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Traveling slippery patches produce thickness-scale folds in ice sheets

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Key Points:
- Slippery patches beneath ice sheets propagate downstream
- Propagating slippery patches create massive stratigraphic structures
- The internal structure of ice sheets is related to the history of basal slip

Supporting Information:
- Readme
- Text S1 and Figures S1-S6
- Movie S1

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Abstract Large, complex stratigraphic folds that rise as high as 60% of the local ice thickness have been observed in ice sheets on Antarctica and Greenland. Here we show that ice deformation caused by heterogeneous and time-variable basal sliding can produce the observed structures. We do this using a thermomechanical ice sheet model in which sliding occurs when the base approaches the melting point and slippery patches develop. These slippery patches emerge and travel downstream because of a feedback between ice deformation, vertical flow, and temperature. Our model produces the largest overturned structures, comparable to observations, when the patches move at about the ice column velocity. We conclude that the history of basal slip conditions is recorded in the ice sheet strata. These basal conditions appear to be dynamic and heterogeneous even in the slow-flowing interior regions of large ice sheets.

1. Introduction

Ice-penetrating radar transects collected over several decades show nearly horizontal reflections in the upper part of the ice sheet indicating layered strata of meteoric ice formed from snowfall [e.g., *Bailey et al.*, 1964]. In recent years, the collection of closely spaced transects of high-quality aerogeophysical data [*Li et al.*, 2013] has allowed the identification of widespread irregular reflectors in the lower part of the ice sheet [*Bell et al.*, 2011, 2014; *NEEM Community Members*, 2013]. These reflectors form structures that do not conform to the bed or surface and that disturb the overlying stratigraphy into anticlines, synclines, and overturned folds (Figure 1). These folds normally have older, reflective ice surrounding a core of featureless ice. In some cases, reflectors emerge from near the bed. The folds can be quite large, with thicknesses up to ~1000 m in ice that is about 2000 m thick, widths of ~10–20 km, and lengths on the order of 10–100 km [*Bell et al.*, 2014].

Several hypotheses exist for the formation of these folds. In Antarctica, these features are closely associated with topographically confined water networks [*Wolovick et al.*, 2013] and resemble the refrozen ice sampled and imaged over Lake Vostok [*Jouzel et al.*, 1999; *MacGregor et al.*, 2009]. In the interior of Greenland, large-scale folds do not have a clear relationship to basal water networks, although the folds may originate near a warm to cold bedded transition [*Aschwanden et al.*, 2012]. Alternately, rheological contrasts within the ice column may create these folded features through a shear instability in the flowing ice [*NEEM Community Members*, 2013]. Here we consider the possibility that this folded stratigraphy results from changing basal slip. We show how warm slippery patches at the ice sheet base travel downstream and that the stratigraphic structures created by traveling slippery patches can explain many of the radar observations.

Stratigraphy deflects vertically where ice flow crosses areas with different basal slip rates [*Weertman*, 1976]. Such gradients in slip rate cause either divergence or convergence of the horizontal flow field near the bed, depending on whether the bed becomes more or less slippery along flow, respectively. To maintain mass balance in the lower part of the ice sheet, there is a vertical flow component that lifts up or draws down the stratigraphy. In some cases, the vertical flow causes vertical motion of the surface [*Sergienko et al.*, 2007]. In other cases, deformation with one sign in the lower part of the ice sheet is compensated for by deformation with the opposite sign in the upper portion of the ice sheet, and the surface maintains a steady state [*Hindmarsh et al.*, 2006]. The latter case has been observed where ice flowing over Lake Vostok experiences thickening near the bed and thinning near the surface when it regrounds [*Bell et al.*, 2002]. However, stationary contrasts in slip rate produce small layer deflections compared with bed topography or surface accumulation, and basal slip rate has been considered a secondary factor in the analysis of ice sheet stratigraphy [*Hindmarsh et al.*, 2006; *Leysinger Vieli et al.*, 2007; *Parrenin and Hindmarsh*, 2007]. Here we consider slippery patches that travel downstream over time and produce a larger integrated deflection of the overlying stratigraphy.

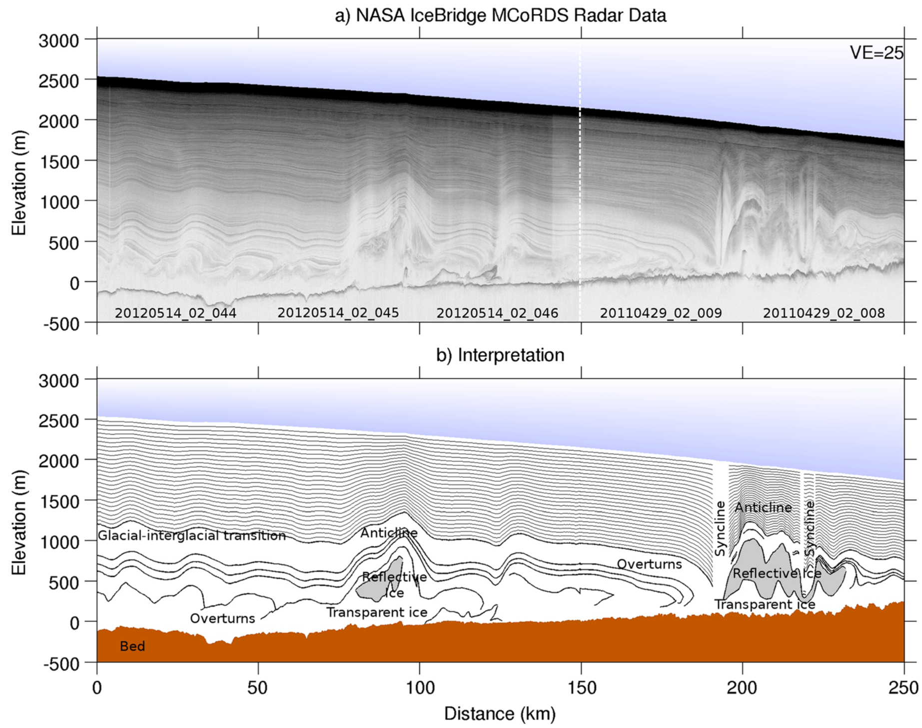

Figure 1. Radar image from northern Greenland depicting basal units and associated deformation. Image comes from two separate flight lines stitched together near 150 km. Data are located at ftp://data.cresis.ku.edu/data/rds/, and the identification numbers in Figure 1 (top) can be used to locate the data in question. (top) The radar data and (bottom) our interpretation of the data. Vertical exaggeration is 25. Flight line is slightly oblique to flow.

Basal slip rate can change in space or time because of a variety of factors, including basal temperature [*Cuffey et al.*, 2000; *Clarke*, 2005; *Stokes et al.*, 2007], till coverage [*Tulaczyk et al.*, 2000; *Clarke*, 2005; *Stokes et al.*, 2007], subglacial water pressure [*Creyts and Schoof*, 2009; *Hewitt*, 2013], and till drainage [*Tulaczyk et al.*, 2000; *Clarke*, 2005; *Stokes et al.*, 2007]. While all these factors are capable of moving over time, for simplicity, we focus on slip variations caused by basal temperature. We use a thermomechanical model described below to show that basal slippery patches are capable of migrating downstream and that the resulting dynamic contrasts in slip rate create large and intricate stratigraphic structures.

2. Methods

We simulate ice flow using a higher order two-dimensional thermomechanical flow line approach [*Blatter*, 1995; *Pattyn*, 2002; *Cuffey and Paterson*, 2010]. Particle tracking allows us to follow englacial stratigraphy and model isochronous layers. A key feature of our model is that it allows for both water flow and longitudinally variable basal boundary conditions while conserving mass and energy at the ice sheet base. Basal temperature is coupled to water flow along the base as follows: temperature is held at the melting point where water is present while a geothermal gradient is prescribed elsewhere. The model can contain multiple warm-based patches, each with an internal mass balance between melting and freezing. Basal slip rate falls off exponentially as basal temperature drops below the melting point [*Fowler*, 1986], consistent with field observations that suggest limited but nonzero basal slip at cold temperatures [*Cuffey et al.*, 2000]. Subfreezing basal slip is due to a combination of premelt films at the ice-rock contact [*Cuffey et al.*, 2000] and unresolved thermal heterogeneity at the bed [*Fowler*, 1986]; for our purposes, it is not necessary to distinguish these mechanisms. The boundary conditions and model setup are shown schematically in Figure S1, and a more detailed description is given in the supporting information. The steady state initial condition is obtained by allowing the model to run unperturbed for 100 ka. We performed resolution tests on the model and found that the error is inversely proportional to the number of grid cells (Figure S3 in the supporting information).

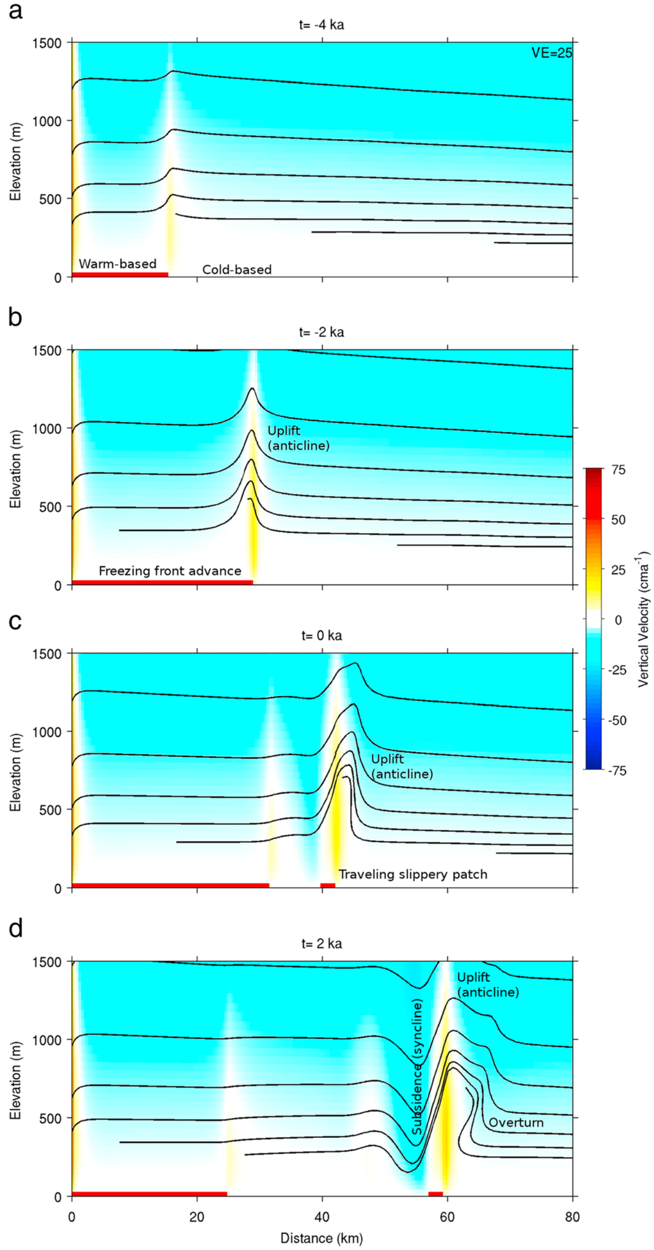

Figure 2. (a–d) Close-up of the initial triggering of a traveling slippery patch. Plots represent four snapshots of model output taken 2 ka apart. Time is measured relative to the peak in the water influx perturbation. Color represents vertical velocity, and lines represent stratigraphy. Red bars on the bottom of the plots represent areas where the bed is at the melting point.

Our model domain simulates a thick interior region with a warm base that transitions downstream into a thinner, cold-based region (Figure S1). This setting was chosen to mimic conditions that may exist upstream of Petermann Glacier and in the flanks of the Northeast Greenland Ice Stream [*Aschwanden et al*., 2012], where many large folds are observed [*Bell et al*., 2014]. Water flows along the smooth base from the warm interior region toward the cold downstream area.

3. Results

When we perturb the upstream water flux, a warm slippery patch forms and travels downstream. At the leading edge of the slippery patch, a freezing front marks the boundary between an upstream, warm-based, water-rich, rapidly sliding patch and a downstream, frozen, slowly sliding region. We find that the perturbation must be sufficient to cause the freezing front to move downstream at approximately the column average ice velocity in order to trigger a traveling slippery patch. The perturbation we used is equivalent to an increase in melt rate of only 0.05 mm/yr over the course of about 5000 years, assuming a cross-flow width of 10 km and an upstream catchment of 10^4 km^2. This increase is small compared to geothermal melt rates that are about 1 mm/yr.

The position of the freezing front is determined by a balance between the water flux and the freezing rate. As the upstream water flux increases, the freezing front advances downstream as latent heat is released. Near the freezing front, basal slip rate drops and produces ice convergence that results in englacial uplift (Figures 2a and 2b). This uplift moves warm ice higher into the ice sheet, reducing the local temperature gradient. Eventually, the conductive heat flux out of the basal interface does not match the geothermal heating and frictional heating from sliding, creating a local melt source just upstream of the freezing front. Once the uplift at the freezing front allows local melt, the slippery patch is no longer limited by the mass balance of the upstream water system. The patch detaches from the upstream water system and travels downstream without further forcing (Figures 2c and 2d).

The main part of a slippery patch is composed of a broad melting region that supplies water to a narrow but intense freezing region near the front (Figure S2). Water is assumed to flow downstream throughout the

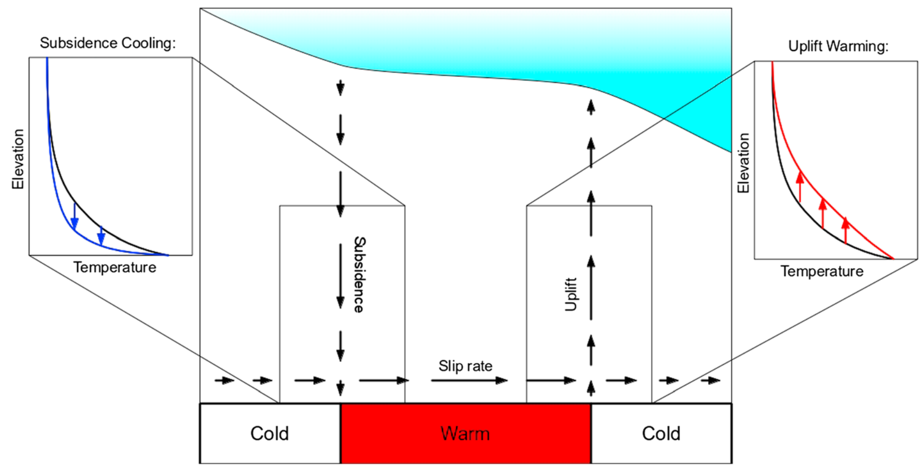

Figure 3. Schematic representation of the thermal feedback that causes slippery patches to move downstream. Englacial uplift at the front (downstream) edge of the patch produces advective warming of the ice column, while subsidence at the back (upstream) edge of the patch produces advective cooling. Inset temperature profiles are meant to be representative of the changes that take place at the margins of the slippery patch. Black lines in the inset represent background temperature profiles, and red and blue lines represent changes due to uplift or subsidence, respectively.

patch due to the gradient in ice overburden pressure. Lower temperatures behind (upstream of) the slippery patch reduce basal slip rate, leading to divergence in the bottom of the ice sheet. This divergence pulls cold ice downward, increasing the conductive heat loss across the basal interface and cooling it below the melting point. Because water is driven downstream, freeze-on does not occur at the back of the patch. The combination of cooling at the back (upstream) and warming at the front (downstream) causes the entire slippery patch to propagate downstream (Figure 3). Previous work has shown that perturbations to basal slip modify ice sheet thermal structure [*Sergienko and Hulbe*, 2011]. Our model shows that closing the feedback loop from thermal structure back to slip can cause the slip perturbation to move. The front of a slippery patch propagates faster than the back because latent heat carried forward by basal water is transported faster than the ice flows. The difference in propagation speed causes the patch to grow over time.

A secondary patch can be generated as a consequence of the subsidence and cooling behind an initial slippery patch. Subsidence immediately behind the slippery patch causes cooling relative to the temperature far upstream. This subsidence produces a basal slip rate contrast upstream of the initial slippery patch, leading to convergence and uplift there. Eventually, the upstream convergence warms to the melting point, creating a new, trailing slippery patch (Figure 4d). Over time, this sequence repeats, creating a train of patches and deformation.

Patches in a train can combine when the fast-moving front of one patch overtakes the slow-moving back of the next one. Basal temperatures in the intervening cold region rise to the melting point. As temperatures rise, drag drops, and the entire ice sheet domain responds with faster flow and surface lowering. The speedups can produce 20–50% increases in ice flux with an overall duration on the order of 1000 a. These flux increases can have rapid onsets (~100 a) when the slippery patches merge.

4. Discussion

Our results show that traveling slippery patches can produce both significant uplift and subsidence within ice sheets. The model produces folds (Figure 4) that are similar to the deformed and overturned ice stratigraphy observed in Greenland [*NEEM Community Members*, 2013; *Bell et al*., 2014] and Antarctica [*Bell et al*., 2011]. The largest observed structures rise to over half the ice thickness. The model produces both anticlines and synclines, mimicking the structures observed in the radar data (Figure 4, cf. Figure 1). The model structures are much larger than equivalent ones produced by stationary patches (Figures S4 and S5). The model predicts overturning within the cores of the uplifted folds, whereas radar data show no reflections within the cores [*Bell et al*., 2014]. The lack of observed features in the cores may be because of data limitations, such as difficulty in imaging contorted layers, or because of processes not included in the model, such as dynamic

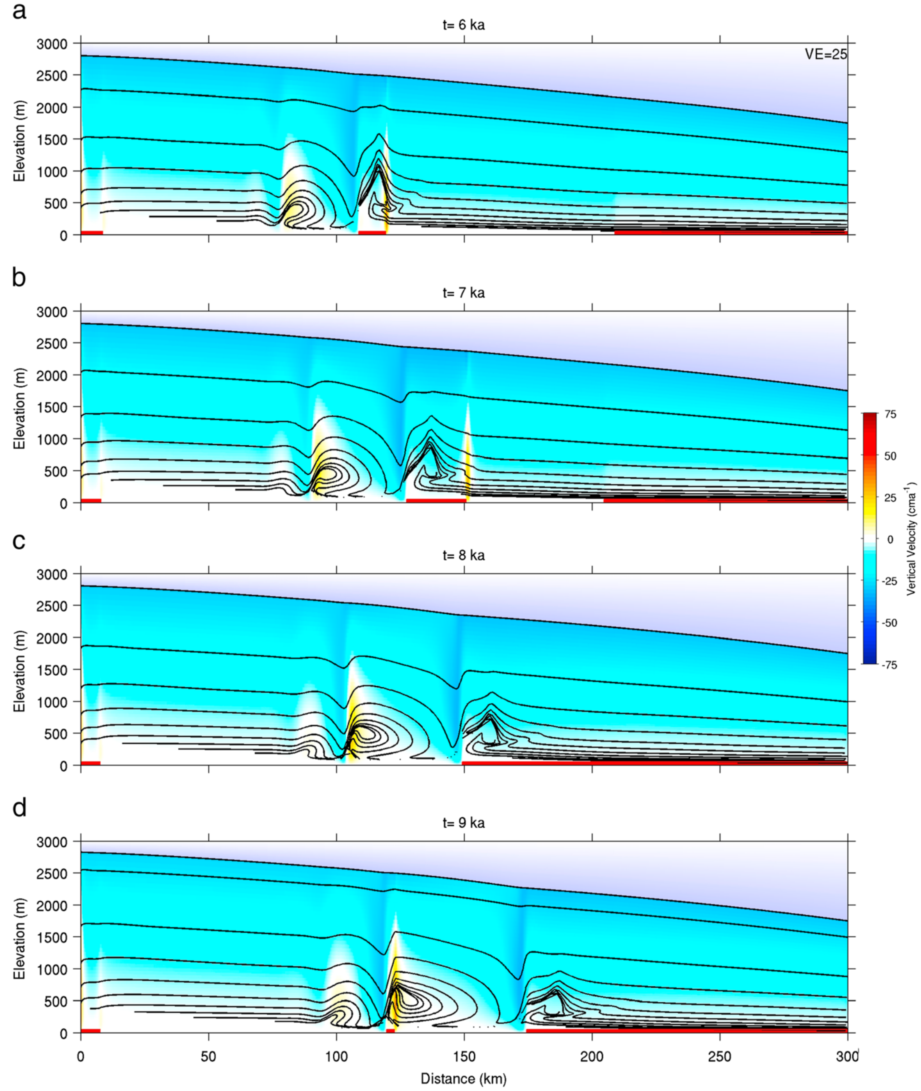

Figure 4. (a–d) Snapshots of the entire model domain showing the development of a mature train of traveling slippery patches. Four snapshots are separated by 1 ka each. Time is measured relative to the peak in the water influx perturbation. Color represents vertical velocity, and lines represent stratigraphy. Red bars on the bottom of the plots represent areas where the bed is at the melting point.

recrystallization [*Cuffey and Paterson*, 2010]. Additional work is required to understand how traveling slippery patches will behave in a complex 3-D environment and how the patches might be modified by bed topography and heterogeneous subglacial geology.

Other proposed mechanisms that can contribute to the deformed and overturned ice stratigraphy include basal freeze-on [*Bell et al*., 2011] and rheological contrasts within the ice column [*NEEM Community Members*, 2013]. A combination of mechanisms likely operates to produce the observed basal structures. For example, our modeled traveling slippery patches contain a small amount (tens of meters) of freeze-on ice. Melting and freezing in our model is limited to low volumes because of latent heat constraints and our choice of boundary conditions. Nevertheless, traveling slippery patches amplify the size of englacial structures beyond the volume of the freeze-on ice alone. Our model includes only variations in ice rheology caused by temperature, but other variations, such as crystal size and fabric [*Cuffey and Paterson*, 2010; *NEEM Community Members*, 2013], could be compatible with our proposed mechanism.

Our model produces substantial variations in ice column temperature and, thus, rheology. Vertical flow pushes warm ice upward at the front of the traveling slippery patches and pulls cold ice downward at the back of the patches. This advection modifies the ice column temperature, creating a heterogeneous distribution of warm, soft areas and cold, stiff areas (Figure S2).

Vertical motion associated with traveling slippery patches can uplift basal ice to about 50% of the ice thickness. The process moves basal debris and deep ice upward. Uplift of debris from ice near the bed [*Lawson et al.*, 1998; *Rempel*, 2008] could delay debris melt-out, including melt-out after iceberg calving. This would aid in the transportation of ice-rafted debris long distances during Heinrich events [*Hemming*, 2004]. Additionally, recovery of old ice normally found near the bed [*Fischer et al.*, 2013] may be easier when that ice is uplifted. However, repeated sections and age inversions may complicate ice core interpretation [*NEEM Community Members*, 2013]. Such complications could have produced the overturned and repeated sections observed in the North Greenland Eemian Ice Drilling (NEEM) core [*NEEM Community Members*, 2013] and potentially played a role in creating the difference between the Greenland Ice Sheet Project 2 and Greenland Ice Core Project cores near Summit, Greenland [*Grootes et al.*, 1993].

Our model makes several predictions about the relationship between the structures in radar echograms and observations that could be made in boreholes. We predict that the majority of the large basal units will have the gas content and geochemical signature of meteoric ice, rather than that of refrozen ice [*Souchez et al.*, 2003], and will be highly deformed with age inversions and repeated sections. If the structures are active, we also predict that synclines mark boundaries in basal thermal conditions, with cold beds upstream and warm beds downstream. This prediction contrasts with the typical assumption that layer drawdown marks the location of concentrated basal melt driven by extreme geothermal heat fluxes [*Fahnestock et al.*, 2001]. However, once the structures become inactive and are passively advected with the flowing ice, they no longer represent variations in basal conditions.

According to our model, stratigraphy responds strongly to moving variations in basal slip. Inversions of basal slip [*MacAyeal*, 1993] from remote observations of surface velocity and ice sheet geometry show heterogeneity in both Greenland [*Sergienko et al.*, 2014] and Antarctica [*Sergienko and Hindmarsh*, 2013], but the temporal variability is poorly constrained. Critical factors for basal slip, such as water availability, sediment coverage and strength, and basal temperature, are difficult to observe directly, but radar stratigraphy can be observed relatively easily. Ice sheet interiors likely have heterogeneous and time-varying basal conditions that leave a record in the ice sheet stratigraphy. Thus, the history of subglacial slip may be estimated from analysis of englacial structures.

This history can include perturbations to basal water flux that trigger the traveling slippery patches. Basal water flux perturbations can be caused by changes in the basal thermal regime or changes in subglacial water routing. Either of these can be related to rapid changes in surface temperature and accumulation rate or ice thickness changes that propagate inland from the margins. These are often related to rapid climate change. We speculate that some of the features observed today in Greenland could have been triggered by sudden climate changes that occurred during deglaciation following the Last Glacial Maximum.

5. Conclusion

Moving patches of subglacial slip create large stratigraphic folds within polar ice sheets as shown in the thermomechanical model used here. These folds create large disturbances to ice sheet thermal structure that feed back to basal temperature and cause the slip patch to migrate. Traveling slippery patches can explain some of the dramatic recent observations of large basal units in Greenland and Antarctica. Our results show that massive thickness-scale folds in ice sheets can be explained with only small amounts of freeze-on. Borehole measurements of ice provenance within the basal units could be used to confirm or disprove our hypothesis. Englacial structures created by traveling slip patches record basal processes and can be used to infer a time-variable history of basal slip. The wide variety of behaviors in ice sheet interiors, including traveling slip patches, suggest that ice sheet interiors can play a critical role expanding our knowledge of ice dynamics and discharge to the margins.

Acknowledgments
Data presented in Figure 1 are available publicly at ftp://data.cresis.ku.edu/data/rds/. Flight date, batch, and file numbers are displayed in the figure itself. All model code available at smb://gravity.ldeo.columbia.edu/gravity/SFTP/PUB/Wolovick_etal_2014/. The authors acknowledge support from NASA and NSF for this paper.

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