Reviewer 1:

In this paper, the authors use a state-of-the-art ice sheet model to carry out (slightly compressed) glacial interglacial simulations on the Ekstrom ice catchment and ice shelf. They use a stokes ice-sheet model with a grounded to floating transition and step forward the ice-sheet model over time scales considerably longer than those generally considered in such simulations. By make use of two different linear system solvers, as well as testing the response to a highly uncertain physical parameter dealing with the nature of the sea bottom in the ice-shelf cavity (at least, this is my understanding, see below). The improved numerical solver shows great gains in terms of computational cost, and the bed geology of ice-shelf cavities is shown to play a large role in fluctuation of ice-sheet volume and potentially lead to hysteretic behaviour.

I believe this paper can potentially by published in this journal. The simulations they have done are impressive from a computational standpoint alone, but also raise interesting questions regarding our ability to model past behaviour of ice sheets when we know so little about the marine geology of ice-shelf cavities. I have a few general comments, and a number of detailed comments, however, that should ideally be addressed.

We thank the referee for the thoughtful and thorough review of our paper. We appreciate you taking the time to complete the reviews and welcome your helpful comments. We have revised the manuscript to address your review comments (see below). Throughout this response to review document your (referee review) comments are provided in regular, non-italic font text, our response comments are provided in red font (as here).

General Comments:

1. Though this is quite specific, it is quite important, and i would like to see it clarified, ideally in the response and in a revision. The central science result hinges around the effect of different bed strength. The methods seems to suggest that, between the “hard bed” and “soft bed” runs, the only difference between the bed frictional coefficient (C) is in areas where this CANNOT be inferred from an inversion of velocities as described in 3.4 â ˘A ˘T in other words, bed within the current ice-shelf cavity â ˘A ˘T meaning in both experiments, C is identical in currently grounded areas. Is this correct???

Yes, this is correct. We are only investigating the effect of different geology underneath present-day ice shelves (floating parts), while the inferred distribution of the friction coefficient (C) for the present-day ice sheet (grounded parts) is the same in all simulations.

I ask because section 3.4 would imply this, though i could not find any other part of the paper that made this clear. If the only difference is indeed below currently floating ice, this has very strong implications; however, i fear that (a) i have misunderstood and (b) even if i have not, other readers might. This aspect of the methodology should be pointed out with crystalline clarity to the point that maybe even the experiment names should change to emphasise this. (And i should add if “hard bed” means 10-1 everywhere and soft bed means 10-5everywhere, then the results overall are not very surprising â ˘A ˘T so this is why it is general point #1)

To make this more clear, we have added the following to the introduction: “Here, we present the first regional scale FS simulations investigating the effect of different ocean bed properties under contemporary ice shelves on ice sheet geometry over a glacial cycle.”
We also added the following statement to the model initialisation: “The model is initialised to the present day geometry using the commonly applied snapshot initialisation in which the basal traction coefficient C is inferred under the grounded ice sheet by matching observed surface velocities with modelled surface velocities.”

Please note that in the Experimental Design section, we also explicitly state that different values for basal traction are only applied underneath present day ice shelves: ”We consider two end member basal property scenarios by prescribing either soft ocean bed conditions (mimicking sediment deposits) or hard ocean bed conditions (mimicking crystalline rock) under all present day ice shelves in the modelling domain.”

2. A key scientific result put forth is that of hysteresis with a strong (ice shelf cavity?) bed, in that the grounding line (GL) does not return to its original position. This is used to argue that even without a retrograde slope (line 364) there can be hysteretic behaviour. I point out that there have been previous studies suggesting that a continuum of grounding lines were possible, but these (and other) authors later showed that correct treatment of grounding zone boundary layers removed this degeneracy, but this treatment involved resolving the grounding zone, the length of which scales inversely with bed strength. In the context of Stokes, Nowicki and Wingham (2008) found that with an effectively non-sliding bed, there was not a unique sliding solution in the presence of a frozen grounding zone. While the authors’ results are interesting, they should allow for the possibility that (a) the grounding zone is not sufficiently resolved or (b) there is not a unique solution to the Stokes contact problem with an effectively non sliding bed (therefore raising the question of whether the model finds the physically correct one) rather than assuming that the model results are correct, and hysteresis of ice sheet is possible without retrograde beds.

We thank the reviewer for pointing this out. We have added the following sentence to the discussion section of the modelled hysteresis to acknowledge this: “However, this result could also be caused by a combination of the non-uniqueness of the Stokes contact problem for non-sliding beds and an under resolving of the grounding line zone (e.g. Nowicki and Wingham, 2008).”

3. There are extensive mentions of ensemble modelling in the paper; while you do not say outright you are doing ensemble modelling, you don't say that you are not (aside from a mention that your approach is "complimentary" to it, line 310, which is confusing; it is not ensemble modelling because it does not vertically average?) I would argue you tested 2 end members of a (albeit important) physical parameter (the choice of solver is not a physical parameter), so perhaps you should be as clear as you can be that this is NOT an ensemble of experiments.

The reviewer is correct that we are not presenting ensemble simulations, but provide a pair of envelope simulations covering two extreme scenarios. We have added a sentence to the introduction to state this explicitly. It reads: "We do this by investigating end-member scenarios as opposed to ensemble modelling."

We also expanded on the statement in line 310 now explicitly stating that our approach of using a complex ice-physics model investigating end-member scenarios and ensemble modelling using simplified ice physics both have their advantages and disadvantages, but both are worth pursuing.

It now reads: "For example, radar isochrones for floating ice shelves could be incorporated more easily into the model tuning, because the FS approach does not apply a vertical average in these areas unlike ice models using a simplified force balance. We believe that ensemble modelling using simpler ice physics models and our approach of employing a complex ice-physics model and investigating end-member scenarios can both provide different new insights. Hence, both approaches should be pursued in future. This also holds for shallow ice approximation-FS hybrid approaches (Ahlkrona et al., 2016) which can build on the results shown here."

A series of 4 experiments are done, varying one of each: bed strength, and numerical solver â˘A˘T and the results of the paper are presented as dual: the effects of the hard bed, and the effectiveness of the solver. It is therefore confusing whether this paper is meant to be about numerics (in which case it might be better suited for a different journal) or about the scientific results. Both aspects are presented quite prominently making the message of the paper a bit unclear. If the paper is to be about science, then aspects dealing with numerical methods such as scaling should perhaps be in an appendix and not feature in the abstract (though i do have comments about these aspects as well).

The goal of the paper is indeed two-fold. One goal is to show that progress towards faster full-Stokes simulations is being made, so that in the near future simulations over longer time scales (>1,000 years) with this type of model should be possible for regional domains. As the scaling of the ParStokes solver is integral to the speed-up in computation time, we would like to keep this Figure in the main text.
The second goal of the paper is to present the effect of different geology under present-day ice shelves for ice-sheet evolution over a glacial cycle. The main scientific messages are however independent of the solver setup. We tried to make

this separation clear by putting technical and scientific results and their discussion in separate sections. However, to make our intentions clear to the reader from the beginning, we have rewritten the last paragraph of the introduction.

It now reads: “Here, we present the first regional scale FS simulations investigating the effect of different ocean bed properties under contemporary ice shelves on ice sheet geometry over a glacial cycle. We do this by investigating end-member scenarios as opposed to ensemble modelling. This means, we specify either very soft and slippery or very hard and sticky conditions under present-day ice shelves. The goal of the paper is hence two-fold. First, we present methodological advances by extending the feasibility of regional FS ice sheet simulations by an order of magnitude using the open source code Elmer/Ice (Gagliardini et al., 2013). We do this with a highly parallelised numerical scheme allowing to maintain a high mesh resolution (∼1 km) and a freely evolving grounding line over glacial/interglacial timescales. Second, we present new scientific insights regarding the effect of different ocean bed properties seawards of today’s grounding line and quantify its impact on the evolution of the entire catchment. This is done for the Ekström Ice Shelf catchment, Dronning Maud Land, East Antarctica (Fig. 1).”

Detailed comments:

line 73, data of shelf cavity â ˘A ˘T should point out this is only relevant to the present study *up to the farthest point of grounding line advance* in your experiments.

Correct. We added the following sentence: “For our simulations, this difference is only relevant up to the point of farthest grounding line advance.”

line 174 and potentially elsewhere; please say something about the FEM basis functions in your scheme(s). It is important to establish that the basis functions are LBB conforming and that the solutions are exactly mass conserving (ie. not using penalty methods) â ˘A ˘T the latter perhaps not being as important for short term runs but very important for long term.
line 174 how many cells? how many DoFs?

We have added this information to the mesh generation and refinement section.

It now reads: “The 3D mesh consists of ∼200,000 nodes and therefore ∼800,000 degrees of freedom. We are using stabilised P1P1 elements and an algorithm that deduces a mass-conserving nodal surface to avoid artificial mass loss (Gagliardini et al., 2013).”

line 203: are you sure? all physical uncertainty? what about ice shelf crevassing weakening? Not to mention these physical parameters, if i understand correctly, are only varied in the ice-shelf cavity (see General Comment 1)

The reviewer is correct. We qualified this statement. It now reads: “ Hence they also account for some uncertainties in model parameters, forcings, and physics of the applied ice sheet model.”

line 223: since this is exceptional is it really of value for general knowledge?

We agree that there is not much value for the compute times using a now decommissioned system. We therefore removed the paragraph from the manuscript and also deleted the corresponding compute times in Figure 4.

Also:

a) it is odd to compare one solver on one system and another on another system. How about an additional test (only a few time steps) of both on the same system, with walltimes so a comparison can be made.

We agree that this would be the preferred option. However, larger jobs get priority on the SuperMUC-NG and hence queuing times would be much longer on this system for smaller jobs like the MUMPS jobs. We did however test whether absolute compute times for a few time steps between the systems are similar. As this was the case, we believe that the absolute numbers provided in Figure 4 are most informative.

b) ParStokes has great scaling but what about absolute time for a fixed core count on the same system compared against MUMPS?

We added the following statements to the solver setup comparison section (sec 4.1): “This speed-up is in part due to using more CPUs in the ParStokes simulations. When comparing absolute runtime of the scaling simulations, ParStokes provides faster computations for >168 CPUs. This means the minimum requirement for faster simulations with ParStokes is a supercomputer with more than 168 CPUs. The exact CPU number may however very well vary from system to system depending on the available hardware.”

line 228: following on from comment on line 174, which of the two uses a stabilisation method? if not both, then what about the other one?

They both use stabilisation methods, but different ones (stabilized method for MUMPS and bubble stabilisation for ParStokes) for stability reasons. We added the following sentence: “ When using the MUMPS solver, the stabilised method is used, while for the ParStokes solver we use bubble stabilisation (Gagliardini et al., 2013). This results in slightly different systems that need to be solved.”

Figure 8: Here or in an appendix you should show a similar plot comparing ParStokes and Mumps for soft and hard bed (whichever shows poorer agreement). It is important to establish that the effect of solver, while having a large difference on performance, has very small effects on Volume and GL position relative to the

effect of the physical parameter. If MUMPS and ParStokes differ, at most one is correct â ˘A ˘T if the difference is large, how are we to trust the physical results?

We do not think that this proposed Figure would add information that is not already there. We are showing grounded area differences in Figure 5 which is indirectly a measure for the differences in ice volume. In addition, Figures 6 and 7 show differences in grounding-line position and ice thickness between the solver setups for different time snapshots. We believe that these Figures provide a solid case that differences between the solver setups are small.

line 250: following what?

Changed to: “Following the period of volume gain, ...”

line 257: then there is a volume decrease â ˘A ˘T what causes it?

Apologies, but we could not find the statement of a volume decrease the reviewer is referring to.

line 289, some funny maths. How does increasing performance by a factor of 6 allow runs of 40,000yrs when using the MUMPS solver allowed less than 1000a? this is not a factor of 6. did you do something to increase the time step that was not mentioned?

I think we confused the reviewer here. These numbers have nothing to do with each other. The speed-up refers to the difference in computation time between MUMPS and ParStokes. MUMPS does allow computations of 40,000 years, they just take 3-6 times as long as with ParStokes. But given enough time, these type of runs are possible. To avoid this confusion we removed the 1,000 year maximum from the sentence.
It now reads: “The new setup allows 3D full-Stokes ice sheet simulations on the regional scale over 40,000 years now in under a month’s time.”

line 349: explain what you mean by grounding line
We changes this to: “The earlier onset of grounding line motion in the retreat phase ...”

Reviewer 2 (Stephen Cornford):

This paper describes the application of a full-Stokes ice sheet model to a modest sized region (about 100 x 300 km) of Antarctica over 40,000 years at a reasonable 1km resolution. Full Stokes models are computationally expensive and have typically been used only for shorter simulations: various approximations are normally applied (quite often at coarser resolution too). The paper also makes use of new data and explores the importance of bed friction in the region, so would be of interest even if it did not manage the full Stokes model. Give the use of the higher-fidelity model, this is an important and clearly written paper. I have a few minor comments only.

We thank the referee for the thoughtful and thorough review of our paper. We appreciate you taking the time to complete the reviews and welcome your helpful comments. We have revised the manuscript to address your review comments (see below). Throughout this response to review document your (referee review) comments are provided in regular, non-italic font text, our response comments are provided in red font (as here).

The abstract perhaps emphasises the Stokes model, but there are two conclusions in the paper – one relates to the importance of the basal boundary condition (sliding law), which might have been reached with a more approximate model. At the same time, there is no less-than-Stokes model considered, so the paper provides us with no information on whether 'uncertainties due to physical approximations [have been] be reduced.', at least compared to the uncertainties that would be common to models (e.g the sliding law)

We agree that our paper does not show that uncertainties due to different physical approximations have been reduced. Rather, the goal of the paper is to provide a first step towards being able to do this in the near future. To reflect this appropriately in the text, we changed the corresponding sentence to: "Therefore, there is a need to extend the applicability of regional FS ice sheet models to timescales longer than 1,000 years so that uncertainties due to physical approximations in the force balance can be quantified and reduced in the near future."
We also agree that at least qualitatively, we could have reached the same conclusions regarding different levels of bed friction with a more approximate model. However, the magnitude of grounding-line advance and retreat over such a long time period will most likely be different across different ice mechanical models. This has been shown in the previous intercomparison studies using idealised geometries (e.g. Pattyn et al. 2013).

Specific comments:

L42: "The rationale behind this tuning is that if the model matches the constraints well,then confidence is high that the model also reproduces ice sheet changes at other times. The risk involved is that the matching may overcompensate for the simplified model physics leading to higher uncertainties in future predictions where model constraints are absent" I don't disagree with the overall statement, but I would suggest that the rationale is simply that if a model matches constraints poorly, then it should be rejected (or given a lower score).

Agreed. We changed this accordingly.

L105; The thermodynamic equation – how is temperate ice treated?

We added the following: "The ice temperature T is bounded by the pressure melting point T_m, so that $T \leq T_m$."

L153 "A linear viscous sliding relation (m=1) was chosen to guarantee consistency between model intialisation and forcing simulation." This is not needed – the inverse problem provides both C1 and |ub|, so you could carry out runs. with any value of m so long as C1|ub|=Cm|ub|m. Linear sliding is probably the worst choice (see e.g Joughin 2010) and although many (me included) have used these rules in the past, as a community we should move on. I am not suggesting new runs, but an acknowledgement that the authors understand this position.

Yes, we are aware of this and agree with the reviewer here. We changed this sentence to read: "A linear viscous sliding relation (m=1) was chosen. Alternative and physically more realistic sliding relations exits (e.g. Joughin et al., 2019) and the consequences of our choice of using a linear sliding relation on the results are discussed below (see section 5.5)."

L183 "While robust, direct solvers do not take advantage of the sparse structure of the matrix and require large amounts of memory." That is certainly true of e.g. LAPACK solvers, but the MUMPS solver is the MUltifrontal Massively Parallel sparse direct Solver, designed for these sorts of problems. That is not to say that an iterative solver has no advantages, but frontal solvers like MUMPS are specialised over general dense solvers.

We agree with the reviewer here. Our formulation was not precise enough. MUMPS is certainly tailored towards solving large sparse linear systems. However, the fact that it remains a direct solver still leads to the solver being memory bounded. Therefore, it does not scale at all beyond 80 CPUs. We adjusted the sentence as follows: "While robust, direct solvers require large amounts of memory."

L226 "We note however that we do not expect a perfect match between the two solver setups due to small differences in the finite element formulation" This needs a bit more emphasis/elaboration. If you were solving the same problem, you would expect the solvers to give the same results (assuming the iterative technique was successful). But the problems are different?

Yes, the problems are slightly different due to different stabilisation methods employed by using MUMPS or ParStokes (see response to other reviewer).

ParStokes does seem to work well though (I would have liked see SSA in the same comparison, but in a follow up paper, perhaps).

We agree that this would have been interesting. However, as of today there is no thermomechanical coupling available when using reduced models in Elmer/Ice (e.g. SSA, SSA*) and that's why we did not perform the same simulations with a reduced model.

L220 "For both simulations, there is good agreement in terms of grounding line position over time, with differences never exceeding 5% (Figure 5)." - the difference is in total grounded area.

Yes. Thanks for spotting this. Changed accordingly.

L264: "Stable grounding line positions for both simulations are associated with periods of ice sheet stability (Fig. 8). " Steady rather than stable? I agree that you are unlikely to see unstable equilibrium in practice, so steadiness tends to imply stability.

Yes, steady might be the better term to use here. Changed accordingly.

L289 "The high mesh resolution required to adequately capture grounding line migration (Pattyn et al., 2013) is hereby maintained.".
Perhaps – there is no convergence study in this paper so it relies on external references, and the only Stokes model in Pattyn 2013 is Elmer/Ice which ran at around 50 m resolution.

The reviewer is correct that we did not perform a convergence study. Given the runtime of the model, we do not think it is feasible to carry out a convergence study for long-term simulations at the moment. Moreover, a mesh resolution of 50 m is certainly only ever applied in simplified settings and for shorter simulation times. To acknowledge the fact that we cannot show that this resolution is adequate, we reformulated the sentence as follows: "We hereby maintain a mesh resolution (~1 km) that is finer than in most other paleo ice sheet simulations (Pollard and DeConto, 2009; Golledge et al., 2014; Albrecht et al., 2020) albeit at a regional scale."

L385 "The difference between Weertman and pressure limited relations is that the latter take effective pressure into account. This means that basal drag goes to zero near the grounding line and reduces to a plastic sliding relation (Brondex et al., 2017). However, this lower basal drag area is limited to a few kilometers upstream of the grounding line."
There is another important difference, which is the independence of Tb and |u| in the region in question, which could be substantial. See for example Joughin

2019 which provides evidence for Coulomb-like sliding a long way from the grounding line. No need to speculate, but please, acknowledge Joughin 2019 Â ˘ahttps://doi.org/10.1029/2019GL082526

We have expanded this section and added the reference. It now reads: “The difference between Weertman and pressure limited relations is that the latter take effective pressure into account. This means that basal drag goes to zero near the grounding line and reduces to a plastic sliding relation (Brondex et al., 2017). This results in the basal drag becoming independent of the sliding velocity. Most previous studies using pressure-limited relations confine areas of lower basal drag to within a few kilometers upstream of the grounding line (e.g. Schannwell et al., 2018; Brondex et al., 2019). There is however evidence from observations and modelling that areas of low basal drag can extend much farther inland (Joughin et al., 2019).”

Pattyn, F., Perichon, L., Durand, G., Favier, L., Gagliardini, O., Hindmarsh, R., . . . Wilkens, N. (2013). Grounding-line migration in plan-view marine ice-sheet models: Results of the ice2sea MISMIP3d intercomparison. Journal of Glaciology, 59(215), 410-422. doi:10.3189/2013JoG12J129

Quantifying the effect of ocean bed properties on ice sheet geometry over 40,000 years with a full-Stokes model

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Abstract. Simulations of ice sheet evolution over glacial cycles requires integration of observational constraints using ensemble studies with fast ice sheet models. These include physical parameterisations with uncertainties, for example, relating to grounding line migration. Ice dynamically more complete models are slow and have thus far only be applied for <1,000 years, leaving many model parameters unconstrained. Here we apply a 3D thermomechanically coupled full-Stokes ice sheet model to the Ekström Ice Shelf embayment, East Antarctica, over a full glacial cycle (40,000 years). We test the model response to differing ocean bed properties that provide an envelope of potential ocean substrates seawards of today's grounding line. The end member scenarios include a hard, high friction ocean bed and a soft, low friction ocean bed. We find that predicted ice volumes differ by >50% under almost equal forcing. Grounding line positions differ by up to 49 km, show significant hysteresis, and migrate non-steadily in both scenarios with long quiescent phases disrupted by leaps of rapid migration. The simulations quantify evolution of two different ice sheet geometries (namely thick and slow vs. thin and fast), triggered by the variable grounding line migration over the differing ocean beds. Our study extends the timescales of 3D full-Stokes by an order of magnitude to previous studies with the help of parallelisation. The extended time frame for full-Stokes models is a first step towards better understanding other processes such as erosion and sediment redistribution in the ice shelf cavity impacting the entire catchment geometry.

1 Introduction

Shortcomings in the description of ice dynamics remain one of the limitations for projecting the evolution of the Greenland and Antarctic ice sheets (Pachauri et al., 2014). If current sea level rise rates continue unabated, up to 630 million people will be at annual flood risk by 2100 (Kulp and Strauss, 2019), making improved ice sheet model projections important to assess the socioeconomic impact. Due to the high computational costs of full-Stokes (FS) models that solve the complete ice dynamical equations, current long term (>1,000 years) ice sheet simulations rely on simplifications to the ice dynamical equations. This choice is justified because it allows for ensemble modelling and tuning of unknown parameters using observations. There are two drawbacks to this approach. First, it is uncertain whether the transition zone between grounded and floating ice is adequately represented in existing long term simulations (Pattyn and Durand, 2013). Second, the omission of membrane and bridging stress gradients hamper disentangling the relative contributions of basal sliding and ice deformation to the column averaged ice discharge (MacGregor et al., 2016; Bons et al., 2018). The former is one of the main uncertainties for projecting the sea level contribution of contemporary ice sheets (Durand et al., 2009; Pattyn and Durand, 2013). The latter is a bottleneck for the inclusion of basal processes such as erosion and deposition of sediments which critically depend on the magnitude of basal sliding (e.g. Humphrey and Raymond, 1994; Egholm et al., 2011; Herman et al., 2011; Yanites and Ehlers, 2016; Alley et al., 2019) and may govern the formation and decay of ice streams (Spagnolo et al., 2016).

A number of simplified model variants of the full ice flow equations have been successfully applied to sea level rise reconstructions over timescales of >1,000 years (e.g. Golledge et al., 2012; Briggs et al., 2014; Pollard et al., 2016). In order to reproduce past ice sheet geometries paleo ice sheet models rely on observations that constrain the lateral as well as the vertical extent of the ice sheet (e.g. Briggs et al., 2014; Bentley et al., 2014; Golledge et al., 2014). Ice sheet extent is commonly inferred from marine sediment core data or geomorphological data, ice sheet elevation from exposure dating, and changes in ice thickness from ice cores or ice rises (e.g. Bentley et al., 2010; Golledge et al., 2013; Briggs et al., 2014). Fast paleo ice sheet models employ ensemble simulations in which poorly known model parameters are tuned such that they match the constraints. This allows to gauge the uncertainties regarding for example atmospheric and oceanic boundary conditions over glacial cycle timescales (e.g. Golledge et al., 2012; Briggs et al., 2014; Pollard et al., 2016; Albrecht et al., 2020). Each ensemble member simulation is then evaluated against the constraints present at that particular timeslice. To determine the goodness of the fit of individual ensemble members, modelling studies apply statistical methods ranging from weighted scoring schemes (e.g. Briggs et al., 2014; Albrecht et al., 2020) to statistical emulators (Pollard et al., 2016). The rationale behind this tuning is that if the model matches the constraints ~~well, then confidence is high that the model also reproduces ice sheet changes at other times~~poorly, then the model should be rejected. The risk involved is that the matching may overcompensate for the simplified model physics leading to higher uncertainties in future predictions where model constraints are absent. Due to the high computational demands, both, in terms of mesh resolution and the physics required to solve for a freely evolving grounding line (Gillet-Chaulet et al., 2012; Seddik et al., 2012; Favier et al., 2014; Schannwell et al., 2019), FS models up to now have been restricted to individual simulations and simulation lengths of <1,000 years for real world geometries. Therefore, there is a need to extend the applicability of regional FS ice sheet models to timescales longer than 1,000 years so that uncertainties due to physical approximations ~~can be reduced~~in the force balance can be quantified and reduced in the near future.

For glacial cycle simulations with an advance and a retreat phase, the particular challenge arises that the ice sheet advances and retreats over ocean beds where bathymetry and its geological properties are often poorly known. Ensemble modelling studies identified basal properties of ocean beds as a major source of uncertainty in ice dynamic models (e.g. Pollard and DeConto, 2009; Pollard et al., 2016; Whitehouse et al., 2017; Albrecht et al., 2020). This holds especially for drainage basins where such geological constraints are absent. Under contemporary ice sheets, estimating basal friction parameters (e.g. basal friction between the ice sheet and the underlying substrate) is virtually impossible by direct measurements and can only be inferred indirectly on a continental scale by solving an optimisation problem matching today's surface velocities and/or ice thickness (e.g. MacAyeal, 1993; Gillet-Chaulet et al., 2012; Cornford et al., 2015). Furthermore, the inferred basal friction coefficient is often spatially heterogeneous and can vary by up to five orders of magnitude under the present day Antarctic ice sheet (Cornford et al., 2015). To what extent this variability truly reflects variability in geology and/or hydrology, or is falsely introduced by the approximations in the ice dynamical equations or omission of ice anisotropy is unknown.

Here, we present the first regional scale FS simulations investigating the effect of different ocean bed properties under contemporary ice shelves on ice sheet geometry over a glacial cycle. We ~~hereby extend~~ do this by investigating end-member scenarios as opposed to ensemble modelling. This means, we specify either very soft and slippery or very hard and sticky conditions under present-day ice shelves. The goal of the paper is hence two-fold. First, we present methodological advances by extending the feasibility of regional FS ice sheet simulations by an order of magnitude using the open source code Elmer/Ice (Gagliardini et al., 2013). We do this with a highly parallelised numerical scheme allowing to maintain a high mesh resolution (∼1 km) and a freely evolving grounding line over glacial/interglacial timescales. ~~Our simulations focus on~~ Second, we present new scientific insights regarding the effect of different ocean bed properties seawards of today's grounding line and ~~to quantify their~~ quantify its impact on the evolution of the entire catchment. This is done for the Ekström Ice Shelf catchment, Dronning Maud Land, East Antarctica (Fig. 1).

2 The Ekström catchment, Dronning Maud Land, East Antarctica

We have chosen the Ekström catchment for our study because it hosts the German overwinter station Neumayer III and is therefore particularly well constrained by geophysical and climatological observations and boundary datasets. Uncertainties in the contemporary ice sheet geometry are small because of previous dense airborne radar surveys (Fretwell et al., 2013). Unlike many other ice shelves, the bathymetry in this area is known to an unprecendented extent from seismic reflection surveying (Smith et al., 2019). This has been complemented with bathymetry modelling via gravity inversion from airborne gravity data to cover the whole cavity (Eisermann et al., 2020). In comparison to the Bedmap2 dataset (Fretwell et al., 2013), the updated cavity is up to 1,000 m deeper. For our simulations, this difference is only relevant up to the point of farthest grounding line advance. We use the Eastern Dronning Maud Land (EDML) ice core (Graf et al., 2002) as proxy for past temperature variations in the region. The location of the EDML ice core is about 700 km to the south-east of the modelling domain on the Antarctic plateau. The Ekström catchment contains also two ice rises (Schannwell et al., 2019; Drews et al., 2013) with independent ice flow centres from the main ice sheet. Ice rises archive the regional ice sheet history in their internal stratigraphy. Therefore, their stability or lack thereof provides indications about past ice flow changes of the area. Furthermore, while geological constraints about the retreat history since the LGM are still uncertain, there is evidence in this area from multiple geophysical observations (Kristoffersen et al., 2014) and geological signatures (Eisermann et al., 2020) about contrasting ocean bed properties. There is also growing evidence that the catchment is close to steady state (e.g. Drews et al., 2013; Schannwell et al., 2019) which we consider beneficial for our model initialisation. While much recent research has focused on the fast flowing outlet glaciers of Antarctica, we stress the importance of also studying catchments characterised by slower moving ice (<300 m/yr), as they occupy ∼90% of the contemporary Antarctic grounding line and account for 30% of the total ice discharge (Bindschadler et al., 2011; Rignot et al., 2011). The results we obtain for the Ekström Ice Shelf catchment could therefore be relevant for many other catchments around Antarctica and hence the total budget.

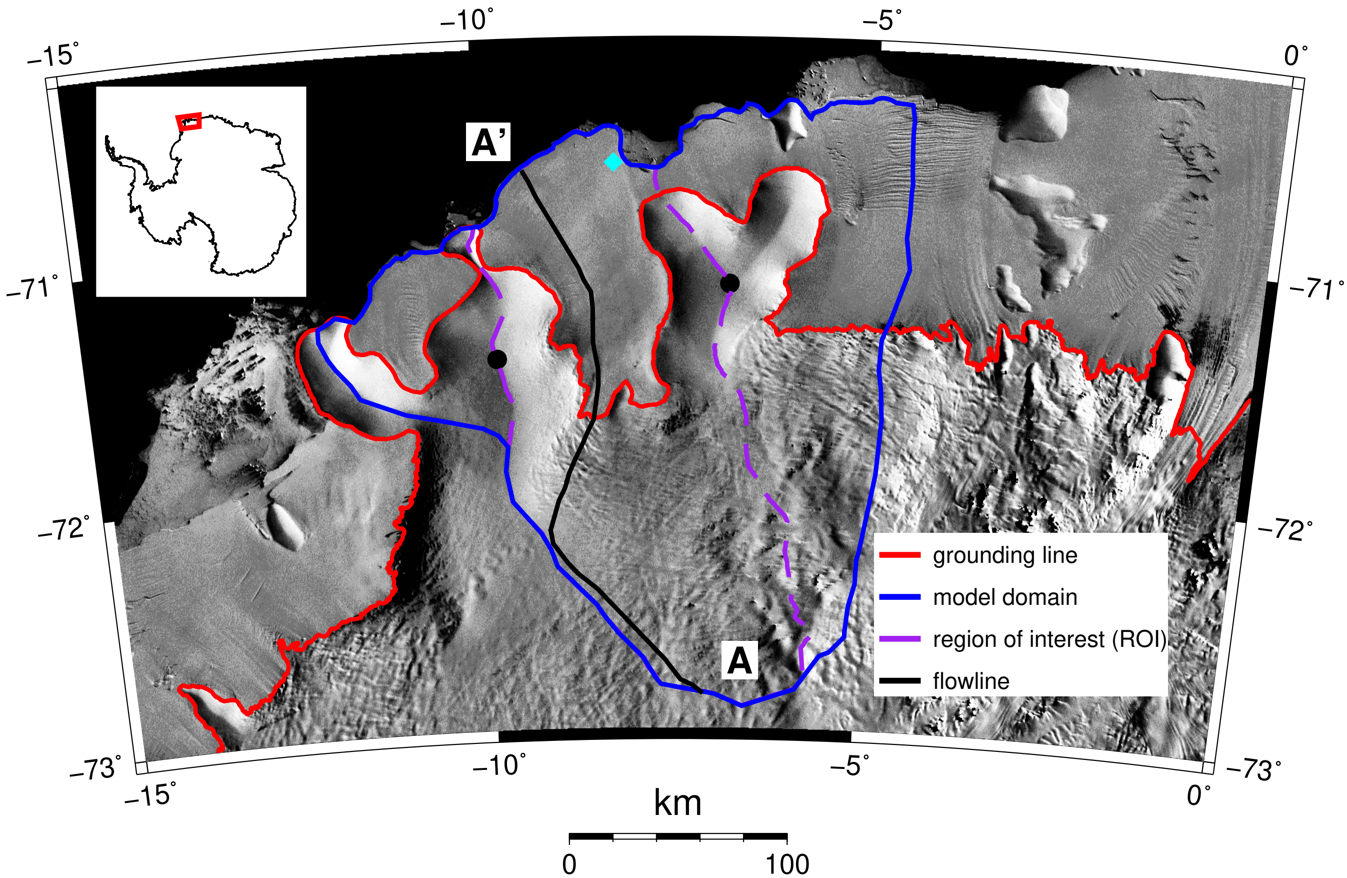

Figure 1. Overview of the Ekström Ice Shelf catchment with present day grounding line (Bindschadler et al., 2011) and model domain. Cyan square shows location of Neumayer Station III. Filled black circles indicate location of ice rises. Flowline (A-A') is shown in Fig. 10. Background is the MODIS Mosaic of Antarctica (Scambos et al., 2007).

3 Model description

3.1 Ice flow equations

Ice flow is dominated by viscous forces which permits the dropping of the inertia and acceleration terms in the linear momentum equations. The Elmer/Ice ice sheet model (Gagliardini et al., 2013) solves the complete 3D equation for ice deformation. This results in the Stokes equations described by

$$\nabla \cdot \boldsymbol{\sigma} = -\rho_i \boldsymbol{g}. \tag{1}$$

Here, $\boldsymbol{\sigma} = \boldsymbol{\tau} - p\boldsymbol{I}$ is the Cauchy stress tensor, $\boldsymbol{\tau}$ is the deviatoric stress tensor, $p = -tr(\boldsymbol{\sigma})/3$ is the isotropic pressure, $\boldsymbol{I}$ the identity tensor, ρ_i the ice density, and $\boldsymbol{g}$ is the gravitational vector. Ice flow is assumed to be incompressible which simplifies mass conservation to

$$\nabla \cdot \boldsymbol{u} = 0, \tag{2}$$

with $\boldsymbol{u}$ being the ice velocity vector. Here we model ice as an isotropic material. Its rheology is given by Glen's flow law which relates the deviatoric stress tensor $\boldsymbol{\tau}$ with the strain rate tensor $\dot{\boldsymbol{\epsilon}}$:

$$\boldsymbol{\tau} = 2\eta\dot{\boldsymbol{\epsilon}}, \tag{3}$$

where the effective viscosity η can be expressed as

$$\eta = \frac{1}{2}B\dot{\epsilon_e}^{\frac{(1-n)}{n}}. \tag{4}$$

In this equation B is a viscosity parameter that depends on ice temperature relative to the pressure melting point computed through an Arrhenius law, n is Glen's flow law parameter (n=3), and the effective strain rate is defined as $\dot{\epsilon_e}^2 = tr(\dot{\boldsymbol{\epsilon}}^{\boldsymbol{2}})/2$.

3.2 Ice temperature

The ice temperature is determined through the heat transfer equation (e.g. Gagliardini et al., 2013) which reads

$$\rho_i c_v \left(\frac{\partial T}{\partial t} + \boldsymbol{u} \cdot \nabla T\right) = \nabla \cdot (\kappa \nabla T) + \dot{\boldsymbol{\epsilon}} : \boldsymbol{\sigma}, \tag{5}$$

where c_v and κ are the specific heat of ice and the heat conductivity, respectively. The **:** operator represents the colon product between two tensors. This last term of the equation represents strain heating. The ice temperature T is bounded by the pressure melting point T_m, so that $T \leq T_m$.

3.3 Boundary conditions

3.3.1 Ice temperature

Our parameterisation of surface temperature changes follows Ritz et al. (2001). We parameterise relative surface temperature changes to present day as a function of relative surface elevation change with respect to present day elevations and a spatially uniform surface temperature variation that is derived from the nearby EDML ice core (Graf et al., 2002). The surface temperature is then given by (Ritz et al., 2001, eq. 11):

$$T_a = T_{a0} - \gamma_a(z_{s0} - z_s) + \Delta T_{clim}. \tag{6}$$

Here, T_a and T_{a0} are the surface temperatures at the current timestep and present day. The present day temperature distribution is taken from Comiso (2000). z_s and z_{s0} are the surface elevations at the current timestep and present day, and ΔT_{clim} is the climatic forcing derived from the EDML ice core. As in Ritz et al. (2001), we apply a spatially constant lapse rate (γ_a) of

0.00914 K/m (Table 1).

At the grounded base of the ice sheet, where the ice is contact with the subglacial topography, we prescribe the geothermal heat flux (Martos et al., 2017). This heat flux is time invariant. Ice temperature is set to the local pressure melting point for the boundary condition underneath the floating ice shelves.

3.3.2 Surface mass balance (SMB) and basal mass balance (BMB)

A kinematic boundary condition determines the evolution of upper and lower surfaces z_j:

$$\frac{\partial z_j}{\partial t} + u_x \frac{\partial z_j}{\partial x} + u_y \frac{\partial z_j}{\partial y} = u_z + \dot{a}_j, \tag{7}$$

where $\dot{a}_j$ is the accumulation/ablation term and $j = (b, s)$, with s being the upper surface and b being the lower surface (base) of the ice sheet.

For the surface mass balance (SMB) parameterisation, we closely follow Ritz et al. (2001) again. We assume that no melt occurs in all our simulations. This is justified because SMB models simulate little melt at present day conditions (Lenaerts et al., 2014) and these are the warmest years in our simulations. As for the surface temperature, our SMB parameterisation uses a present day distribution of the SMB (Lenaerts et al., 2014) as input. Variations of the SMB over time are then proportional to the exponential of the surface temperature variation (Ritz et al. (2001), eq. 12):

$$\dot{a}_s(T_a) = a_{s0}(T_{a0}) exp(\Delta a(T_a - T_{a0})), \tag{8}$$

where a_{s0} is the present day SMB, a_s is the SMB at the current timestep, and the parameter $\Delta a = 0.07$ K^{-1}. This means that for a surface temperature drop of 10 K, the SMB is reduced by 50% (Ritz et al., 2001).

Sub shelf melting underneath the floating ice shelves is based on the difference between the local freezing point of water under the ice shelves and the ocean temperature near the continental shelf break (Beckmann and Goosse, 2003). The freezing temperature (T_f) is calculated through:

$$T_f = 0.0939 - 0.057 S_o + 7.64 \times 10^{-4} z_b, \tag{9}$$

where z_b is the base of the ice shelf and S_o is the ocean salinity (Table 1). The basal melt rates ($\dot{a}_b$) are then computed by

$$\dot{a}_b = \frac{\rho_w c_{p_o} \gamma_T F_{melt} (T_O - T_f)^2}{L \rho_i}. \tag{10}$$

In this equation, ρ_w is the density of water, c_{p_o} is the specific capacity of the ocean mixed layer, γ_T is the thermal exchange velocity, L is the latent heat capacity of ice, F_{melt} is a tuning parameter to match present day melt rates, and T_O is the ocean temperature (Table 1). The ocean temperature is initially set to -0.52°C (Beckmann and Goosse, 2003). F_{melt} is chosen such that present day basal melt rates do not exceed ∼1.1 m/yr. This is in accordance with melt rates derived from satellite observations and mass conservation (Neckel et al., 2012). Applied variations of the ocean temperature are a damped (∼40%) and delayed (∼3,000 years) version of the climatic forcing for surface temperature ΔT_{clim} (Bintanja et al., 2005).

3.3.3 Basal sliding and sea level

Where the ice is in contact with the subglacial topography a linear Weertman-type sliding law of the form

$$\boldsymbol{\tau_b} = C|\boldsymbol{u_b}|^{m-1}\boldsymbol{u_b}, \tag{11}$$

is employed. Here $\boldsymbol{\tau_b}$ is the basal traction, m is the basal friction exponent which is set to 1 in all simulations, and C is the basal friction coefficient. A linear viscous sliding relation (m=1) was chosen~~to guarantee consistency between model intialisation and forcing simulation. The consequences of this choice~~ . Alternative and physically more realistic sliding relations exist (e.g. Joughin et al., 2019) and the consequences of our choice of using a linear sliding relation on the results are discussed below (see section 5.5). For the present day grounded ice sheet, C is inferred by solving an inverse problem (see section 3.4), and for the present day ocean beds a uniform basal friction coefficient of $10^{\cancel{-1}\,-5}$ MPa m^{-1} yr and $10^{\cancel{-5}\,-1}$ MPa m^{-1} yr is prescribed for the soft (sediment based) bed and hard (crystalline rock based) bed simulations. Underneath the floating part of the domain basal traction is zero ($\boldsymbol{\tau_b} = 0$), but hydrostatic sea pressure is prescribed. We initialise the present day sea level to zero and apply sea level variations according to Lambeck et al. (2014).

Table 1. Numerical values of the parameters adopted for the simulations

Parameter	Symbol	Value	Unit
Ice density	ρ_i	917	kg m^{-3}
Ocean density	ρ_w	1028	kg m^{-3}
Glen's exponent	n	3	
Gravity	g	9.81	m s^{-2}
Atmospheric lapse rate	γ	0.00914	K m^{-1}
Tuning parameter SMB	Δa	0.07	K^{-1}
Ocean salinity	S_0	35.0	PSU
Heat capacity	c_{p_o}	3974	J kg^{-1} °C^{-1}
Latent heat of fusion	L	3.35×10^{-4}	J kg^{-1}
Tuning parameter BMB	F_{melt}	0.383×10^{-4}	
Thermal exchange velocity	γ_T	1×10^{-5}	m s^{-1}

3.4 Model initialisation

The model is initialised to the present day geometry using the commonly applied snapshot initialisation in which the basal traction coefficient C is inferred under the grounded ice sheet by matching observed surface velocities with modelled surface velocities. We take advantage of the quasi steady state of the catchment and use same optimisation parameters as in Schannwell et al. (2019). Similar to Zhao et al. (2018), we employ a two step initialisation scheme. In the first iteration, the optimisation problem is solved with an isothermal ice sheet with ice temperature set to –10°C. The resulting velocity field is then used to solve the steady state temperature equation before the optimisation problem is solved again with the new temperature field. This type of temperature initialisation approach provides similar results to a computationally more expensive temperature spin up over several glacial cycles (Rückamp et al., 2018), as long as the system is close to steady state.

3.5 Mesh generation and refinement

We initially create a 2D isotropic mesh with a nominal mesh resolution of $\sim$6 km everywhere in the domain. To ensure that we simulate grounding line dynamics at the required detail, we use the meshing software MMG (http://www.mmgtools.org/, last access: 28 February 2020) to locally refine the mesh down to $\sim$1 km in the region of present day Ekstöm Ice Shelf (Figure 2) with areas away from the region of interest remaining at $\sim$6 km resolution. The mesh is then vertically extruded ~~, consisting of~~ using 10 layers and the horizontal mesh size is kept constant throughout the simulations. The 3D mesh consists of $\sim$200,000 nodes and therefore $\sim$800,000 degrees of freedom. We are using stabilised P1P1 elements and an algorithm that deduces a mass-conserving nodal surface to avoid artificial mass loss (Gagliardini et al., 2013).

3.6 Block preconditioned ParStokes solver

Because of the non-Newtonian rheology of ice and the dependence of viscosity on strain rates, the resulting Stokes equations are non-linear and have to be solved iteratively. In three dimensions the arising systems of linear equations become large (10^6–10^7 degrees of freedom) at high mesh resolution. Standard iterative methods (Krylov subspace methods) in conjunction with algebraic preconditioners (e.g. Incomplete Lower Upper (ILU) decomposition) do often not converge for real world geometries in glaciology. High aspect ratios of the finite elements and spatial viscosity variations of several orders of magnitudes, strongly affect accuracy and stability of the numerical solution (Malinen et al., 2013). This means that most glaciology applications with Elmer/Ice revert to using a direct method for solving the Stokes equations. While robust, direct solvers ~~do not take advantage of the sparse structure of the matrix and~~ require large amounts of memory. In three dimensions their memory requirements increase with the square of the number of unknowns. Therefore, we use a stable parallel iterative solver (ParStokes) in our simulations that is implemented in Elmer/Ice, but has so far been rarely used. ParStokes is based on block preconditioning (Malinen et al., 2013) that improves the solvability of the underlying saddle point problem through clustering of Eigenvalues. As we will show below the Krylov subspace methods now converge better and lead to improved scaling with more Computer Processing Units (CPUs).

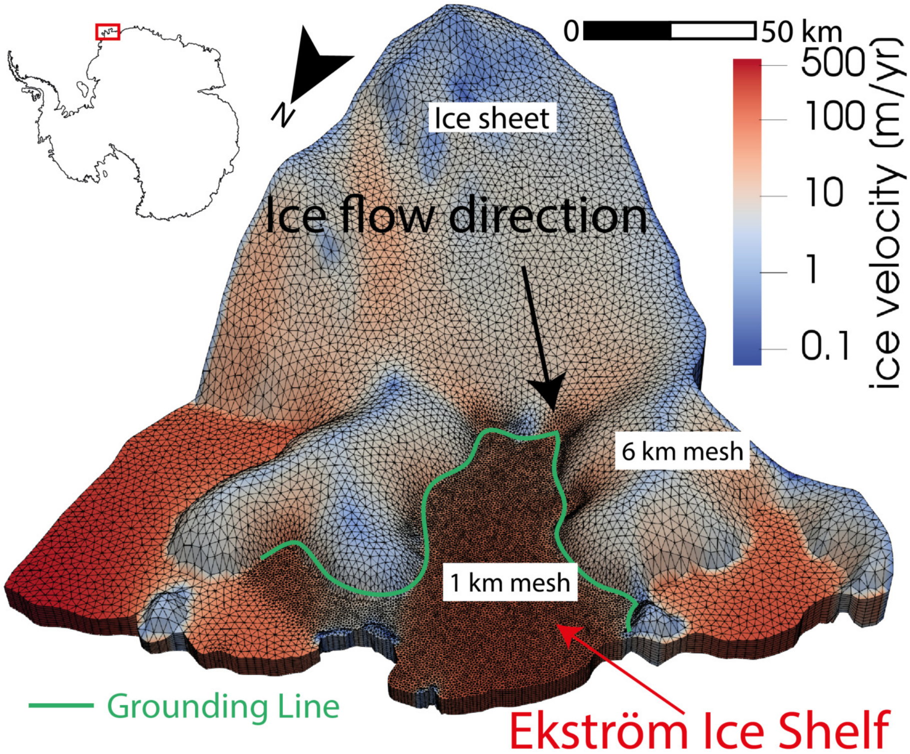

Figure 2. Model domain of Elmer/Ice in 3D including numerical mesh of Ekström Ice Shelf catchment, East Antarctica, with ice velocity in the background

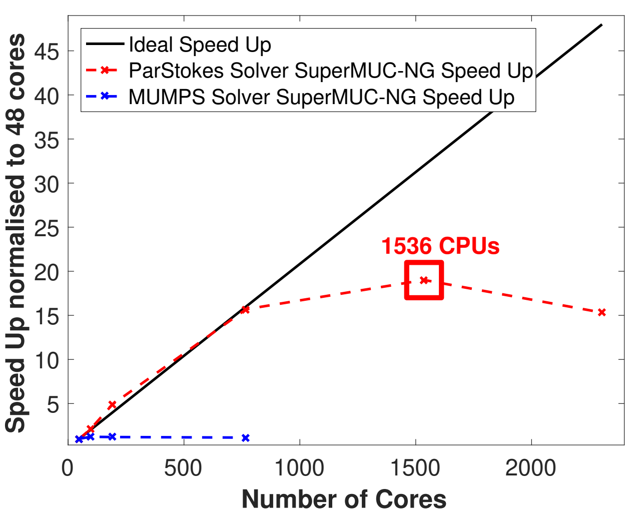

Figure 3. Scaling behaviour of iterative solver (ParStokes) and direct Solver (MUMPS) for Elmer/Ice on the SuperMUC-NG supercomputer. Red square denotes number of CPUs selected for this study.

3.7 Experimental design

We demonstrate a FS simulation of ice sheet growth and decay over 40,000 years. During the first 20,000 years the atmospheric and oceanic forcing simulates the transition from an interglacial to a glacial (henceforth called the advance phase). We then symmetrically reverse the climate forcing to simulate deglaciation (henceforth called the retreat phase). The symmetrical reversal of the model forcing enables investigation of hysteresis effects. The interglacial starting conditions are chosen with present day properties and characteristics, so that the best possible basal friction coefficient beneath the grounded ice sheet can be found using today's ice sheet geometry and surface velocities (Schannwell et al., 2019). The glacial conditions are chosen to resemble the Last Glacial Maximum for which we have good constraints for atmospheric forcing from the nearby EDML ice core. We consider two end member basal property scenarios by prescribing either soft ocean bed conditions (mimicking sediment deposits) or hard ocean bed conditions (mimicking crystalline rock) ~~for~~ under all present day ~~ocean cavities~~ ice shelves in the modelling domain. The tested scenarios of basal traction coefficients encompasses what other ice sheet models have inferred (e.g. Cornford et al., 2015) for the grounded portion underneath the present day Antarctic ice sheet (basal traction coefficient ranging from $10^{\cancel{-1}-5}$ MPa m^{-1} yr for sediments to $10^{\cancel{-5}-1}$ MPa m^{-1} yr for crystalline bedrock). Those end member values do not reflect a true range of sliding coefficients for a given sliding law, but were derived as tuning parameters. Hence they also account for some uncertainties in model parameters, forcings, and physics of the applied ice sheet model. That is why we consider those values as end members and regard simulated differences in ice volume and grounding line position as the maximum envelope of uncertainties resulting from different ocean bed properties. We perform the simulations with a) the standard Elmer/Ice setup using the Multifrontal Massively Parallel Sparse (MUMPS) direct solver for ice velocities; and b) using a stable iterative solver for ice velocities (see section 3.6), resulting in a total of four simulations. We carried out the ~~simulation on three~~ simulations on two different high performance computing systems: the ZDV cluster ~~, the now decommissioned SuperMUC system,~~ and the SuperMUC-NG system.

4 Results

The results can be divided into methodological advances and new scientific insights. In the following, we first present the technical improvements of the presented Elmer/Ice model setup in comparison to the "classic" setup employed in previous studies (e.g. Schannwell et al., 2019). This is followed by the analysis of the performed model simulations in terms of ice flow behaviour and an analysis of the role of the subglacial strata characteristics for advance and retreat dynamics.

4.1 Comparison between direct Stokes solver (MUMPS) and ParStokes

The ParStokes solver allows for a much better scaling of the required computation time with increasing numbers of CPUs (Figure 3). While there is no speed up for the "classic" solver setup using the direct solver MUMPS, there is a linear speedup for the ParStokes solver up to $\sim$700 CPUs before the rate of speedup tapers off and vanishes for more than 1536 CPUs. This much better scaling behaviour results in a total compute time for the iterative solver on the SuperMUC-NG system that is faster between a factor 3–6 in comparison to the MUMPS solver setup on the ZDV system. For our simulations, this means that the 40,000 year simulation now takes 23 days instead of 141 days for the hard bed case, and 27 days instead of 94 days for the soft bed case (Figure 4). ~~In comparison, on the now decommissioned SuperMUC system, total compute time savings were only 20 days in comparison to the MUMPS solver setup. The reason for this were the long queuing intervals in between simulations, leading to an additional~~ This speed up is in part due to using more CPUs in the ParStokes simulations. When comparing absolute runtime of the scaling simulations, ParStokes provides faster computations for $>$~~80 days of waiting for simulations to run in comparison to the other systems. This is a direct consequence of the system being in the process of shutting down and hence only running at 50% capacity~~168 CPUs. This means the minimum requirement for faster simulations with ParStokes is a supercomputer with more than 168 CPUs. The exact CPU number may however very well vary from system to system depending on the available hardware.

We use predicted grounding line position and ice thickness as metrics to compare the "classic" solver setup using MUMPS with the new solver ParStokes. We note however that we do not expect a perfect match between the two solver setups due to small differences in the finite element formulation (e.g. stabilisation method). ~~For~~ When using the MUMPS solver, the stabilised method is used, while for the ParStokes solver we use bubble stabilisation (Gagliardini et al., 2013). This results in slightly different systems that need to be solved. However, for both simulations, there is good agreement in terms of grounding line position over time, with differences in grounded area never exceeding 5% (Figure 5). Because the soft bed simulation exhibits smaller magnitude grounding line motion over the simulation, agreement between the two solver setups is better, with differences well below 1% for almost the entire simulation length. In the hard bed simulation, where larger magnitudes of grounding line motion are predicted, the ParStokes solver's grounding line is not as far advanced as the MUMPS solver grounding line (Figure 6). Moreover, at times of rapid grounding line motion, the response of the grounding line in the ParStokes solver is delayed by up to $\sim$3,500 years. This leads to differences in transient grounding line positions ($<$5%). However grounding line positions for steady state situation differ negligibly ($<$1.5% difference). The predicted ice thickness differences are larger, particularly for the hard bed run, where ice thickness change is larger overall. Locally these differences can be as large as $\sim$460 m ($<$25% of the ice thickness) in transient scenarios. They are most pronounced in periods of delayed grounding line response. Once a stable grounding line position has been reached, thickness differences are notably smaller (Figure 6, 7). Overall, the ParStokes solver provides comparable results to the MUMPS solver, but is much superior in terms of the required computation time. Therefore, the remainder of the results section will be based on the ParStokes solver simulations.

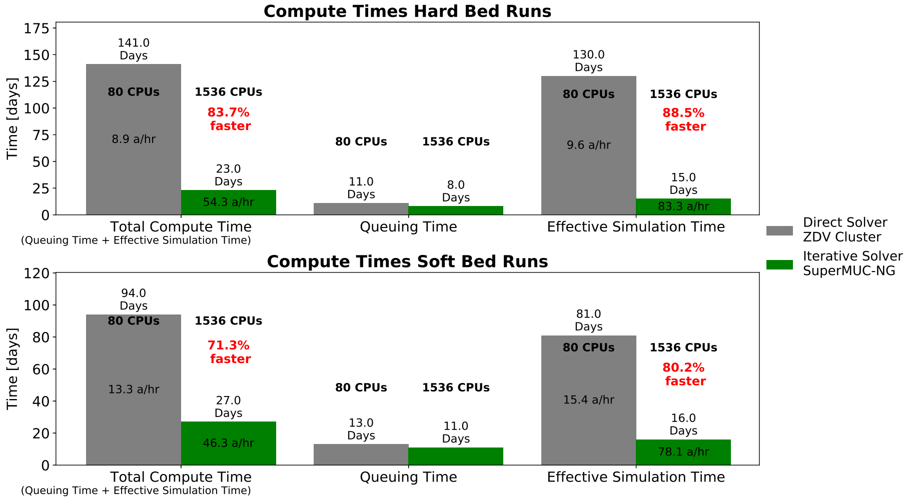

Figure 4. Speed up of iterative Solver (ParStokes, green and blue bars) in comparison to direct solver (MUMPS, gray bars) for the hard bed cavity (upper panel) and soft bed cavity (lower panel) simulations. Simulations were performed on three different high performance computing systems (ZDV ~~, SuperMUC,~~ and SuperMUC-NG).

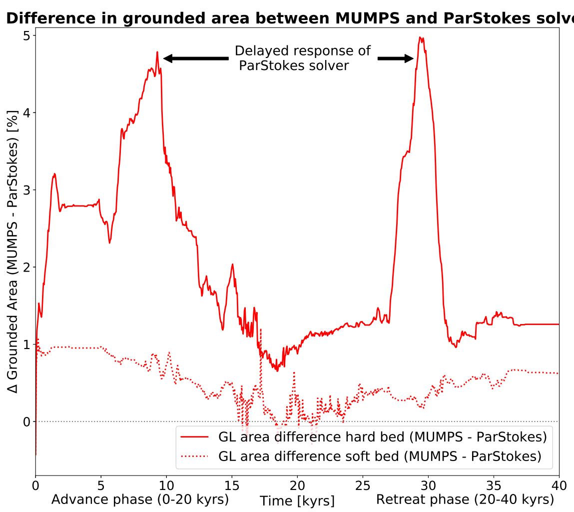

Figure 5. Differences in grounded area between the classic MUMPS and ParStokes solver setup for the soft bed and hard bed simulations.

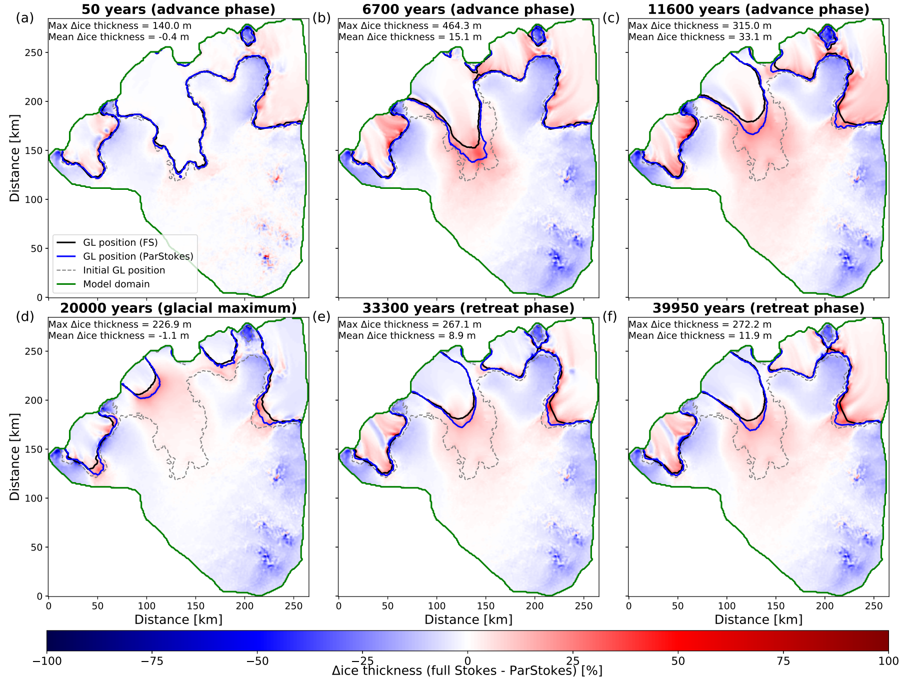

Figure 6. Differences in grounding line position and ice thickness between the classic MUMPS and ParStokes solver setup for the hard bed simulation at specific time slices.

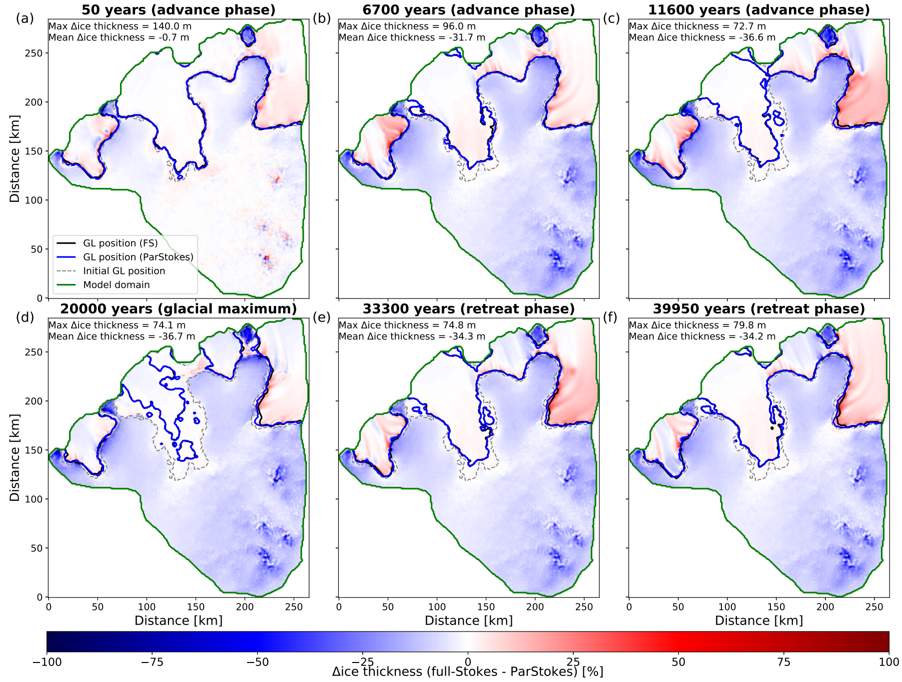

Figure 7. Differences in grounding line position and ice thickness between the classic MUMPS and ParStokes solver setup for the soft bed simulation at specific time slices.

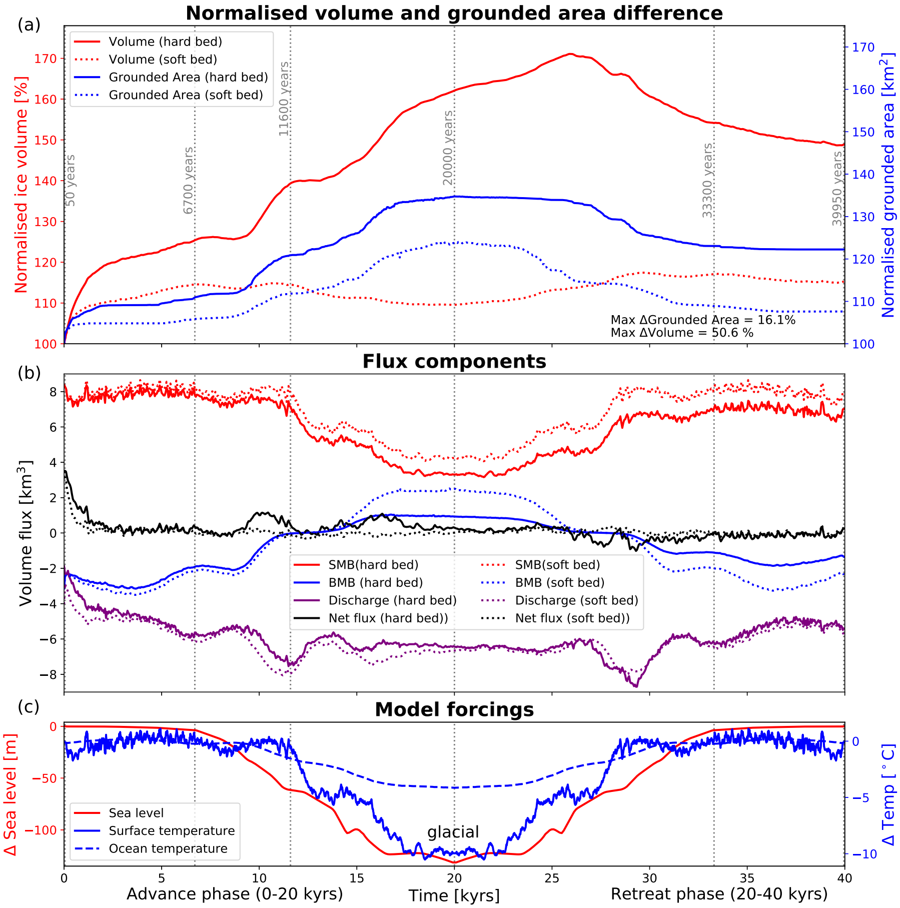

Figure 8. Ice sheet evolution and model forcing for soft and hard bed simulations. (a) shows volume and grounded area evolution normalsied to present day. (b) shows corresponding mass balance fluxes, and (c) shows most important model forcings. Vertical grey stippled lines show time slices shown in Figs. 6, 7, 9, and 10.

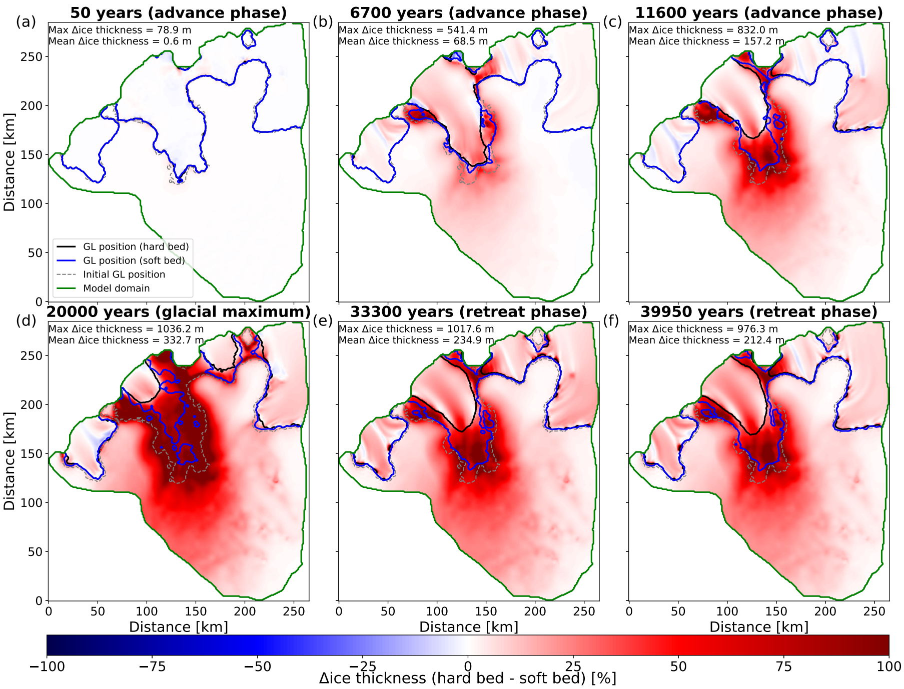

Figure 9. Differences in plane view of ice thickness and grounding line positions between the hard and soft bed simulations at selected time slices. (a-d) show differences in the advance phase and (e,f) show differences in the retreat phase.

4.2 Influence of bed hardness on ice sheet growth and decay

As expected, the hard and soft bed simulations result in different ice sheet geometries. Quantitatively, both scenarios differ significantly in transient and steady state volumes (Fig. 8), fluxes, and grounding line positions (Figs. 9 and 10). The simulated hard bed ice sheet is in many areas more than twice as thick as the soft bed ice sheet, with maximum ice thickness differences between hard and soft bed reaching 1,036 m or 120% (Fig. 10). In more detail, the differences between these simulations are as follows. First, the hard bed ice sheet results in a thick, slow, and large volume ice sheet after 20,000 years at glacial conditions. During the advance phase, volume increases occur step-wise with three distinct periods of volume increases (Fig. 8). These periods of volume increase in the region of interest are short (<2,000 years) and are interrupted by longer periods of little ice volume change. At the glacial maximum, the volume increase in comparison to the interglacial is ∼60%. During the first ∼8,000 years in the retreat phase, the hard bed simulation continues to gain volume albeit at a slow rate. ~~In the following~~ Following this period of volume gain, the ice sheet starts to loose volume. However, the rate of volume loss is small, such that after a full glacial cycle, the total ice volume is still ∼47% more of what is was at the beginning of the simulation.

Second, unlike the hard bed simulations, the soft bed simulation leads to a thin, fast, and small volume ice sheet at glacial conditions. During the advance phase, this simulation does not show a step-wise volume gain pattern. In fact, apart from an initial volume gain in the first 1,000 years of the advance phase (∼10%), there is very little volume change. This leads to a volume increase of merely ∼8% at the glacial maximum. The trend of little volume variations continues during the retreat phase, where in the first 10,000 years a volume increase of $\sim$8% occurs, before the volume remains approximately constant for the remainder of the retreat phase.

The entirely different ice sheet geometries for soft and hard bed simulations have consequences for the two ice rises present in the catchment (Fig.1). While both ice rises and their divide positions are very little affected by the soft bed simulations, they are partly overrun in the hard bed simulation such that their local ice flow centre vanishes (SI video 1).

4.3 Grounding line and ice sheet stability

~~Stable~~ Steady grounding line positions for both simulations are associated with periods of ice sheet stability (Fig. 8). There are three distinct periods of grounding line stability in the advance phase and one period of grounding line stability in the retreat phase. All of these four periods are longer than 3,000 years. Periods of grounding line advance in comparison are characterized by short leaps taking no longer than 1,000–2,000 years (Fig. 8). During the advance phase, differences in grounding line positions between the hard bed and soft bed simulations gradually increase from 7 km after $\sim$1,500 years to over 37 km after 11,600 years, and finally to its maximum difference of 49 km at the glacial maximum (Fig. 10). Grounding line advance for the hard bed is more than twice as far ($\sim$110% larger) than its soft bed counterpart in the advance phase. In the retreat phase, the soft bed simulation shows higher grounding line fidelity compared to the hard bed simulation. The soft bed starts to exhibit grounding line retreat after $\sim$4,000 years into the retreat phase, whereas the hard bed does not show grounding line retreat for $\sim$8,000 years into the retreat phase.

4.4 Hysteresis of ice sheet simulations

Next we compare the ice sheet geometries during a full glacial cycle in which atmospheric and oceanic forcing are essentially symmetrically reversed. There is a significant grounding line advance in the first $\sim$300 years in both simulations. In the following, hysteresis is analysed with respect to this position, rather than the start of the simulation. Only the hard bed simulation shows significant hysteresis behaviour, while the soft bed simulation has negligible hysteresis (Fig. 11). For the hard bed simulation, the grounding line after a full glacial cycle is $\sim$38 km further downstream of its initial position. This means that during the retreat phase, the grounding line retreats only $\sim$48% in comparison to the simulated grounding line advance during the retreat phase of the hard bed simulation.

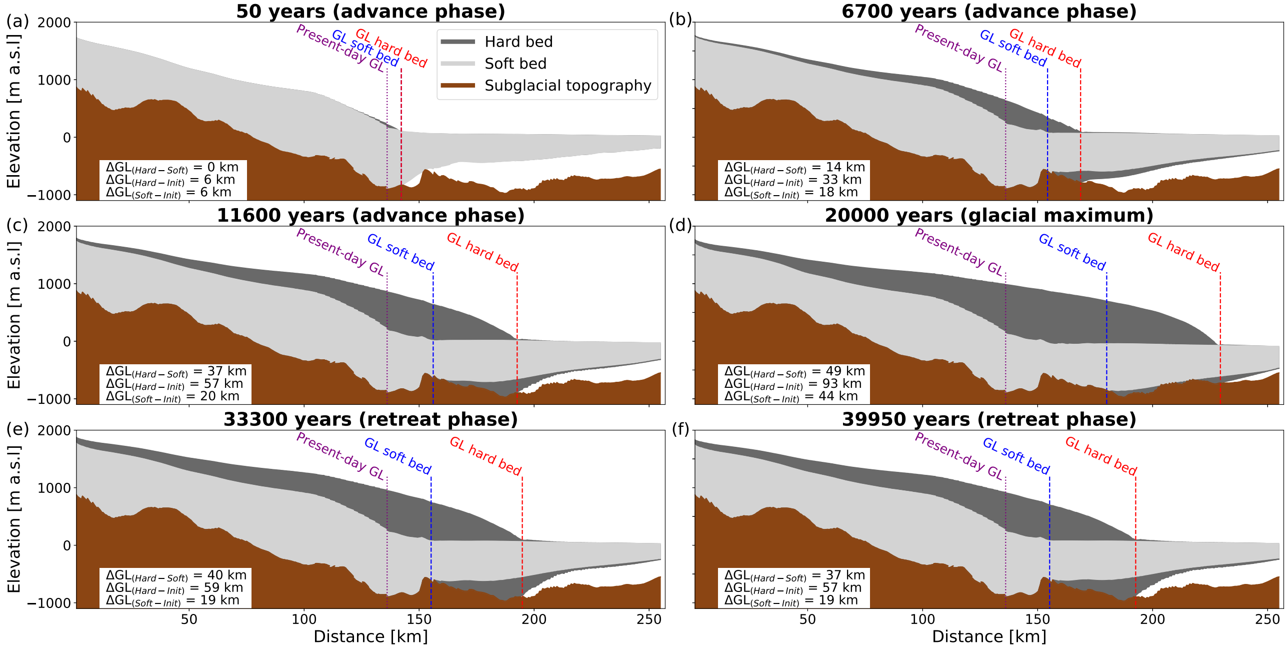

Figure 10. Difference in ice sheet geometry and grounding line position along a flowline (A-A' in Fig. 1) for the soft and hard bed simulations. (a-d) show differences in the advance phase and (e,f) show differences in the retreat phase.

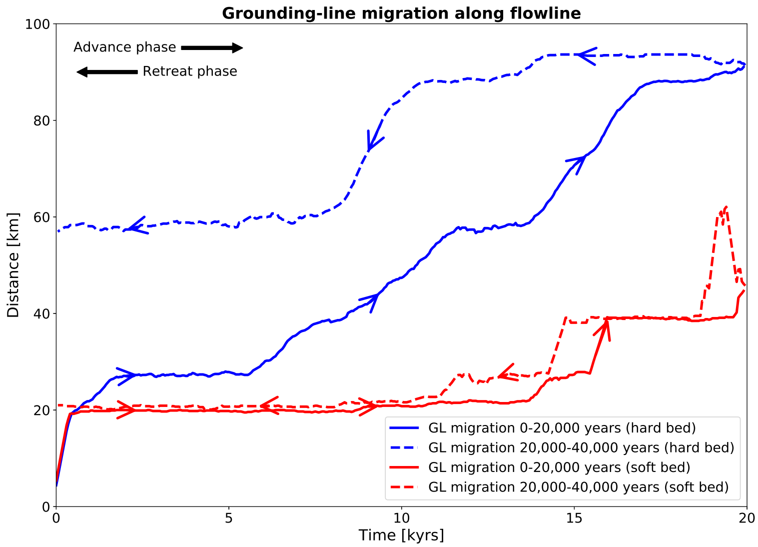

Figure 11. Grounding line migration along a flowline (A-A' in Fig. 1) for the soft and hard bed simulations for the advanced (solid lines) and retreat phase (dashed lines).

5 Discussion

5.1 Extending the feasibility timescales of full-Stokes models

The inclusion of the iterative ParStokes solver results in a speed up by a factor 3-6 compared to the direct solver. While grounding line positions agree well between the two solver setups, during periods of rapid grounding line migration, positions can differ by up to $\sim$5%. We note, however, that we do not expect a perfect match between the two solver setups due to small differences in the finite element formulation (e.g. stabilisation method). Therefore, differences in grounding line positions were expected between the solver setup, but they turn out to be small. The new setup ~~extends the time range of~~ allows 3D full-Stokes ice sheet ~~models~~ simulations on the regional scale ~~from $\leq$1,000 years previously to~~ over 40,000 years now ~~. The high mesh resolution required to adequately capture grounding line migration (Pattyn et al., 2013) is hereby maintained~~in under a month's time. We hereby maintain a mesh resolution ($\sim$1 km) that is finer than in most other paleo ice sheet simulations (Pollard and DeConto, 2009; Golledge et al., 2014; Albrecht et al., 2020) albeit at a regional scale. However, while the time range is now significantly extended, our modelling approach only brackets the effect of ocean bed properties. As detailed below (section 5.5), many other factors influencing ice sheet evolution, such as the applied BMB and SMB parameterisations, and basal sliding relation, remain poorly constrained or are even excluded (e.g. glacial isostatic adjustment). Ensemble modelling (e.g. Golledge et al., 2012; Briggs et al., 2014; Golledge et al., 2014; Pollard et al., 2016; Albrecht et al., 2020) using simplified ice physics is better suited for this, because these models can more easily include other important model sub systems (e.g. basal hydrology, basal sliding) and evaluate their respective uncertainties.

Our efforts aim towards including higher order ice physics into paleo ice sheet simulations. The advantages of our FS simulations are as follows. By retaining all terms in the force balance, we have a solid physical representation of internal deformation and grounding line dynamics over glacial timescales. This permits an improved quantification of the relative contributions from basal sliding and ice deformation to the column averaged ice discharge, opening the door for a better understanding of basal processes such as erosion and deposition of sediments and the formation of ice streams. We are also able to quantify the effect of ocean bed properties onto the grounded ice sheet as the backstress provided by the contrasting ocean bed properties is correctly transmitted upstream by our FS model. Grounding line migration also needs to be interpreted in relation to observed bedforms. For example, the bedrock bump at 150 km in Figure 10 is interpreted as a potential overdeepening, carved out by the confluence of two paleo ice streams (Smith et al., 2019). Our study presents the numerical framework to test hypotheses such as this. Even though we are still not able to constrain our model with paleo observations due to the computation requirements, our study provides an important first step towards it. In addition, computing the full 3D ice velocity field from the linear momentum equations may help to include thus far unused paleo data as constrains. For example, radar isochrones for floating ice shelves could be incorporated more easily into the model tuning, because the FS approach does not apply a vertical average in these areas ~~. Ensemble modelling~~ unlike ice models using a simplified force balance. We believe that ensemble modelling using simpler ice physics models and our approach ~~are in that regard complimentary. Both~~ of employing a complex ice-physics model and investigating end-member scenarios can both provide different new insights. Hence, both approaches should be pursued ~~as improvements to either are mutually beneficial for both~~in future. This also holds for shallow ice approximation-FS hybrid approaches (Ahlkrona et al., 2016) which can build on the results shown here.

5.2 Influence of bed hardness on ice sheet growth and decay

The completely different ice sheet geometries for the hard and soft bed simulations are a consequence of the different levels of basal friction provided by the hard and soft bed, respectively. The predicted differences between the hard bed and soft bed simulations underline the high significance of a proper choice of basal properties used for ocean beds. The higher basal friction in the hard bed case leads to elevated back stress and corresponding dynamical thickening of the inland ice sheet far upstream of the grounding line. Although the SMB and BMB forcings equally depend on the ice sheet geometry through the applied parameterisations, these effects are small compared to the ice dynamically induced thickening (Fig. 9). This clearly shows that in the absence of other forcing mechanisms, ocean bed properties exert an important control on ice sheet growth and decay. The importance of ocean bed properties on ice sheet evolution is long known (e.g. Pollard and DeConto, 2009; Whitehouse et al., 2012; Pollard et al., 2016; Whitehouse et al., 2017; Albrecht et al., 2020). Here we quantify upper and lower bounds of this effect for the first time on a regional scale with a FS model. Our results indicate that spatial changes of basal friction coefficients in the cavities are likely very important for ice sheet growth and decay behaviour. This is relevant for the Ekström Ice Shelf embayment and probably most of Dronning Maud Land, as evidence from geophysical data show that the ocean bed of the Ekström cavity consists at least partly of crystalline bedrock (Kristoffersen et al., 2014; Smith et al., 2019). This feature is more than 1000 km long. A new compilation and interpretation of airborne geophysics data by Eisermann et al. (2020) shows that the northern edge of a strong magnetic anomaly coincides with the location of the outcrop of the Explora Volcanic Wedge (Smith et al., 2019), where subglacial material changes from ocean sediments to crystalline rock. This transition cross-cuts the Ekström Ice Shelf cavity from ENE to WSW over its full width. Based on our simulations, such crystalline outcrops under ice shelves will result in a thicker but slower ice sheet over the last glacial cycle, compared to a thin and fast ice sheet linked to soft ocean beds which are mostly assumed for areas that lie below present day sea level (Pollard and DeConto, 2009; Pollard et al., 2016; Whitehouse et al., 2017). Interestingly, today's north-eastern most grounding line of Halfvarryggen ice rise coincides with this magnetic anomaly and the Explora Volcanic Wedge outcrop and thus likely with the presence of subglacial crystalline strata (Smith et al., 2019; Eisermann et al., 2020). We can therefore hypothesize that the spatial variations in subglacial strata also influence the position of present day grounding lines. Finally, the ramifications of heterogeneous ocean bed properties go beyond ice volume considerations. Different levels of basal traction strongly affect the magnitude of basal sliding. This in turn determines how much material is eroded underneath the ice sheet and transported across the grounding line. As erosion rates are commonly approximated as basal sliding to some power (e.g. Herman et al., 2015; Alley et al., 2019; Delaney and Adhikari, 2019), any differences in basal sliding velocities are exacerbated when erosion volumes are computed. This uncertainty in eroded material produced has implications for how much sediment is available at the ice bedrock interface and therefore if it is a hard or soft bed interface and its temporal variability.

5.3 Grounding line and ice sheet stability

The identified stable grounding line positions are not controlled by a single specific forcing alone, but are due to a combination of sea level forcing, basal traction of the ocean bed, and ocean bathymetry. Other forcing mechanisms such as the SMB and BMB are of secondary importance. However, the relative stability of grounding line position (<7 km of grounding line retreat) in the last 9,000 years of the retreat phase in both simulations coincides with the period of little sea level variations, leading us to conclude that at least for the retreat phase, sea level forcing is the most important model forcing. The ~~modelled higher grounding line fidelity~~ earlier onset of grounding line motion in the retreat phase for the soft bed can be attributed to the fact that ice discharge for the soft bed simulation is dominated by basal sliding and higher ice velocities. In comparison, in the hard bed simulation ice discharge is dominated by internal deformation and almost no basal sliding, resulting in a much thicker ice sheet. This means that more ice needs to be removed before the grounded ice can detach from its subglacial material and initiate grounding line motion, thereby resulting in a much slower response time to changes in the model forcing. While our employed modelling approach makes it unlikely that the timing of our modelled stable grounding line positions are correct, they can still serve as rough spatial markers of areas where depositional landforms such as grounding zone wedges or other geomorphological markers may be found.

5.4 Hysteresis of ice sheet simulations

The modelled grounding line advance in the first ∼300 years, we attribute to the fact that our ice sheet geometry is not completely in steady state after initialisation. This is due to inconsistencies of the model forcing (e.g. BMB parameterisation) in combination with boundary datasets (e.g. cavity topography). However, this does not affect our conclusions regarding ice sheet hysteresis. Our results highlight the importance of different ocean bed properties onto the ice sheet's hysteresis behaviour. This underlines the dependence of the final ice sheet geometry on the model's initial state over timescales of a glacial cycle or longer. While bedrock geometry has long been identified as a cause for hysteresis behaviour in ice sheet models (e.g. Schoof, 2007) and remains an important indicator for future ice sheet vulnerability, our simulations show that in the absence of retrograde sloping bedrock topography, ~~hysteresis can also be introduced by~~ varying ocean bed properties also have the potential to induce hysteresis. However, this result could also be caused by a combination of the non-uniqueness of the Stokes contact problem for non-sliding beds and an under resolving of the grounding line zone (e.g. Nowicki and Wingham, 2008). Despite very similar model forcing, our simulations result in a non-linear response of ice sheet evolution that is exclusively controlled by ocean bed properties, revealing an additional challenging problem for model simulations over at least one advance and retreat cycle (Pollard and DeConto, 2009; Gasson et al., 2016). This also means that the employed modelling framework will likely not result in the correct ice sheet geometry at the LGM due to non-linear feedback mechanisms such as the marine ice sheet instability (Schoof, 2007; Durand et al., 2009), the height mass balance feedback (Oerlemans, 2002), and remaining uncertainties regarding the subglacial topography.

5.5 Model limitations

The primary focus of the modelling framework was to extend the applicability of FS ice sheet models to glacial cycle timescales. This means that simplifications were made to other model components that we list here. We regard each of these simplification as a future avenue to improve upon the presented results.

The modelling approach presented here is tailored towards capturing ice and grounding line dynamics to high accuracy at the cost of comparatively naive parameterisations for the SMB and BMB which can be improved in the future. Also, by approximating hard and soft ocean beds through a time and space invariant friction coefficient, we omit spatial gradients in the thickness, grain size and cohesion of the ocean bed substrate. We therefore assume that properties of hard bed and soft bed areas at the start of the simulation remain constant throughout the simulation. This means, areas in which little or enhanced basal sliding occurs in the modelling domain stay constant.

At the underside of the grounded ice sheet, we use a linear Weertman sliding law that relates the basal shear stress to the basal sliding velocity. In comparison to the non-linear Weertman sliding law, the linearised version has a tendency to reduce grounding line fidelity (e.g. Schannwell et al., 2018; Brondex et al., 2019). While this type of sliding law is still widely used (e.g. Ritz et al., 2015; Cornford et al., 2015; Nias et al., 2016; Yu et al., 2017; Schannwell et al., 2018; Brondex et al., 2019), pressure limited sliding relations (e.g. Tsai et al., 2015) are becoming more popular in the modelling community. The difference between Weertman and pressure limited relations is that the latter take effective pressure into account. This means that basal drag goes to zero near the grounding line and reduces to a plastic sliding relation (Brondex et al., 2017). ~~However, this~~ This results in the basal drag becoming independent of the sliding velocity. Most previous studies using pressure-limited relations confine areas of lower basal drag ~~area is limited to~~ to within a few kilometers upstream of the grounding line (e.g. Schannwell et al., 2018; Brondex et al., 2019). There is however evidence from observations and modelling that areas of low basal drag can extend much farther inland (Joughin et al., 2019). Studies that have investigated the effect of the different sliding laws on grounding line retreat have found that the pressure limited relations lead to enhanced grounding line retreat (e.g. Schannwell et al., 2018; Brondex et al., 2019) in comparison to Weertman sliding laws. However, it is difficult to judge how much a pressure limited sliding law would affect our results as up to now no study has investigated this effect over an advance and retreat cycle.

Moreover, we have not considered glacial isostatic adjustment (GIA). Until recently, GIA was considered to be only important on timescales exceeding 1,000 years. However, recent progress has revealed that due to lower than previously assumed mantle viscosities, response times of GIA to ice unloading can be as short as five years for certain sections in Antarctica (Barletta et al., 2018; Whitehouse et al., 2019). While present day GIA rates for East Antarctica are relatively low ($\sim$1mm/yr, see Martín-Español et al. (2016)) in comparison to regions of high mass loss in Antarctica, the effect over 20,000 years could amount to $\sim$20 m of elevation drop for the subglacial topography. This number is small in comparison to, for example, sea level variations ($\sim$130 m), but may nevertheless result in a grounding line position that is not as far advanced at the glacial maximum as presented in our simulations.

6 Conclusions

Our simulations unlock a new time dimension for the applicability of FS ice sheet models on the regional scale. Application of an iterative solver reduced computation times in comparison to previous simulations by $\sim$80% and extended the temporal range of FS simulations by a factor of 40 compared to previous studies. This provides an important step towards including higher order physics into paleo ice sheet simulation and reduce uncertainties arising from approximations to the ice flow equations. Being able to simulate ice deformation to high accuracy over glacial timescales also opens opportunities for a better understanding of a number of subglacial processes (e.g. basal erosion).

We find ice volume differences of $>$50% over a glacial cycle that are exclusively caused by differing ocean bed properties. The different ocean bed properties also result in different ice sheet growth and decay pattern with the thick and slow flowing hard bed simulation exhibiting strong hysteresis behaviour. This is completely absent in the thin and fast flowing soft bed simulation. As recent geophysical observations (e.g. Gohl et al., 2013; Smith et al., 2019; Eisermann et al., 2020) indicate a more hetereogenous substrate distribution (sediments vs. crystalline bedrock) than previously thought, this could have important consequences for past stable ice sheet geometries and grounding line positions as well as for the present and future response of the ice sheet's grounding line to ocean warming.

Code availability. The Elmer/Ice code is publicly available through GitHub (https://github.com/ElmerCSC/elmerfem, lastaccess: 05 November 2019). All simulations were performed with version 8.3 (rev. 74a4936). Elmer/Ice scripts including all necessary input files to reproduce the simulations are available at https://doi.org/10.5281/zenodo.3564168 (Schannwell (2019), last access: 28 February 2020).

Video supplement. There is one video supplement SI video 1

Author contributions. CS and RD conceived the study with input from OE, TAE, and CM. Simulations were run by CS. E.C.S and H.E. provided new cavity topography data. The manuscript was written by CS and RD, and all authors contributed to editing and revision.

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