakes (Bell et al., 2007; Langley et al., 2011) and Lake Vostok (Studinger et al., 2003) as well as a small hump at the downstream end. Figure 3b shows surface profiles for the three different lake sizes, each with the characteristic flattening of the surface as well as the dip and ridge features on each side of the lake.

Figure 4a shows a comparison of surface profiles for the L_{M} lake size, made with a constant ($\eta_{\mathrm{const}} = 10^{14}$ Pa s) vs. a nonlinear viscosity ($n = 3$ in Glen's flow law) and a pressure and temperature dependent one ($n = 1$) but otherwise using the same setup. The hummocky feature is noticeably absent from simulations with a constant viscosity. Figure 4b and c shows vertical profiles, over the center of the lake, of horizontal velocity and viscosity for the three different viscosity cases. The horizontal velocity has been scaled with the horizontal velocity at the surface (the uppermost point of the vertical profile). For both the $n = 1$ and $n = 3$ cases, the horizontal velocity at depth is larger than at the surface.

A weaker contrast in basal traction inside and outside the lake, or a less pronounced switch in flow mode (Leysinger Vieli et al., 2007; Parrenin and Hindmarsh, 2007), decreases the amplitude of the dip and ridge features as can be witnessed in Fig. 5a, where surface profiles for simulations with different sliding coefficients and ice thickness are compared ($n = 3$). Isochrones relative to the ice base are shown in Fig. 5b for two of the simulations presented in Fig. 4, ($n = 3$ (blue line) and $\eta = 10^{14}$ (red line)), in addition to isochrones for a simulation with a lower contrast in basal traction and a lower bed inclination than the other two (cyan line). The structure of internal layers depends heavily on both the contrast in basal traction and the viscosity formulation used. Low contrast results in small deflections of internal layers whereas a fixed viscosity (compared to flow with nonlinear viscosity and a similar surface velocity) results in larger deflections of internal layers due to the generally smaller ice flux at depth outside the lake compared to over it.

Lakes drain and fill on different time scales. Several studies have documented relatively rapid drainage events in subglacial lakes in Antarctica (Gray et al., 2005; Wingham et al., 2006; Bell et al., 2007). Typically, drainage occurs over the course of several years but take much longer to refill. To simulate such an event and what effect it could have on the internal structure of the ice, we set up a model run where the lakes diameter shrinks during a 10 year period from the maximum (L_L) to the smallest lake size used in the paper (L_S). Figure 6 shows four different time slices of horizontal velocity, with black lines indicating isochrone layers. As the velocity field adjusts in time to the new boundary conditions, a travelling wave is created at depth within the isochrone structure that transfers downstream with the flow of ice.

4 Discussion

The frictionless boundary levels the surface over the lake, changing surface gradients and causing the ice to speed up in the vicinity of the lake. The increase in velocity is further amplified by the effect of velocity gradients on ice viscosity. Outside the lake, the velocity field is primarily affected by changes in surface gradients and longitudinal stresses and, to a lesser degree, by changes in ice viscosity and basal velocities. Over the lake, the ice moves more like an ice shelf, with almost uniform horizontal velocity throughout the ice column. The increase in velocity (Fig. 2a) is due to the lack of basal friction over the lake, causing the highest velocities to appear there but secondary velocity peaks (Fig. 3) can also be discerned at the lake edges, which result from the interaction of surface evolution and ice dynamics. The velocity increase for the two fringe peaks propagates from the surface and downward whereas the velocity peak over the lake is mostly caused by acceleration in the basal layers of the ice sheet.

4.1 Thermal regime

For the given model setup, the geothermal flux is sufficient to ensure that the basal temperature reaches the pressure melting point everywhere in the model. Internal deformation then ~~raises the specific enthalpy~~ adds to the available thermal energy with the potential to form a temperate layer with non-zero microscopic water content in the vicinity of the lake ~~, effectively creating a thin layer~~ as seen in Fig. 2b, where a thin temperate layer has formed ($\sim 70\,\mathrm{m}$, 2 vertical cells)~~of temperate ice with non-zero microscopic water content~~. Lakes in ice streaming areas, where a considerable portion of the surface velocity arises due to sliding at the base can be expected to have a much gentler transition in flow mode at the lake edges and much less deformational energy available. No temperate layer forms for simulations where the transition in sliding is considerably less sharp and more appropriate for streaming areas as in the cases with $C = 10^{10}$ (Fig. ~~2b).~~ 5).

The steeper temperature gradient over the lake efficiently leads away excess heat created by internal deformation at the upstream end, refreezing whatever microscopic water created upstream. Large quantities of microscopic water, in reality, would drain to the base ($\gtrsim 3\,\%$), although drainage is not included here as water content is relatively low.

Two factors would contribute to facilitating freeze-on at the interface between ice and lake water, if it were included in the model. First, draw-down of cold ice from ~~the surface~~ above increases the temperature contrast in the lower part of the ice sheet, second, heat from internal deformation ceases with the removal of basal traction over the lake surface. Both result in a steepening of the temperature gradient close to the base, which more efficiently removes heat from the base and shifts the balance in favour of freeze-on.

As lake size increases so too do horizontal velocities over the lake (Fig. 3), as well as the effective strain rates and ~~as a result, the thickness of the temperate layer at the lake boundaries also increases~~the available deformational energy for internal heating. Accretion rates at the ice/lake interface would be limited by the amount of latent thermal energy that the ice above it can efficiently lead away and any temperate layer formed by internal deformation at the lake boundary would, during its existence, block heat flow and thus accretion, until it completely refroze first. ~~In addition, the larger the lake, the further down cold ice from the upper layers of the ice sheet are advected, further increasing the temperature contrast at depth.~~

4.2 Transition zone

The softening effect of local increases in effective stress at the lake edges ~~,~~ effectively creates a vertical layer of soft ice in between ~~,~~ higher viscosity ice (Fig. 2c). Strong vertical flow at the edges of the lake results from the localized lack of basal traction. A clear difference can be seen between simulations with constant viscosity and viscosity that depends on pressure and temperature (Fig. 4). The softening effect of increasing temperature and pressure with depth not only causes velocity changes in the vertical to be concentrated in the lower layers of an ice sheet but also means that for areas with varying basal traction, such as subglacial lakes, that the ice at depth will support less lateral shear and longitudinal stresses compared to the upper layers where viscosity is higher and the ice stiffer. This in return means that as the ice encounters a slippery spot or a spot with a sharp decrease in basal traction, such as a subglacial lake, that the force balance will be different than in the isotropic case, where viscosity is ~~everywhere the same~~constant. The imbalance in mass flux at depth must be compensated by a more localized increase in vertical flow and a subsequent drop in surface elevation at the upstream side and an upwelling at the downstream side of the lake due to the limited vertical extent of a typical ice sheet.

For large subglacial lakes, where the basal traction is zero, the horizontal velocity at depth is predicted to be slightly larger (here about 0.1 % for $n = 3$, Fig. 4b) than at the surface. Lakes, such as subglacial lake Vostok, should therefore experience extrusion flow at the base, where the basal horizontal velocity exceeds that at the surface. Extrusion flow is not a requirement though for the formation of these dip and ridge features, as it is sufficient to have a sharp transition in sliding ~~velocity~~ along with a ~~weaker lower layer~~ negative downward gradient in ice viscosity to form them and they are predicted to form as well for situations where there is simply a strong decrease in basal traction, not necessarily a complete disappearance.

4.3 Drainage experiment

Drainage cycles in Antarctica, cycles including both draining and filling i.e. can have frequencies on decadal to centennial time scales or potentially even larger (Pattyn, 2008; Wingham et al., 2006). Although the drainage takes course during 10 years here, it can be seen as instantaneous given the generally slow flow of ice. The velocity field adjusts rather rapidly ($\sim 100\,\mathrm{a}$) to the new basal boundary conditions relative to the time it takes for the isochrones to respond as the flow of ice is relatively slow. For episodic drainage cycles, the strength of the response will partly depend on how long it takes for the lake to refill. After roughly 2000 years, the wave has moved far enough downstream to be more or less separated from its initial location (Fig. 6d). As both the upstream and downstream lake boundaries move during the drainage event, a two-troughed wave is created. In draining lakes where one end is much deeper than the other a single wave would be expected, as only the shallower end is likely to move significantly. The velocity of the moving boundary also ~~effects~~ affects the amplitude of the resulting isochrone disturbance (Wolovick et al., 2014), where a slip boundary, moving with the ice, is capable of distorting isochrone layers to a much greater extent than stationary slip boundaries. For maximum effect, the boundary should be moving at a velocity comparable to the averaged ice column velocity. Only slip boundaries moving with the ice are capable of distorting layers to a greater extent than stationary boundaries. Boundaries moving in opposite directions to ice flow, like our downstream lake boundary, will have a smaller impact on internal layers than a stationary boundary as the relative horizontal velocity increases.

Water in very active subglacial systems, such as recently discovered in West Antarctica (Gray et al., 2005; Fricker et al., 2007) has relatively short residence times and fast circulation, contrary to previous beliefs. The impact of such short drainage cycles, where both drainage and refilling happen on decadal time scales, is unlikely to have a strong effect on the internal isochrone layers as it takes a long time for the ice to respond. Regular drainage events in subglacial lakes that have a much shorter cycle than the time it takes for a particle of ice to be brought up by vertical flow at the edge of the lake, will therefore probably not be easily detectable in the isochrone structure. If the frequency of drainage cycles is high, the ice will have little time to respond and the amplitude of the resulting wave will be small. Drainage of large lakes in areas with low basal melt rates and consequently long filling times on the other hand could be expected to generate travelling waves, downstream of the lake, with a sufficiently large amplitude to be detectable within downstream isochrones. The amplitude of the travelling wave would be expected to be similar in magnitude as the steady state isochrone disturbance over the lake itself (Fig. 6). Here, on the order of $\sim 100\,\mathrm{m}$ at depth, well within the bounds of being measurable by modern radar systems. Drainage of subglacial lakes where the transition in flow mode is less abrupt, would result in smaller amplitudes (Fig. 5b). As layer stratigraphy is often quite complex, a numerical model of ice age and velocities would be needed to separate the effect of temporally changing lake size, or basal conditions, from layer deflections caused by varying basal topography, or rheology.

For our particular setup, the isochrone disturbance, or the travelling wave, should eventually overturn and create a fold as the stress situation downstream of the lake is essentially one of simple shear without any longitudinal extension (Waddington et al., 2001; Jacobson and Waddington, 2005). Resolving this in the model would require an equally fine resolution downstream of the lake, as over the lake itself, ~~so this~~ which is not done here. For a subglacial lake situated at the onset of streaming flow, a fold might not be expected though, as overturning would be counteracted by longitudinal extension and vertical compression, which would tend to flatten all layer disturbances. In general, both horizontal shear and longitudinal extension can be assumed to be present and thus, whether a layer disturbance develops into a fold or flattens out, to eventually disappear, will be decided by the balance between the two (Waddington et al., 2001).

5 Conclusions

Subglacial lakes represent areas of the ice sheet base which are incapable of exerting any horizontal stress on the overlying ice. ~~The rapid transition from~~ A rapid transition between little to full sliding at the lake edges causes intense deformation and internal heating, leading to both an increase in enthalpy and strain softening of ice~~at the lake edges, and can result in the formation of a temperate layer of ice~~. The decrease in viscosity with depth leads to preferential deformation of the lower layers of an ice sheet which ~~, for subglacial lakes~~along with a sharp transition in sliding, results in a hummocky surface feature above the upstream and downstream ~~lake edges~~ edges of subglacial lakes. These dip and ridge features can be taken as evidence for a rapid transition in basal sliding, with strong vertical movement of ice as a result. Lakes without this feature would be expected to experience a more gradual increase in sliding, approaching the lake, a lower contrast in basal traction or continued fast flow downstream, as with subglacial lakes at the onset of ~~ice streaming.~~ or in ice streaming areas. Over the lake itself, the ice moves more like an ice shelf, with zero basal traction and virtually no internal strain heating. As a result, temperature gradients over the lake steepen, making them more prone to freeze-on at the ice/water interface~~than other areas of the ice sheet~~. A rapid decrease in lake size (or basal friction) causes a travelling wave to be created at depth within isochrone layers, suggesting that certain aspects of lake history are preserved within them and could potentially be deciphered from radio echo sounding data downstream of lake locations when combined with output from numerical models.

Acknowledgements. Funding for this work came from the Research Council of Norway (RCN), Statoil, Det Norske ASA and BG group Norway (grant 200672) to the PetroMaks project "Glaciations in the Barents Sea area (GlaciBar)", and from the Research School in Arctic Marine Geology and Geophysics (AMGG) at the University of Tromsø. This is also a contribution to the Centre of Excellence: Arctic Gas Hydrate, Environment and Climate (CAGE) funded by RCN (grant 223259). In addition, we would like to thank Nina Wilkens and Martin Rückamp for help with getting started with COMSOL, ~~and~~ the Stallo support team for invaluable assistance and Gwendolyn Leysinger Vieli and one anonymous reviewer for constructive reviews that greatly improved the quality of the manuscript.

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Table 1. Values for constants used in the study.

	Constants	Values
α	bed inclination	$0.3^\circ, 0.1^\circ$
ρ	density of ice	$910\,\mathrm{kg\,m^{-3}}$
g	gravitational acceleration	$9.81\,\mathrm{m\,s^{-2}}$
n	flow law exponent	$\cancel{3}$ $3, 1$
γ	Clausius–Clapeyron constant	$9.8 \cdot 10^{-8}\,\mathrm{K\,Pa^{-1}}$
${T_\mathrm{m}}^0$	melting point at atm.pressure	$273.15\,\mathrm{K}$
A_0	pre-exp. constant ($T \leq 263.15\,\mathrm{K}$)	$3.985 \cdot 10^{-13}\,\mathrm{s^{-1}Pa^{-3}}$
	– ($T > 263.15\,\mathrm{K}$)	$1.916 \cdot 10^{-3}\,\mathrm{s^{-1}Pa^{-3}}$
Q	activation energy ($T \leq 263.15\,\mathrm{K}$)	$60\,\mathrm{kJ\,mol^{-1}}$
	– ($T > 263.15\,\mathrm{K}$)	$139\,\mathrm{kJ\,mol^{-1}}$
R	universal gas constant	$8.3145\,\mathrm{J\,(mol\,K)^{-1}}$
k	thermal conductivity	$2.1\,\mathrm{W\,(m\,K)^{-1}}$
c	heat capacity	$2009\,\mathrm{J\,(kg\,K)^{-1}}$
L	latent heat of fusion	$3.35 \cdot 10^5\,\mathrm{J\,kg^{-1}}$
d_χ	diffusion coefficient	$10^{-13}\,\mathrm{m^2\,s^{-1}}$
C	friction coefficient	$\cancel{10^{13}}$ $10^{13}, 10^{10}\,\mathrm{(s\,m^2)\,kg^{-1}}$
$\cancel{H}$ h	ice thickness	$1500\,\mathrm{m}, 3000\,\mathrm{m}$
a_s	surface accumulation	$0\,\mathrm{m\,s^{-1}}$
T_s	surface temperature	$-30\,^\circ\mathrm{C}$
q_geo	geothermal flux	$55\,\mathrm{mW\,m^{-2}}$
η_const	ice viscosity (constant)	$10^{14}\,\mathrm{Pa\,s}$

Table 2. Values for lake sizes used in the paper.

Lake size	Major axis	Minor axis
L_{S}	10 km	5 km
L_{M}	20 km	10 km
L_{L}	30 km	15 km

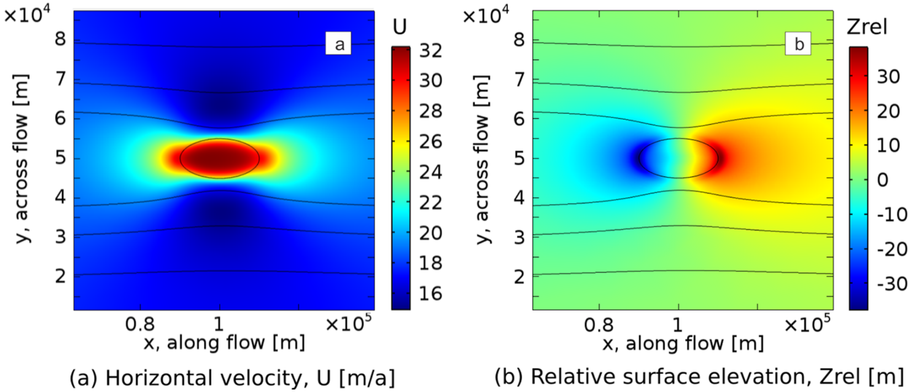

Figure 1. 2-D surface plots of horizontal velocity **(a)** in [$\mathrm{m\,a^{-1}}$] and relative surface elevation **(b)** in [m]. The outline of the lake and streamlines are shown in black and flow is from left to right.

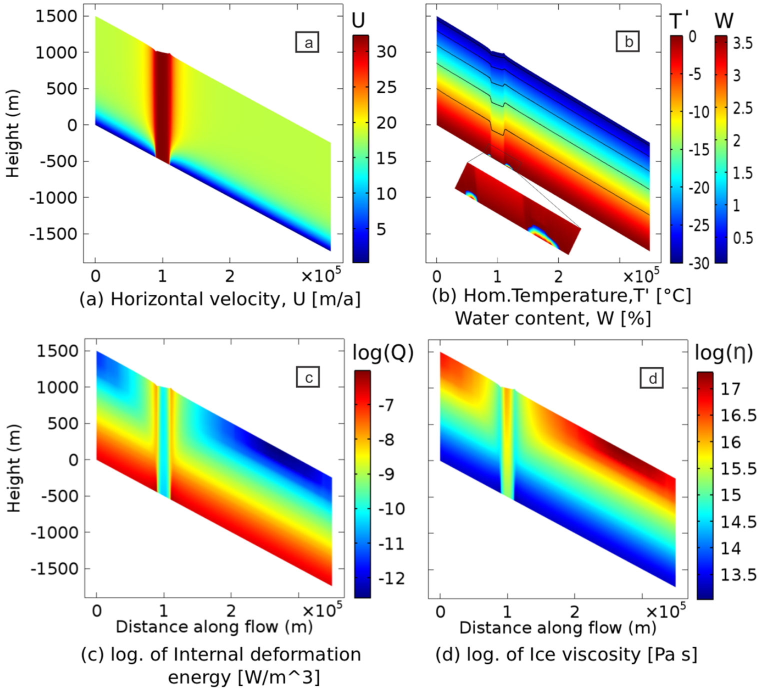

Figure 2. ~~Cross-sectional plots in the~~ Along flow ~~direction~~profiles, through the center of the lake (L_M lake size, $n = 3$). **(a)** shows horizontal velocity in [m a^{-1}]. **(b)** shows the homologous temperature in °C and water content in percentage. The black lines represent streamlines. The inlet figure shows a close-up of the temperate layers at the lake edges (same color scales). In **(c)** we see the logarithm of the internal deformation energy [W m^{-3}] while **(d)** shows the logarithm of the viscosity [Pa s].

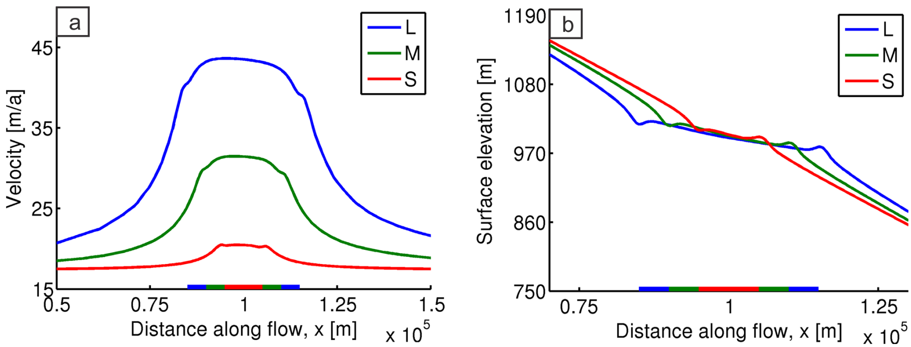

Figure 3. Surface velocity **(a)** and elevation **(b)** ~~across the lake~~along flow ($n=3$), ~~in~~ through the ~~flow direction~~center of the lake for various lake sizes. Velocity and surface profiles for the different lake sizes (L, M, S) are marked with different colors and the horizontal extent of each lake is marked with a ~~vertical~~ horizontal bar at the bottom. Note the different horizontal scale for the two figures.

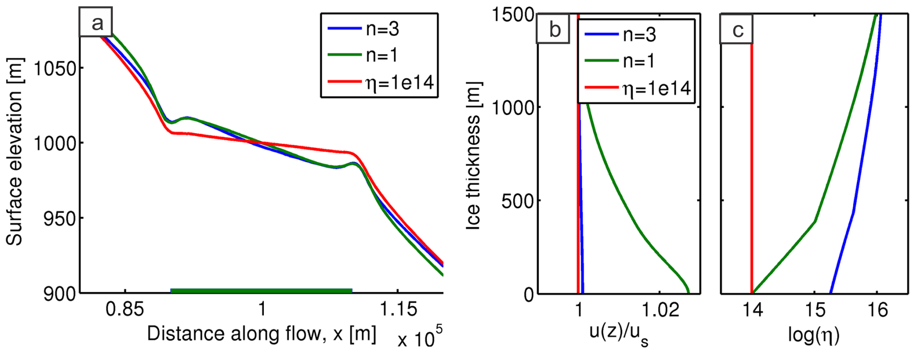

Figure 4. (a) Profiles of surface elevation in [m] for the L_{M} lake size and three different viscosity cases; fixed viscosity ($\eta_{\mathrm{const}} = 10^{14}$ Pa s, red), with flow exponent $n = 1$ (green) and $n = 3$ (blue) in Glen's flow law (Eq. 4). Vertical profiles of **(b)** ~~Scaled~~ the scaled horizontal velocity and **(c)** the logarithm of ice viscosity, ~~in a vertical profile~~ over the center of the lake. The horizontal velocity has been scaled with the velocity magnitude at the surface.

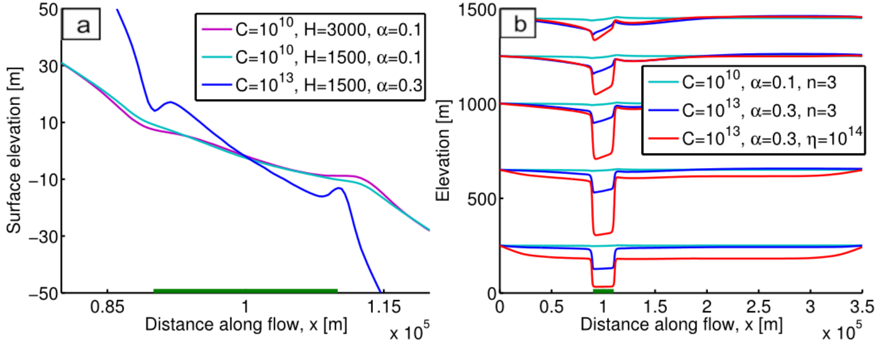

Figure 5. ~~Four~~ **(a)** Profiles of relative surface elevation in [m] for the L_M lake size for different ~~time slices~~ values of the ~~velocity~~ sliding coefficient and ~~isochrone adjustment to a drainage event~~ice thickness. The ~~color~~ blue line is the same surface profile as the blue line in Fig. 4a only displaced vertically for comparison. Note the different vertical scale ~~represents horizontal velocity and yellow and black lines indicate streamlines and isochrones respectively~~compared to the previous figure. ~~**(a)** represents~~ **(b)** Isochrone layers for the ~~initial stage~~ $n=3$ viscosity case with different values of the sliding coefficient (~~$t=0$~~cyan and blue lines) ~~, **(b)** a 100later~~ and for the fixed viscosity case (~~$t=100$a~~red line)~~, **(c)** $t=500$a~~. The lake size is L_M and ~~**(d)** $t=2000$a~~the isochrones are presented relative to the base.

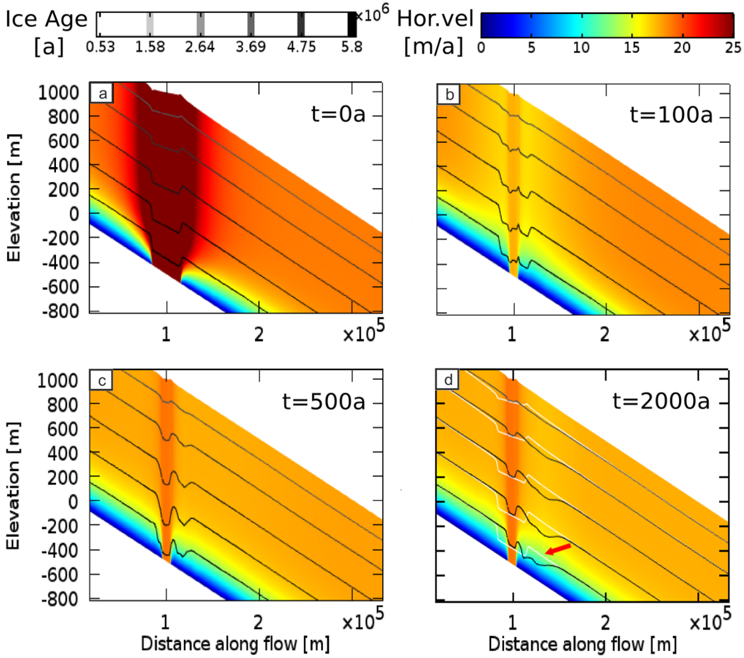

Figure 6. Four different time slices of the velocity and isochrone adjustment to a drainage event. The color scale represents horizontal velocity [m a^{-1}] and gray to black lines indicate isochrones. **(a)** represents the initial stage ($t = 0$), **(b)** a 100 years later ($t = 100$ a), **(c)** $t = 500$ a and **(d)** $t = 2000$ a. The white lines in the last time slice are the initial isochrone layers from $t = 0$ and the red arrow points to the isochrone disturbance that advects downstream with the flow of ice.