The Influence of Marine Topography on the Antarctic Ice Sheet

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THE INFLUENCE OF MARINE TOPOGRAPHY ON THE ANTARCTIC ICE SHEET

A Thesis

Submitted to the Graduate Faculty of the
Louisiana State University and
Agricultural and Mechanical College
in partial fulfillment of the
requirements for the degree of
Master of Science

in

The Department of Geology and Geophysics

by

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B.Sc., Virginia Tech, 2013
May 2015
Acknowledgements

I would like to thank my advisor, Dr. Phil Bart, and my committee members Dr. Nick Golledge, Dr. Karen Luttrell, and Dr. Sophie Warny for their input. I would also like to thank the makers of the Parallel Ice Sheet Model – particularly the ones responsible for maintaining the ‘help@pism-docs.org’ email account. This research made use of resources provided by the LSU High Performance Computing Centre.
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Abstract

The primary controls on Antarctic Ice Sheet (AIS) volume and extent include changes in water temperature, precipitation, atmospheric temperature, and sea level. Here we evaluate the influence of a 5th control, the depth of the continental shelf. Shelf depth affects ice sheet mass balance in the marine environment by controlling the flux of ice that can be exported to the marine environment. This is significant because heat exchange between ocean and ice sheet have been demonstrated to be a dominant control on marine ice volume (Pollard and DeConto, 2009). We used the Parallel Ice Sheet Model (PISM) to simulate AIS dynamics using seven different geologically realistic configurations of a shallower marine scape (i.e., water depth and morphology) to determine how AIS grounding line translations and ice volumes respond to the same climate forcing. Our results show that during advance on an overdeepened and foredeepered shelf, high ablation via calving is such that mass balance is only slightly positive with a net effect of gradual grounding line advance. On a shallow shelf, the relatively low ablation via calving is such that mass balance is more positive – the net effect being that grounding line advance is rapid. For grounding line retreat, we see rapid retreat for deep continental shelf configurations and relatively slower retreat on shallow continental shelves. Additionally, the largest ice volumes occur in the configuration with the shallowest continental shelves. As has been recently suggested (Wilson et al., 2013), these results confirm that relative shelf depth is an important factor to consider when using both direct and proxy evidence to reconstruct AIS history
1) Introduction

The dynamic nature of the Antarctic Ice Sheet (AIS) is primarily controlled by four factors: (1) atmospheric temperature, (2) precipitation rates, (3) variations in sea level, and (4) ice/ocean heat exchange. The magnitudes, times and rates of these changes during the last climate cycle are better constrained than for any other geologic time. In decreasing order of the degree to which these components affect AIS ice volume, the factors are ice ocean heat exchange, sea level, atmospheric temperature, and precipitation rates (Pollard and DeConto, 2009).

Considering that the two most important factors which influence ice volume involve ice-ocean interaction, it is apparent that when examining ice dynamics one must pay close attention to the relationship between ice shelves and the ocean. Heat is transferred between ocean and ice through the ice shelf, i.e., the floating marine termination of the ice sheet. This ice shelf overlies the continental shelf topography below sea level. Over geologic time, this creates an envelope of interaction between ocean and ice that is primarily modulated by changes in sea level and topography (Figure 1). It follows logically that if changes in sea level have already been demonstrated to be an important factor in ice dynamics, then changes in continental-shelf topography should also play an important role.
Our study evaluated the effect of changes concerning a fifth controlling factor, the continental topography. We hypothesized that given similar duration and magnitude glacial-cycle forcings, an ice sheet on a shallow continental shelf configuration will experience relatively fast grounding line advance rates during climatic transitions to glacial periods and relatively slow retreat rates during transitions to interglacial periods. Conversely, the modern overdeepened and foredeepened (OD/FD) topographic configuration would predispose the ice sheet to slower advance during transitions to glacial periods and relatively rapid retreat during transitions to interglacial periods. Our reasoning is that given Last Glacial Cycle (LGC) forcings, the sub ice shelf (SIS) area change from a shallow to OD/FD topography modulates shelf fluxes so that shallow shelf mass balance (highly positive interglacial to glacial, slightly negative from glacial to interglacial) transitions to an OD/FD mass balance scheme (slightly positive interglacial to glacial, highly negative glacial to interglacial).
2) Methods

The Parallel Ice Sheet Model (PISM, Bueler and Brown, 2009, Winkelmann et al., 2011) was used to simulate 122,000 years of AIS dynamics of one Interglacial-Glacial-Interglacial (IGI) cycle over seven different topographic configurations. The magnitude and time of forcings for this interval mirror that of the LGC, but model results are not tuned to create a LGC; the choice was made to create a baseline simulation that looked like the LGC because constraining the most recent advance and retreat of the AIS is made easier by the abundance of data documenting this cycle. The first baseline simulation used the modern continental topographic configuration from the BEDMAP-2 dataset (Figure 2, Fretwell et al., 2013). In addition to the contemporary shelf morphology, six additional topographies were constructed (Figure 3). These additional topographies are progressively shallower versions of BEDMAP2. These topographies have no changes made in their lateral geometries – only shelf depths are changed between each transformation. The modified topographies were created using MATLAB (See Appendix B for display of all modified topographies and initial fields and Appendix C for map view of the differences between modified and original topographies). The domains over which these topographic transformations took place ranged from -300 meters to −2500 meters with respect to sea level. The transformation operated on all values between a prescribed minimum depth (one unique depth per simulation) and to a maximum depth of -2500 meters (deep enough to cover the most overdeepened parts of the interior Antarctic continental topography). Based on the value of an area’s topographic depth within the range from -300 meters (or whichever targeted normalized depth is prescribed) to -2500 meters, the respective topography is shifted up towards the maximum value. It is important to recognize that for a certain normalized topographic depth value, all shallower values were left unmodified. In other words, if a depth of -300 meters was
targeted as a shelf configuration depth, all cells containing topographic values between -300 meters and -2500 meters were modified to become closer to -300 meters with the deeper valued cells receiving a higher modification. Any cells with topographic configurations at or shallower than -300 meters were left unmodified. This allowed for smooth transitions between modified and unmodified cells. Target depths were chosen so that there were a similar number of cells modified between simulations. This process resulted in six modified topographic configurations with ‘normalized’ depths of -600 meters, -525 meters, -485 meters, -450 meters, -375 meters, and -300 meters (Figure 3).

Figure 2: Original BEDMAP2 data set for Antarctic topography (Fretwell et al., 2013). Color bar ranges from 1000 meters to 0 meters below sea level to better display continental shelf topography. Line A-A’ is displayed as Figure 3.
Most glacial modelling experiments consist of two phases, an initial ‘spin up’ phase followed by an experimental phase (Golledge et al., 2014; Pollard et al., 2015). The spin up phase insures that any unphysical ice behavior is allowed to pass before experimental phase data is calculated. Unphysical behavior arises because present and future ice sheet behavior are a function of its past condition, and often times the spin up includes a period where constant forcings are applied over long periods of time until a steady state is achieved. Because a spin up to steady state will result in a different starting ice configuration for the different topographic configurations, the spin up process is uniquely challenging when trying to use bed topography as an independent variable. As such, the spin up process used in this experiment consists of the following: 1) an initialization run, 2) a smoothing run, and 3) an enthalpy development run. The initialization (or bootstrapping) run provided PISM with all unobservable fields necessary to simulate ice motion (e.g. internal ice fields). During this initialization phase, the 5km resolution
dataset was coarsened to 30km. Vertical resolution was 24m at all times. This period was followed by a 100 year smoothing run where ice was free to move but calved at the present day ice extent – i.e. the ice geometry was held fixed. Calving at this specified location allowed for any abnormalities in the upper ice surface to smooth. The third stage of spin up was a 200ky enthalpy development run to allow an internal ice enthalpy field to be established. During the spin up, the initial fields used were topography, basal heat flux, air temperature, precipitation, and ice thickness (Comiso, 2000; Shapiro and Ritzwoller, 2004; Vaughan et al., 1999; Fretwell et al., 2013). During the spin up phase, all forcings were held constant, i.e. there were no offsets applied from the modern spatial variations. In our experiments, each of the different topographies received the same spin up phase leading into the same experimental phase forcings and model parameters.

After the spin up, the experimental phase was begun at a model time of 122,000 years ago. The initial model conditions were set to match those of the modern interglacial (MIG), which we assumed to be a reasonable representation of an average interglacial period of the late Pleistocene. Over the next 102,000 years, the forcings were changed to progressively approach a full glacial state (again, assumed to be approximated by the forcings that describe the last glacial maximum). The glacial maximum climate conditions were then followed by a shift back to interglacial conditions that characterize the modern beginning at ~18,000 years ago. To avoid confusion and comparison to the LGC, our 122ky simulation will be referred to as one interglacial-glacial-interglacial (IGI) simulation.

The IGI forcing factors include the following: (1) atmospheric temperature offsets; (2) precipitation rate offsets; (3) sea level offsets; and (4) sub shelf melt rate offsets (Figure 4). All forcings were applied as deviation from present day values except sub shelf melt rate, which was
applied as an offset from pressure melting derived rates. Additionally, there are no spatial variations in how these offsets are applied to the initial fields. These changes correspond to the climatic changes of the past 122,000 years. This time period represents one full cycle of interglacial-glacial climate changes, which roughly corresponds to the last eccentricity orbital cycle.

The first simulation used the BEDMAP2 topographic configuration. The model was parameterized so that AIS advance and retreat patterns conformed to the geologic constraints from the continent in terms of grounding line location (e.g., Bentley et al., 2014, Anderson et al., 2002), and so that total ice volume progressively increased to a value of ~7.6 million cubic kilometers higher than from the experimental start time (Denton, 2002). These ice volume values are about the same as recent reconstructions of the last glacial cycle (Golledge et al., 2014). Again, our results are not meant as an accurate representation of the last glacial cycle; they are merely a full glacial cycle created for the purpose of comparing the effects of changes in topography. The simulation using the BEDMAP2 topography was used as a baseline against which the simulations for the six progressively shallower topologies were compared. For each 1000 years of model simulation, grounding line location was recorded. At 250-year intervals, a series of scalars including ice volume were also recorded. These measurements of the changing grounding line and ice volume for the different topologies provided a basis of determining the degree to which the topographic configuration affects AIS dynamics. After the initial baseline experiment on the BEDMAP2 topography, each of the six subsequently shallower topographies were modelled using the same climate forcings.
Figure 4: A. Atmospheric forcings were prescribed to change over the 122,000 year run as shown. Precipitation changes were 7.3% for every degree change in temperature (Petit et al., 1999). B. Oceanic forcings were prescribed to change over the 122,000 year as shown Both sub shelf freeze on and sea level were scaled from Imbrie and McIntyre (2006).
3) Results

3.1) Grounding line location

Here we present the results from three key cases – the modern BEDMAP2 (or deep) topographic configuration (Run 1), an intermediate continental shelf depth configuration (Run 4), and a shallow depth configuration (Run 7).

As expected, the modern BEDMAP2 topographic configuration produced the contemporary grounding line location at 122$kya$. As the glacial forcings were applied, the ice sheet and shelves advanced to their maximum extent by 14$kya$. As the prescribed forcings shifted towards interglacial conditions, a period of rapid retreat occurred culminating in an ice sheet configuration in which the grounding line shifted back to the initial configuration (Figure 5).

For the intermediate topographic configuration (Run 4), the grounding line (GL) at 122$kya$ is established in a slightly seaward position relative to the overdeepened shelf (Run 1). By 53$kya$, grounded ice reached the shelf edge. For the next 39$ky$, the GL remained near the shelf edge until roughly 14$kya$ where forcings shifted the GL back towards interglacial configuration. There was relatively minor GL retreat during this transition from glacial to interglacial (Figure 5). All of the GL retreat occurred at places where ice was grounded below sea level, and was most noticeable on the relatively-wide shelves of Antarctic Peninsula (AP) and the Weddell and Ross Sea areas. In this run, the final interglacial GL location was markedly seaward of that which was created at the end of the Run 1 (BEDMAP2 dataset) simulation.

The shallowest topographic configuration (Run 7) produced a full-shelf GL after 15$ky$ of simulated time or 107$kya$ (Figure 5). Grounded ice remained at the shelf edge until the full transition to maximum glacial forcings. After 14$kya$ when interglacial forcings began, there was
no appreciable GL retreat. One of the most noticeable disparities between Run 7 and all deeper topographic runs is the starting GL location. For the Run 7 combination of topography and starting ice thickness, many areas began the simulation with grounded ice.

Figure 5: Resultant grounding line locations from a glacial cycle simulations on an original and modified continental shelf configurations (Runs 1, 4, and 7 with 1 being original, 4 being intermediate, and 7 most shallow). From left to right: starting configuration, glacial configuration, ending configuration. Green color indicates grounded ice, orange indicates floating ice, and red indicates ice free ocean location. Black outline is made to fit Run 1 initial grounding line location.
3.2) Ice volume

In general, for each progressive shallowing of the topography, there was a resultant increase in ice volume present on the Antarctic continent (Figure 6A). The most noticeable changes in the ice volume occurred after a slight delay (roughly 4ky) from the atmospheric and oceanic forcing files provided to the model as input (Figure 4). For the deeper topographies (Runs 1-4) there was a distinct tiering or stair stepping of volume increases over time. The steps in these changes occurred at times of high forcing rate of change – not necessarily times of the lowest or highest forcing. The deeper topographies clearly exhibited higher magnitude ice volume changes over time when compared to the shallower topographies (Figure 6A, Runs 5-7). Also apparent in the plot were the high rates of initial ice volume increase for the more shallow topographies. In general, there was significantly smaller decrease in ice volume for the shallower plots as the ‘interglacial’ forcings were applied.

3.3) Ice shelf area

The ice shelf area is used as a proxy for the sub ice shelf area. While they are not the same, relative changes in a total BEDMAP2 (Run 1) ice shelf area and shallow ice shelf areas (Runs 2 through 7) are representative of the patterns of change in sub ice shelf area. The values are reported by summing the floating ice cells visible from map view. When viewing floating ice area, the deeper topographies show initial higher shelf area values. All values then approach minimum areas of near $3 \times 10^5$ km$^2$ at the time grounding line has advanced to the continental shelf edge (Figure 6B). The observation that all runs approach similar minimum values is indication that all maximum glacial configurations occurred with grounded ice fully at the continental shelf edge (observable in section 3.1).
3.4) Sub shelf mass flux

The flux of ice underneath the ice shelves is reported as a scalar over time called sub shelf mass flux (SSMF). This SSMF is a result of summing the entire flux in or out of the sub ice shelf area. This flux is motivated by both the area (Section 4.2) of the sub ice shelf and the forcing (Figure 4) applied as an offset and given to the model. Because our forcing values are negative, SSMF values are positive – an indication of sub shelf mass gain (Figure 6C). This is also known as freeze on, which is observed under major modern day ice shelves in the WAIS (Liu et al., 2015). This is similar to the approach used by Golledge et al. (2012) in which negative melt rates were used as forcings. The observed data shows similar patterns in the lines as the floating ice area because one is a consequence of the other (Figure 6A). Where there are times of maximum glacial forcings (i.e. 18kya for BEDMAP2), there were minimum values in sub ice shelf freeze on.

3.5) Calving discharge flux

Within the model domain, the physics surrounding free floating ice (icebergs) is poorly constrained (Winkelmann et al., 2011). As such, when ice is calved (Appendix A for calving laws), it is immediately removed from the model domain. This results in highly negative mass flux via iceberg removal (Figure 6D). These highly negative calving discharge fluxes offset the positive mass gain via the sub shelf (Figure 6C) to produce a realistic IGI cycle. Calving and iceberg removal flux least negative values at the same times of maximum glacial forcings were applied.
Figure 6: A. Ice volume plots for all seven simulations over the entire glacial cycle. B. Floating ice area plots for all seven simulations over the entire glacial cycle. C. Sub shelf ice flux plots for all seven simulations over the entire glacial cycle. D. Calving and iceberg flux plots for all seven simulations over the entire glacial cycle. All horizontal axis range from 122kya to 0kya.
4) Discussion

The model simulations strongly suggest that the pattern and timing of AIS response is highly sensitive to shelf depth. These patterns of GL advance/retreat and ice volume change for the overdeepened shelf are consistent with the geologic evidence for the LGC (e.g., Anderson et al., 2014). The simulations are consistent with our hypothesis that (1) grounding line advance on a shallow shelf should be relatively rapid compared to that on an overdeepened shelf and (2) grounding line retreat on a shallow shelf should be relatively slow. In the following two sections (4.1 and 4.2), we show that the differences in grounding-line and ice-volume response (respectively) were primarily controlled by differences in calving flux. In the final sections (4.3 and 4.4), we briefly discuss the limitations and implications of these finding with respect to the task of reconstructing a margin’s long-term glacial history.

4.1) Grounding line migrations

4.1.1) Gradual grounding line advance and rapid retreat on an overdeepened shelf

The baseline BEDMAP2 simulation succeeded in producing a gradual advance and a rapid retreat. Grounding line advances were most appreciable around West Antarctic margins. In comparison, GL translations in East Antarctic margins were small with the exception of the Prydz Bay area (Figure 5, Appendix C). Grounding line advance was coincident with a lessening in shelf area that translated to reduction in both calving and sub ice shelf freeze on fluxes. The grounding line advance was most abrupt at times of high rates of change in forcings – e.g. high rates of sea level fall between marine isotope stages (MIS) ‘5e’ and ‘5d’. The most significant change involved the reduction in the calving flux (Figure 6D) which greatly reduced ablation and drove the mass balance towards positive values. The freeze on flux (accumulation) is greatly controlled by sub ice shelf area. Although the freeze on rate increased from 7 meters per year to
10 meters per year as we transition from IG to G, the total contribution of freeze on to accumulation diminished (Figure 6C) – a result of the sub ice shelf area decrease.

The retreat phase was rapid; in other words no major pauses were observed. This is not consistent with the geologic events which show several ice streams paused multiple times during retreat (e.g. Shipp et al., 1999, Mosola and Anderson, 2006). Our data time series are too infrequent to capture sub-millennial stillstands. The end state GL was equivalent to that at the start of the simulation. Grounding line snapshots show the most change when the grounded ice was on the inner shelf. This is an indication of a reversal in the mechanisms that instigate the initial advance. In other words, the glacial to interglacial forcings drove the GL landward which increased the SIS area and ultimately led to increased calving rates. This effect was most pronounced on the inner shelf where the overdeepened and foredeepened nature of the shelf is greatest.

The LGC history is well documented for many sectors of the Antarctic margin (e.g., Anderson et al., 2014). The simulation is consistent with those available geologic constraints. The rates of advance produced by marine and terrestrial data from Ross Sea sector indicate that the average rate of advance was 15 meters per year. A large amount of information shows that the AIS retreated rapidly back to the modern GL position at rates near 100 meters per year (Anderson, 1999). Comparatively little is known about the pre LGC times in terms of GL translation rates, however geologic evidence shows that large amplitude translations occurred during the Pleistocene (Bart et al. 2011) and Pliocene (Naish et al., 2009). By the most conservative estimates, the continental shelves were overdeepened at these times and hence the results are consistent with the model result showing large grounding line translations on an OD/FD shelf. Bart et al. (2011) report that over the last 0.7Mya, there are only two full
grounding line advances and eight full advances between 2.0 and 0.7Mya (note the average δ¹⁸O cycle of the last 0.9 Mya occurs at a ~0.1Mya period). While these findings represent a small snapshot in geologic time, we believe the findings to be encouraging in that it does indeed seem that grounding line advance to the continental shelf edge is difficult given the modern topographic configuration. Stratigraphic successions are also indicative of a rapid retreat (Naish et al., 2009).

4.1.2) Rapid grounding line advance and slow retreat on a shallow shelf

The more advanced initial grounding line locations to begin the shallow simulations are a result of the raised topography. Thus, the initial (modern) ice sheet emplaced on shallow continental shelves exhibits grounded ice in more locations. When comparing OD/FD advance phase to the advance phase of the shallower shelves, there were many changes stemming from an overall lessening in ice shelf area. Shallower shelves started with less ice shelf area and have less ice shelf area through the advance phase as a whole (Figure 6B). Through these small ice shelf areas, there was a trend of reduction in both mass gain via sub shelf freeze on and mass loss via discharge flux. The sum of the two showed a trend towards more positive mass balance when compared to an OD/FD shelf. The characteristics of these shallower simulations resulted in significantly earlier GL advance to the shelf-edge position (as early as 107kya, Figure 5). In the most shallow cases (Runs 6, 7), the ‘advance’ phase appears as a rapid increase to an ‘equilibrium’ state where forcings have little to no bearing on ice configuration. We believe these results to be consistent with Schoof (2007) where it is proposed that the equilibrium profiles for marine ice sheets favor flatter or normal dipping slopes. In other words, given Pleistocene forcings on a very shallow continental shelf the ice sheet’s equilibrium state more closely resembles that of a glacial maximum configuration.
Any retreat on a shallow shelf does not entail the similar increase in shelf profile (a result of retrograde topography) when progressing from glacial to interglacial periods. The retreat phase on shallower shelves exhibited little to no grounding line retreat post glacial maximum despite forcing parameters returning to that of time 122kya (or full interglacial forcings). We discount the possibility that interglacial forcings were not given enough time to manifest into interglacial ice configurations because the lag between forcing application and response is ~4ky. On a shallow shelf, there is less ice shelf area made from introducing interglacial forcings, and we see a similar pattern of both inward (SSIF) and outward (calving) fluxes staying closer to zero with a combined effect of smaller negative mass balance via shelves.

Not much is known about GL dynamics on a shallow Antarctic continental margin. In fact there is little consensus for what time intervals a shallow continental shelf did exist in Antarctica – with estimates ranging from as recently as the early Pliocene to as late as the Eocene (Hayes and Frakes, 1975, DeSantis et al., 1995, Bart and Iwai, 2012). Significant glaciation in the northern hemisphere began much more recently in comparison to the Antarctic and hence these shelves may not have yet entered into a fully overdeepened phase. Indeed broad sectors of formerly glaciated Arctic margins are shallow compared to the overdeepened configuration of the modern Antarctic shelf. Previous studies have shown that grounded ice existed on Arctic margins (Bjarnadottir et al., 2013). These Arctic margins have a very different configuration than those of the Antarctic. We know of no data that discusses advance of these relatively shallow Arctic margins. Despite these differences in shelf morphology, the retreat rate for the Kveithola paleotrough in the West Barents Sea averages ~40 meters per year during the transition from LGM to MIG (Bjarnadottir et al., 2013). This is a possible indication of stunted GL retreat on a shallow shelf.
4.2) Ice volume on shallow versus deep continental shelves

In general, the ice volume data (Figure 6A) follow patterns similar to the grounding line data (Figures 5). The ice volume data exhibit a tiering of maximum achievable volumes with shallower shelves showing small variance after reaching maximum configuration. Because all simulations have the grounding line reach the continental shelf edge, it is evident that there are variations in inland ice thickness. This trend of larger ice volumes on shallower shelves is in accordance with the idea that ablation is minimized as ice/ocean interactions at the ice shelf are minimized. The exact magnitude of the maximum volume for each simulation is very much dependent on the prescribed flow enhancement factors of the ice (Lliboutry and Duval, 1985, Paterson and Budd, 1982). With increased ice sheet flow enhancement factors, ice deforms more easily in shear and would be deflated vertically by flow to the margin. The result of this increase in flow enhancement would be less disparity between subsequent runs (the maximum achievable volumes would be more similar). Our ice flow enhancement factors for both floating and grounded ice are high compared to recommended values (Ma et al., 2010) which may mean that our predicted ice volumes are under estimations. This contrasts with Wilson et al. (2013) who showed an AIS simulation for the shallowest approximation of Wilson et al.’s (2012) Eocene-Oligocene (E-O) Antarctic topography (which is significantly shallower than the topography of our Run 7) produced a volume of 35.9 million cubic kilometers. However, their study does use atmospheric and oceanic forcings that characterize the climate of the E-O boundary – a climate that was warmer than the generic late Pleistocene climate oscillation used in our study.

The ice volume data clearly demonstrate the pattern in which ice volume increase is significantly earlier and more rapid on the successively shallower shelves. In the case of the shallowest shelf, a volume comparable to the BEDMAP2 maximum volume occurred 5ky after
the start of the simulation. For this same shallowest run, the maximum ice volume is achieved after ~20ky from initialization, with a volume that is 64% higher than the maximum volume achieved on the BEDMAP2 simulation. The pattern of ice volume change for the OD/FD shelf is consistent with that of the δ¹⁸O record for the LGC with smaller scale oscillations corresponding to individual stages with a small (~4ky) offset between forcing and ice volume manifestation. On the successively shallower shelves, the stage oscillations are progressively suppressed as high ice volumes are reached earlier. There is no geologic proxy for ice volume changes on a shallow shelf to which these patterns can be compared, but the difference between the shallow and the deep topography plots show that the pattern of ice volume changes are very sensitive to shelf depth.

During the transition from glacial to interglacial forcing, the shallower shelves are more likely to remain at high volumes whereas the BEDMAP2 run returned to the initial interglacial volume. In contrast, at time zero (simulation end), the three shallowest simulations (Runs 5, 6, 7) have experienced only a small ice volume reduction – leading to end-state ice volumes higher than the maximum ice volume on the BEDMAP2 simulation (~33.5 million cubic kilometers).

Judging from the grounding line translations and the larger grounding line translations around the WAIS margin, it would seem that most of the added ice is in West Antarctica. This is consistent with geologic data suggesting minimal change in ice thickness to the EAIS during the LGC (Anderson et al., 1999). Flattening of the continental shelf topography removes some of the dynamic topography that allows for both the initialization of ice streaming and pooling of sub ice water which enhances flow (Stokes et al., 2014, Bennet, 2003). Shutting down the ice streams removes critical ablation highways which further enhance the removal of ice from the continent. The degree to which discrete trough development and general overdeepening separately affect
ice dynamics is not easily extracted from the data because our topographic manipulation reduces trough relief in concert with overall continental shelf depth. That being said, recognition of the timing of each event may prove useful in evaluating their separate impacts.

4.3) Model limitations

As noted in earlier, the forcings we used were chosen so that the BEDMAP2 replicated a LGC-type grounding-line oscillation and ice-volume change. The model reproduced gradual grounding line advance to the shelf edge during the interglacial to glacial transition and rapid retreat during the glacial to interglacial transition. In terms of the oceanic heat transfer forcing, we chose freeze-on because any melting caused the GL to rapidly retreat throughout the simulation. This would mean that sufficiently warm waters on a shallow continental shelf would cause GL retreat. In other words, our data should not be taken to mean that an ice sheet once expanded would never retreat. Moreover, modern data shows that there are many areas of freeze on while other areas are experiencing melt (Liu et al., 2015).

It is important to remember that during our experiments we only changed one facet (topography) of a very complex system. A necessary point to address when considering ice dynamics in our simulations is the effect of isostatic rebound which was not implemented in our model simulations. It has been shown that proper coupling of earth (and ocean) models with AIS simulations since the LGM show that there can be a stabilizing influence on grounding line retreat (Gomez et al., 2013). We did not incorporate isostatic adjustments into our simulations because shallower topographies created more ice volume. This greater ice volume would depress the landscape and create similar topographies – a converging of our independent variables for which we wished to keep a broad range. We acknowledge the stabilizing effect of the rebound, but only wish to evaluate the effects of a static change in topography on ice dynamics. On a
similar note, sedimentation may also (on a long time scale) insulate ice shelves from small sea level changes (Alley et al., 2007). Simply modifying the topography would affect the thickening of a thermally isolative sediment blanket for shallow topography runs (creating less basal melt, and therefore less ice streaming). Streaming on the OD/FD shelves may occur due to channeling of melt water into topographic lows (which are removed by our topographic transformation). The channeling of melt water has been shown to be an important condition in initiation of ice streaming (Livingstone et al., 2012). These areas of grounded ice which have high velocity favoring conditions would then pass on high velocities into the ice shelves.

The initial partitioning of grounded and floating ice cells also play an important role when considering ice volume. Where a modern ice thickness on a shallow continental shelf area may accommodate large areas of grounded ice, the opposite is true for a deep shelf. The implication for more ice shelf area is an added dimension that is susceptible to oceanic fluxes. Not included in this model, yet worthy of note is the feedback between ice and ocean. Any changes in sub shelf melt on affecting the heat or salinity composition of nearby ocean cells is not modelled. Similarly, the normalized topography removes ice-eroded troughs through which warm Circumpolar Deep Water (CDW) is directed towards the ice shelf (Rignot, 2002). It is unclear what smearing this spatial distribution of warm water would play on the simulations here, but the conceptualization of the impact of warm water intruding on the shelf and interacting with the sub ice shelf has been explored in other studies (Bart and Iwai, 2014, Golledge et al., 2014) where it is predicted that the warm water intrusions yielded rapid melt phases.

4.4) Implications for reconstructing AIS ice-volume history from proxy data

If our hypothesis is correct, the evolution of the Antarctic continental shelf from a shallow basinward-dipping margin to an overdeepened and foredeepened margin should have
influenced how the AIS responded to the orbitally-paced glacial-interglacial climate forcing through time. The model results are potentially significant because they show that shelf topography exerts an important control of ice-sheet dynamics. At present, reconstructions of long-term glacial history are primarily derived from proxy data, i.e. composite deep sea benthic foraminifera oxygen isotope records (e.g., Zachos et al., 2001). Interpretations of these proxy data tacitly assume that the continental interiors and margins surrounding Antarctica have always been foredeepened and overdeepened (Bart, 2004). However, Wilson et al. (2013) suggested that the continental interior in the Eocene was significantly higher than present, with much of the West Antarctic interior being subaerial. By extension, the continental shelves were shallower and basinward dipping prior to the development of a continental scale AIS that evolved during the early Oligocene. This is consistent with Hayes and Frakes (1975) who proposed that Ross Sea overdeepened during the early Oligocene based on biostratigraphic evidence. However the exact timing of overdeepening is poorly constrained. DeSantis et al. (1995) proposed that Ross Sea overdeepened 20 million years later during the late Miocene based on backstripping experiments. Bart and Iwai (2012) proposed that the Antarctic Peninsula margin overdeepened 15 million years later during the early Pliocene based on seismic stratigraphic and biostratigraphic evidence. These different assessments and the possibility overdeepening may not have been a synchronous event across the continent may require that we rethink certain aspects of how proxy evidence should be used to interpret the long-term glacial history.

Omitting that over time, glacial processes can build out the continental shelf break to create more cumulative continental shelf area, evidence of a steady transition from shallow to deep continental shelf configuration may appear in proxy records as a gradual decrease in the maximum possible contribution of the AIS to global ice volume. This implication is seemingly at
odds with the general δ¹⁸O record from 55Mya onwards (Miller et al., 2005) which exhibits a
general increase in the δ¹⁸O value up to the present. However, the δ¹⁸O record of the past million
years does show an increase in the magnitude of ice volume oscillations between glacial and
interglacial periods which is similar to our predictions for ice volume on an OD/FD continental
shelf. One possible explanation to reconcile these two contradictory aspects is that the δ¹⁸O
record is dominated by northern hemisphere ice sheets for the Pliocene and Pleistocene. For the
early record (solely southern hemisphere signal), the abrupt shift at the E/O boundary suggests
that the AIS expanded rapidly. The δ¹⁸O shows small oscillations in the ice volume signal which
is consistent with our shallow shelf predictions.
5) Conclusions

Our modeling results showed that the shelf depth and profile significantly influence AIS mass balance during a generic late Pleistocene glacial cycle. Mass balance changes affecting grounding line and ice volume were primarily driven by large reductions in the flux of ice shelf calving on shallower shelves versus the BEDMAP2 shelf. During the transition from interglacial to glacial states, mass balance was successively more positive on the shallower shelves, resulting in progressively earlier, rapid grounding-line advance, well before peak glacial forcings were applied. The continent surrounded by the shallowest shelf hosted the largest ice volumes. For successively shallower shelves, the respective maximum volume was reached earlier. Our results have implications for the glacial history reconstructions from proxy data because AIS response would change as continental shelf depth evolved.
References


Appendix A: PISM description

Appendix A Figure 1: PISM’s view of interfaces between an ice sheet and the outside world (from the PISM manual).

ICE/AIR: For the experimental portion of the PISM simulations, the surface of the ice was modelled using precipitation field reinterpreted as climatic mass balance and mean annual air temperature was evaluated as the temperature below the firn layer. This is called the ‘simple’ model. Offsets were applied to the surface mean annual air temperature via a ΔT input file described in the methods (Figure 4). Precipitation was interpreted as changing from the modern precipitation field by 7.3% per degree temperature change after Huybrechts (2002).

ICE/OCEAN: The ocean model supplied was ‘constant’. Sub shelf ice temperature is set to pressure melting and the sub shelf melt rate is assumed to be proportional to the heat flux from the ocean into the ice (PISM Manual). Changes were applied to the sea level and to the sub shelf melt rate via a file outlined in Figure 4.

ICE/BEDROCK: No isostatic adjustment was incorporated in this experiment. Negative feedback loops would lead to converging independent variables. While this is an important
factor, it is not something that we wanted to test. The hydrology model was set to ‘null’ (the default) meaning that water is not conserved. It is stored in the till up to a specified amount. This is based on the undrained plastic bed model. The ice sliding parameterization was set to a pseudo plastic power law model with a ‘q’ value of 0.25. Till effective fraction overburden was set to the default value of 0.02. Basal yield stress at grounded, sub sea level ice cells was reduced by invoking the option of ‘tauc_slippery_grounding_lines’. Till angle was set heuristically using values from the ‘searise-antarctica’ example scripts. These values are 15, 40, -300, and 700. The first two values correspond to the minimum and maximum angles, and the latter two values set a topographic range for which these values are scaled between.

GENERAL ICE PROPERTIES: Stress balance was set to the combined SSA and SIA model. The enhancement factors were 1.9 and 6.0 respectively. Our SSA enhancement factor is high, but produced the best baseline in terms of ice volume and grounding line location. The finite difference (fd) option was set for the SSA numerical solver.

CALVING/SHELF EDGE: Ice calving was set to remove sea ice after it became less than 200m thick or set to eigen calving with a ‘K’ value of 2.0e18. Eigen calving is a physically-based calving parameterization based after Levermann et al. (2002) and Winkelmann et al. (2011). Additionally, ‘cfl_calving’ was set to restrict PISM’s adaptive time-stepping. This is supposed to eliminate dendritic ice shelf structures, but there were still some apparent. The ‘pik’ option was also invoked which does four things: allows for stress boundary conditions at the calving front, removes all icebergs, allows the shelf edge to partially advance (it can move in less than the 30km cell width), and conserves mass for those partially advance ice shelves.
Appendix B: Initial Datasets

Appendix B Figure 1: Modern topography (BEDMAP2 dataset, Fretwell et al., 2013). This topography is referred to as Run 1. Color bar was selected to highlight the largest area of features below sea level.
Appendix B Figure 2: Modified topographies. Color bar ranges from 0 to -1000m relative to sea level to highlight features below sea level. With respect to run number, the topographies were averages to roughly just deeper than: -600, -525, -485, -450, -375, and -300m.
Appendix B Figure 3: Initial climate and ice thickness values applied at initialization of the simulation (Comiso, 2000, Shapiro and Ritzwoller, 2004, Vaughan et al., 1999, Fretwell et al., 2013). Annual mean air temperature and ice accumulation receive offsets from the 'atmosphere' forcing files. Basal heat flux is used to determine heat flow into the base of the ice. Ice thickness is the original ice field emplaced on the topography.
Appendix B Figure 4: Comparison of the difference between the modified topography for each run and the original topography. The ‘z’ axis (the color) is a measure of topographic addition. Note that for the color scale values are capped at +600m to better show differences in between runs. This is not an indication of the maximum topographic additions.
Appendix C: Full grounding line map for each run

Appendix C Figure 1: Grounding line location for every simulation at times of: initialization, glacial maximum and simulation end. Each row shows progression for one simulation. Green indicates grounded ice, orange is ice shelves, and red is ice free ocean. Travelling down the first column, it is apparent that shallowing of topography (down the column, or from Run 1 to Run 7) the floatation criterion are changed i.e. there is more grounded ice. Travelling down the second column, the time to glacial maximum should be noted. Glacial maximum conditions are achieved slightly earlier. The third column shows the trend of stunted retreat phase as topography shallows.
Vita

Dan attended Virginia Tech as an undergraduate. He graduated in 2013 having received his Bachelor’s in Geoscience with a specialty in geophysics and a minor in Math. Dan attended Louisiana State University where he majored in Geology. During the course of his degree at LSU, he was fortunate enough to have the opportunity to present his research at both the Scientific Committee on Antarctic Research (SCAR) 2014 meeting in Auckland, New Zealand and the Geologic Society of America (GSA) 2014 conference in Vancouver, Canada. At LSU Dan was a recipient of an Applied Depositional Geosystems (ADG) Fellowship and additionally received funding via ExxonMobil’s recruiter nominated grant system. Dan spent his final semester at LSU aboard the ice breaker/research vessel (IB/RV) Nathaniel B. Palmer performing research in the Ross Sea, Antarctica. Upon completion of his degree requirements, Dan will spend the summer working as an intern in the Geoscience Workflows arm of ExxonMobil’s IT department.