Anonymous Referee #1

General Comment

Kallenberg et al report on a new approach for estimating ice dynamic rates for mass balance assessment from EO data and apply this for a study site in East Antarctica where an increase in ice mass has been observed in recent years. In the approach the ice dynamic are estimated by combining modelled SMB rates with gravity observations from GRACE and laser altimetry observations from ICESat. The derived IDR is combined with modelled elevation changes due to snow processes for comparison with measured elevation changes from ICESat. The authors find the estimated ice dynamic rates from GRACE and ICESat of similar magnitude and modelled elevation changes in correlation with direct altimetry observations. This is a well written, illustrated an referenced methodological manuscript and a valuable and original contribution for the glaciology community. I would suggest a few minor corrections/clarifications to help improve the manuscript.

Specific Comments

Pg2 – Ln 20-23: There appears to be a mix up here. Ice velocity is derived from satellite radar interferometry and related methods in the mentioned studies, not from altimetry.

Changed on Pg2, Ln24.

Pg5 – Ln 12-17: Studies have shown that leakage from oceanic geophysical signals may bias mass rate estimates for Antarctica from GRACE significantly. How is this quantified or dealt with in the approach?

We converted the altimetry observations into spherical harmonics so that we represent the ice height information with the same spatial resolution as the mass change information. By doing this we impose the same potential leakage on to the altimetry observations. We added a comment to this effect on Pg.13, Ln.18-20.

Pg6 – Ln12: trend due to SMB

Changed on Pg11, Ln 25.

Pg7 – Ln24: Eq. 7 & Eq. 9 do not exist, I assume Eq. 4 & Eq. 5 are meant

This has changed by incorporating the appendix into the body of the manuscript. The new equation numbers have been adapted.

Pg13 – Ln4: Ligtenberg et al.

Changed on Pg 7, Ln8.

Pg13 – Ln18: Eq. A3

This has changed by incorporating the appendix into the body of the manuscript. The new equation numbers have been adapted.

Pg15 – Ln5&Ln11: Equation (A7)

This has changed by incorporating the appendix into the body of the manuscript. The new equation numbers have been adapted.

Pg15 – Ln25: (Appendix A2)

The appendix has been deleted.
Pg16 – Ln14: I assume Eq. 1 is meant here

This has changed by incorporating the appendix into the body of the manuscript. The new equation numbers have been adapted.

Pg16 – Ln15: I assume Eq. A3 is meant here

This has changed by incorporating the appendix into the body of the manuscript. The new equation numbers have been adapted.

Fig. 1: Invert color scale

The plot is of the velocity of ice. The colour scheme is therefore somewhat arbitrary. We reversed the colour scheme but it turns the figure into something very dark, with the background colour over the region being dark blue rather than white. We prefer our original colour scheme and so have chosen to leave it as it was.

Fig. 2: What is the white cross in Fig. 2a

Added, now Figure 3, Pg26.

Fig. 2: Y-axis label: dM/dt in mm or mm/yr?

The Y-axis label has been updated, Fig. 3b, Pg.24.

Fig. 3: (m yr-1): please check scale here

Corrected, now Fig.4, Pg.25

Anonymous Referee #2

Summary:
The manuscript back-calculates dynamic mass loss from East Antarctica using repeat GRACE and satellite altimetry observations paired with models of surface mass balance, firn compaction, and glacial isostatic adjustments. The authors find that rates of dynamic mass change inferred from the two different satellite observational platforms yield similar results, indicating that either platform can provide reasonable estimates of dynamic mass change when paired with current model outputs. The authors also infer that the good agreement between dynamic mass change estimates derived from the two different satellite platforms indicates that the most up-to-date RACMO model provides accurate estimates of surface mass balance for East Antarctica. The results of the paper are interesting in that they show models of SMB, firn compaction, and GIA are accurate enough to allow us to tease-out the ice dynamics signal from repeat gravity and laser altimetry observations. There are several relatively small modifications that would improve the overall quality of the manuscript, including: 1) removal/adjustment of commas (described below), which often break sentences into somewhat awkward fragments, 2) incorporation of data and methods description that is currently contained in the appendices, and 3) change references to "ice dynamics rates" to "mass change rates due to dynamic change" or something similar My biggest concern is in regard to (2) above. I highly recommend that more information on the datasets and methods is included in the body text of the manuscript. The details regarding how (1) you process the GRACE and ICESat data, (2) the versions of RACMO used in the study differ from each other, and (3) the firn compaction converts SMB input to estimates of surface elevation change are incredibly important for assessing the validity of the results and for reproducing the method elsewhere. Although this will lengthen the manuscript, I think that adding more detail regarding the data used in the analysis is imperative. Given that the firn layer in Antarctica can be tens of meters thick and that the interpretation of altimetry data is incredibly sensitive to the accuracy of the firn compaction model (see Zwally et al., Journal of Glaciology, 2015 for an example) this is particularly important for the firn compaction model. As a follow-up, why use RACMO2.1 to run the firn compaction model when RACMO2.3 was found to produce better results for the GRACE-ICESat dynamic change comparison? This is inconsistent.

1) remove/adjust commas

Changes performed as mentioned in detailed comments.

2) move data and method description from appendices into text

The appendices were incorporated into the body of the manuscript.

3) change “ice dynamic rates” to “mass change rates due to ice dynamic changes” or similar.

Changed throughout the manuscript.

4) Why use RACMO2.1 for the firn compaction model if RACMO2.3 was found to estimate SMB more accurately.

As explained within the firn compaction section the model has been tuned to fit available observations. Using RACMO2.3 means we would have to fine-tune the model differently in order to adjust it to available observations, due to the differences in the SMB.
This is already explained in the manuscript and we didn’t add anything further.

Detailed Comments:

p. 2, line 4: Change to “and bedrock uplift rates.”

Changed on Pg2, Ln4.

p. 2, line 5: Move the comma after “ice lost” forward so it is after “balance”

Changed on Pg2, Ln5.

p. 2, line 11: Either move the definition of firn so that it comes earlier in the sentence or remove it entirely. Currently it’s in an awkward location.

Changed on Pg2, Ln14.

p. 2, line 14: Change to “results in a change in the ice sheet surface elevation without...”

Changed on Pg2, Ln15.

p. 2, line 15: Remove “potentially”

Removed on Pg2, Ln17.

p. 2, lines 20 -21: This sentence should be broken - up so it’s easier to read. It took me at least 2 attempts to pause in the appropriate places and follow the entire sentence.
I recommend something like: “Ice discharge is the product of the ice velocity and thickness across the grounding line. Satellite rate altimetry is used to retrieve information about ice surface velocity. Ice thicknesses are estimates from

airborne radar or, in the absence of radar observations, using surface elevation observations under the assumption that the ice is floating."

Changed and rewritten on Pg2, Ln 23-26.

p. 3, lines 6-7: Change to..."(GIA), which is the response of the lithosphere to changes in surface loading."

Changed on Pg3, Ln10.

p. 3, line 10: Remove the comma before "by"

Removed on Pg3, Ln14.

p. 3, lines 14-16: Either remove these lines "We combined our... to direct observations of ice surface height from ICESat." or rephrase. I don't think you need to go into much detail at this point and these two sentences are currently really difficult to follow.

Changed on Pg3, Ln 18-21.

p. 3, lines 18-20: I find this sentence confusing. I follow that you obtain similar estimates of dynamic mass change from GRACE and ICESat observations but I don't understand what you mean that they "can be used to model surface elevation changes that are comparable with altimetry observations". ICESat data are altimetry observations. Do you mean that you can use GRACE data to estimate the surface elevation change expected due to dynamic change using the methods you describe here? If so, you need to revise the sentence so that is clear.

Changed on Pg3, Ln 18-21.

p. 3, line 22 to p. 4, line 6: You start off by stating that there is a positive mass change trend across the study region but then go on to say the region is roughly in balance.
Please revise to present a more consistent background on mass change estimates from the region.

We have restructured and rephrased the sentences to explain the mass balance situation more clearly for our study region.
Changed Pg3, Ln 25 to Pg4, Ln 9.

p. 4, lines 18-26: I assume that the "slope correction" you present in Equation 1 is an effort to account for drifting snow across the ice sheet surface. I believe this is already accounted for in RACMO (as you state in the appendix) so you may be "double -counting" for snowdrift. If you are referring to some other mechanism,

please make that more clear. Additionally, the last sentence here should state specifically where you obtain estimates for these variables, not just what is "typical".

The "slope correction" introduced by Helsen et al. (2008) (supporting material, page 2) is applied to the parameterisation of Kaspers et al. (2004) to estimate surface snow densities. Their relation was found to obtain more realistic values that are in better agreement with available density observations. It is still used by Ligtenberg et al. (2011) to estimate their surface snow densities for the firn compaction modelling. Neither from the Helsen et al. (2008) nor the Ligtenberg et al. (2011) publication it is stated that the "slope correction" is applied to account for drifting snow but rather to fine-tune the surface snow parameterisation.
We reworded this part when incorporating the appendix into the manuscript: Pg7, Ln23 – Pg.8, Ln3.

p. 5, line 1: Replace "seen" with "measured"

Changed on Pg10, Ln12.

p. 5, line 2: Remove comma before "as well as the effect of GIA"

Removed on Pg10, Ln13.

p. 6, lines 5-6: Replace with "The solutions to Equations 4 and 5 are the change in ice mass, DM/dt, and surface elevation, dH/dt, associated with changes in ice dynamics", with the proper subscripts and superscripts added.

Changed on Pg11, Ln 15.

p. 6, line 6: Replace "mass rate and height rate" with "rate of change in mass and surface elevation"

Changed on Pg11, Ln19.

p. 6, lines 17-19: There's an assumption inherent in these conversions that the entire ice sheet thickness is composed of glacier ice when we know this is not the case. It would be helpful to have an estimate provided somewhere of the fraction of the total thickness that is firn versus ice. If the firn column is only ~50m but the ice is ~2000m thick, this assumption is fairly reasonable. However, if the ice is relatively thin and/or the firn column is very thick, then the density used for these conversations should be reduced.

Our assumption is that a mass change observed by GRACE is caused by a change in SMB, GIA and ice dynamic. As firn compaction has no effect on GRACE observations, any changes within the firn layer are covered by the rate of change due

to SMB, which occurs within the firn layer. We estimate the rate of change due to SMB and GIA, and assume that the remaining signal belongs solely to changes in the glacier ice, specified by ice dynamic signal.
Elevation variations detected by altimetry are caused by changes in SMB, firn compaction, GIA and ice dynamic. We estimate the rate of change due to SMB, firn compaction and GIA, and assume that the remaining signal belongs solely to changes in the glacier ice, specified by ice dynamic signal.
As we assume that the remaining signal is solely due to changes within the glacier ice, we have chosen to use the density of glacier ice.
It is true that ice dynamic changes affect the entire ice sheet (ice and firn column) but here we address both columns separate from each other, and changes within the firn layer (SMB+firn compaction) have been considered separately.

We added a brief explanation of this at Pg. 11 Ln.16-19.

p. 6, line 20 to p. 7, line 4: Shouldn't you be adding dH/dt estimate from GRACE to the dH/dt estimates for SMB and firn AND dH/dt from GIA to get a signal that is equivalent to the ICESat dHdt? Also, if dH/dt of ice from GRACE and ICESat are not equal, the discrepancy could also be caused by the spatial and temporal variations in the density of the ice used in the conversion, inability of the SMB and/or firn model to realistically simulate surface changes, in addition to errors/limitations in data processing techniques. You should list all potential sources of error briefly here.

This is correct, changes due to GIA should be included. We left it out of the equation, because the satellite data used has already been corrected for GIA.
We added it to Eq.13 on Pg12..
List of potential error sources added on Pg12, Ln15-17.

p. 7, lines 9-10: Split the sentence so that it reads "...measured by GRACE. Figure 2a shows the map of the GRACE mass change signal and Figure 2b shows a time series for a coastal location near..."

Changed on Pg12, Ln22.

p. 7, lines 17-19: Why not use the ICE_5G_C results? Are the other results more realistic/better for some reason?

The ICE_6G_C model is an updated version of the ICE_5G_C and provides a better estimate on GIA uplift rates according to Peltier et al. (2015) and Argus et al. (2014). We therefore chose to use only their most recent model.

p. 7, line 20. Break into two sentences so that you now have "... snowfall and ice discharge. The GIA-corrected GRACE mass change data suggest a positive mass trend of ~32 +/-8 mm w.e. yr^-1 between 30°E and 70°E and a substantial increase in mass from 2003-2009 (Fig. 3b)." The anomaly you list should be

averaged over this entire region. I think the anomaly is only estimated over a smaller region currently, which is a bit misleading. Also, how can you attribute this to SMB? The SMB signal is actually from RACMO, correct? Are you presenting the mass gain estimated by RAMCO for SMB only or the GRACE SMB+discharge mass signal?

The value of the mentioned anomaly was pointing out the two regions with the strongest observed anomaly, which was stated in the original sentence "with a positive anomaly of ~32+/-8 mm w.e. yr^-1 near 40°E and 55°E".
We did break the sentence down and removed the stated anomaly for a better understanding.
It is also correct that the trend should not be attributed to SMB, it should have said "mass" instead of "SMB". This has been changed.

Changed on Pg13, Ln12-13.

p. 8, line 3: How do you convert the ICESat data into spherical harmonics? What precisely does this mean? Does it mean you spatially average the data in some way? Please elaborate.

The values of the ice dynamic estimates have been converted into the C and S coefficients of spherical harmonics. This allowed us to convert our spatial grid into a spherical harmonic grid. This is a basic conversion and we don't feel it necessary to explain it in detail within the manuscript.

p. 8, lines 5-17: I find this section to be really difficult to follow. You should make it clear that you are using the RACMO models to estimate SMB contributions to the GRACE and ICESat signals. Saying "For both RAMCO2 models the ice dynamic estimates" and "Using RAMCO2.3 the ice dynamic estimates" reads a bit like you are estimating the dynamic signal directly from RACMO.
Are the rates of dynamic mass change averages over the entire time period for each observational platform?
Are the RMSE estimates the RMSE of the difference in SMB between the two RACMO versions over the entire study region?
It would be helpful to have numeric estimates clearly presented in this section along with their error estimates.
It would also be helpful to focus on just the difference in RACMO SMB over the study region, with discussion as to which version produces more realistic results when used to tease-out the dynamic signal, then compare the dynamic change estimates. Right now there's just too much going on at once.

We have modified the entire section to improve our discussion. We clarified that we use the modelled SMB estimates using RACMO: Pg13, Ln23-27.
We added that the mentioned rates of change provide the information that the obtained rates are between the two values within a certain region: Pg13, Ln26-27, and that the provided RMSE estimates are averaged over the study area: Pg14, Ln 6.

p. 8, lines 18-26: As mentioned earlier, I think you need to add dH/dt from GIA into your GRACE-derived dH/dt estimates. You should also include values for the trends you discuss here so that the reader can discern "strong" and "weak" trends.

Initially we included values for the discussed trends. However, following the discussion including values for the trends reads somewhat awkward, which was the reason for us to exclude specific values. The reason why we find this not suitable is that we compare the regions in where the positive and negative trends have been modelled and correlate with the ICESat observations, rather than presenting the actual values of the trends. As stated in the manuscript the values don't correlate precisely with each other. This makes it hard to follow actual trend-results when reading the discussion and comparing the figures. As we find that including values for the trends is not fitting we have not added any values.

dH/dt from GIA added on Pg14, Ln10.

p. 9, lines 12-16: You should include maps of uncertainty with the dynamic change estimates in the text (move Figure A2 to the body of the manuscript).

Uncertainties and figure added to main text on Pg15-16, and P. XY, respectively.

p. 10, line 1: Can you substantiate this remark that the different GIA models have a small effect on the ice dynamic change estimates? There have been rather large error bars in previous Antarctic mass change estimates from GRACE that have been largely attributed to uncertainty in the GIA signal.

Estimated uplift rates due to GIA in our study region are between ~0-1mm (W12a, Whitehouse et al., 2012) and ~0-3mm (ICE-6G_C, Peltier et al., 2015). While the effects are larger on the GRACE observations than the altimetry observations it still remains very small. Consequently, even a 100% error in the modelled GIA values would have only a very small impact on the results shown.

We added a comment to this effect on Pg.13, Ln.10 and Pg.16, Ln1-5.

p. 10, line 3: Replace "We believe" with "Our data suggest"

Changed on Pg16, Ln25.

p. 10, line 5: Is the different statistically significant?

Yes, the difference is statistically significant.
This was added on Pg. 16, Ln27 to Pg. 17, Ln2.

p. 10, lines 12-13: Replace with "Thus, a comparison of estimated changes in ice dynamics derived from GRACE and altimetry observations not only provides information about dynamic mass change, but may also help to identify regions where models fail to accurately simulate variations in SMB."

Changed on Pg17, Ln 8.

p. 12, line 1: Remove comma after "grid"

Removed, now on Pg6, Ln4.

p. 16: What about uncertainties associated with GIA? I expect that these are quite large but they are seemingly overlooked. They are likely difficult to quantify but you could likely obtain uncertainty estimates computed for each GIA model from the model developers.

Uncertainties for GIA models are not provided with the models. GIA models are based on available observations and assumptions about the Earth's viscosity profile and ice history. They are tuned to fit by performing a parameter search for the best-fitting ice sheet history and earth rheology values. Velicogna and Wahr (2006) estimated an uncertainty by defining a lower bound and an upper bound, using the minimum and maximum trend of a GIA model. Their uncertainty corresponds to the bounds of the GIA range, while their best estimate on GIA uplift rates is the midpoint of this range.
Nevertheless, across our study region GIA uplift rates are small and differences between the models are < 2 mm/yr. Therefore, the error in the modelled GIA signal in our study region are likely to be small.

We added a comment on Pg.16 Ln.1-5.

Figure 2: Include the name of the model used in the time series of GIA.

The viscoelastic uplift shown in the figure is derived from GRACE observations using the empirical formula that was shown by Purcell et al. (2011) to be valid for all realistic ice/earth models where there has been no change in load for the past 5000 years. That is, the result is not related to any particular GIA model at all.
This is explained in Section 3.1 on Pg.5, Ln.14-17.

Figure 3: Remove the last sentence in the caption.

Removed, now Figure 4, Pg.27

Figure 4: The dynamic mass change rates are obtained from GRACE and ICESat using RACMO to parse-out the SMB signal.

Changed, now Figure 5, Pg. 28.

A new approach to estimate ice dynamic rates using satellite observations in East Antarctica

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Abstract.

Mass balance changes of the Antarctic ice sheet are of significant interest due to its sensitivity to climatic changes and its contribution to changes in global sea level. While regional climate models successfully estimate mass input due to snowfall, it remains difficult to estimate the amount of mass loss due to ice dynamic processes. It's often been assumed that changes in ice dynamic rates only need to be considered when assessing long term ice sheet mass balance; however, two decades of satellite altimetry observations reveal that the Antarctic ice sheet changes unexpectedly and much more dynamically than previously expected. Despite available estimates on ice dynamic rates obtained from radar altimetry, information about ice sheet changes due to changes in the ice dynamic are still limited, especially in East Antarctica. Without understanding ice dynamic rates it is not possible to properly assess changes in ice sheet mass balance, surface elevation or to develop ice sheet models. In this study we investigate the possibility of estimating ice sheet changes due to ice dynamic rates by removing modelled rates of surface mass balance, firn compaction and bedrock uplift from satellite altimetry and gravity observations. With similar rates of ice discharge acquired from two different satellite missions we show that it is possible to obtain an approximation of the rate of change due to ice dynamics by combining altimetry and gravity observations. Thus, surface elevation changes due to surface mass balance, firn compaction and ice dynamic rates can be modelled and correlate with observed elevation changes from satellite altimetry.

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1 Introduction

Assessing and understanding ice mass balance of the Antarctic Ice Sheet (AIS) is challenging due to the remoteness and extensive ice cover of the continent, resulting in a sparse network of field observations to provide information about the climate, mass balance **and** bedrock uplift rates. In order for an ice sheet to be in balance**,** the amount of ice lost due to the dynamic processes of meltwater runoff and solid ice discharge over the grounding line, needs to be balanced by accumulated snowfall. If one exceeds the other, the ice sheet either gains or loses mass, resulting in a change in ice sheet mass balance (Cuffey and Paterson, 2010). The surface processes of snowfall, snowmelt and subsequent runoff, sublimation, evaporation and snowdrift add, remove or distribute snow and define the surface mass balance (SMB) (e.g. Lenaerts et al., 2012; Van Wessem et al., 2014). Changes in SMB occur primarily in the firn layer that covers the AIS, the intermediate product between snow and ice (Ligtenberg et al., 2011). Temperature variations, overburden pressure, deformation and repositioning of snow grains causes snow to densify until it reached the density of glacier ice (~917 kg m^{-3}) (Herron and Langway, 1980**). The intermediate product is called firn, and changes in SMB occur primarily in the firn layer that covers the AIS** (Ligtenberg et al., 2011). **This results in a change in the ice sheet surface elevation without changing the mass of the ice sheet.**

When thoroughly evaluated with field observations and downscaled using statistical interpolation methods, regional climate models can be used to simulate fields of SMB components, temperature and near-surface wind speed. Ice loss rates can be obtained by combining individual estimates of accumulation, ablation and dynamic ice loss, with the difference between mass input and mass output providing the mass balance of the ice sheet. While SMB can be taken from regional climate models, estimates on ice discharge are limited and difficult to obtain. **T**he amount of ice discharge **can be estimated by obtaining the product of ice velocity and ice thickness** across the grounding-line**. Satellite radar interferometry is used to retrieve information about ice velocity rates. The ice thickness is estimated from airborne radar or, in the absence of direct observations, using surface elevation observations under the assumption that the ice is floating once it has crossed the grounding line** (Rignot and Thomas, 2002; Rignot et al., 2008; Allison et al., 2009). Commonly, **changes due to** ice dynamics are either taken from these by satellite altimetry derived estimates

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(Shepherd et al., 2012; Sasgen et al., 2013), or assumed to be insignificant when studying short-term changes (e.g. Ligtenberg et al., 2011). However, unexpected changes in ice sheet dynamic have been observed in the past decades, with some glaciers found to accelerate, while others decelerated (Rémy and Frezzotti, 2006). In general, ice dynamics are not well known and information about ice dynamic variations are limited (Rignot, 2006; Rignot et al., 2008). This becomes an issue when assessing ice mass balance and surface elevation changes, or establishing ice sheet models.

Although satellite observations help provide information about temporal and spatial changes in ice mass and ice volume, large uncertainties remain when interpreting the signals and assigning the origin of change. Ice mass balance can be measured directly from gravity observations but needs to be separated into the possible changes caused by SMB, ice dynamics and Glacial Isostatic Adjustment (GIA), **which is** the response of the lithosphere to changes in surface loading. Changes in ice sheet thickness can be obtained from altimetry observations but need to be separated into the change caused by SMB, ice dynamics, GIA and/or firn compaction. Observed elevation changes can subsequently be converted to changes in mass by employing firn densities.

In this study we obtain an estimate of **ice sheet changes due to** ice dynamic rates by combining modelled SMB rates using the Regional Atmospheric Climate MOdel (RACMO2), Gravity Recovery And Climate Experiment (GRACE) and laser altimetry observations from the Ice, Cloud and land Elevation Satellite (ICESat). We **found that the attained estimates of ice dynamic changes obtained from GRACE and ICESat are of similar magnitude. In conjunction with our estimates on our rate of change due to ice dynamics we model the rate of change of the ice surface and compare our results with direct observations taken from ICESat measurements.** A study site in East Antarctica has been chosen due to the increase in mass that has been observed there by GRACE and altimetry, suggesting a thickening of the ice sheet.

2 Study area

The chosen study area combines Enderby Land, Kemp Land and MacRobertson Land, and parts of Dronning Maud Land and Princess Elizabeth Land (hereafter referred to as Enderby Land for simplicity). The study area is assumed to be a stable region (e.g. Rignot et al., 2008), with the ice sheet

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predominantly located on bedrock above sea level, making it less vulnerable to changes in ocean temperatures. The major outlet glaciers of this region are the Lambert and Mellor glaciers feeding the Amery Ice Shelf in the east, together with the smaller (~3000 km^2) Fisher, Scylla and American Highland Glaciers. Only smaller glaciers are found along the remaining coastal region of Enderby Land, including the Shirase, Rayner, Thyer and Robert glaciers (Fig. 1). Previous research **based on the mass budget method** found the ice sheet to be largely in balance across this area, possibly even slightly thickening (Rignot, 2006; Rignot et al., 2008; Rignot et al., 2013). **A general positive mass trend across this region has also been recorded by gravity and altimetry observations (e.g. Shepherd et al., 2012; Sasgen et al., 2013).**

3 Data sets and implemented models

We use observational measurements of mass variations from the Gravity Recovery And Climate Experiment (GRACE) and surface elevation changes observed by laser altimetry using the Ice, Cloud and land Elevation Satellite (ICESat). **The regional** climate model RACMO2/ANT is used to model the trend in SMB and to force the firn compaction model. Two versions of the RACMO2 model are used here, RACMO2.1 and RACMO2.3. The SMB used throughout this paper is the sum of snowfall, evaporation/sublimation, snowdrift and runoff. The SMB components are provided in kg m^{-2} t^{-1}, where t is the temporal resolution of the model.

3.1 GRACE

We use the monthly gravity field solutions CNES/GRGS RL03-v3, provided by the Groupe de Researches de Géodésie Spatiale (GRGS). The RL03 solutions have a spatial resolution of degree and order 80 (Lemoine et al., 2013) and have been chosen due to the stabilisation process that is applied to reduce noise in form of North-South striping. This is achieved by regularising the inversion for spherical harmonic coefficients (Bruinsma et al., 2010).

Temporary changes in the Earth's gravity field can be related to changes in surface mass due to the distribution of mass, as well as the elastic and viscoelastic (GIA) response of the lithosphere, the instantaneous and long term signal to changes in surface load (Wahr et al., 1998). We obtain

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mass anomalies by applying the equations that relate mass changes to gravity changes (Wahr et al., 1998) to obtain the change in mass due to SMB:

$$U^{w.e.}(\theta,\lambda,t) = R\sum_{n=2}^{N}\sum_{m=0}^{n} P_{nm}(\cos\theta)\frac{2n+1}{1+k_n^{elast}}\left(\Delta C_{nm}(t)\cos m\lambda + \Delta S_{nm}(t)\sin m\lambda\right) \qquad (1)$$

and due to the viscoelastic deformation, or GIA:

$$U^{visco}(\theta,\lambda,t) = R\sum_{n=2}^{N}\sum_{m=0}^{n} P_{nm}(\cos\theta)\frac{h_n^{visco}}{k_n^{visco}}\left(\Delta C_{nm}(t)\cos m\lambda + \Delta S_{nm}(t)\sin m\lambda\right) \qquad (2)$$

where R is the Earth's radius, P_{nm} are the fully normalised Legendre functions, n and m are degree and order of the spherical harmonic coefficients, θ and λ are colatitude and longitude, and ΔC_{nm}, ΔS_{nm} are the spherical harmonic coefficients, at time t, of the GRACE anomaly fields. k_n and h_n are the elastic Love loading numbers (e.g. Pagiatakis, 1990) and the ratio of viscoelastic Love loading numbers (Purcell et al., 2011), depending on the degree. Purcell et al. (2011) showed that this empirical approximation permitted the accurate computation of viscoelastic uplift that was independent of any particular GIA model, provided that there has been no change in load for the past 5000 years.

3.2. ICESat

Various methods are used to estimate surface elevation changes from ICESat observations, using either along-track measurements or measurements directly taken from the crossover location. Due to perturbations in the orbit, deviations of the repeated ground track occur and it is necessary to determine the surface topography to correct elevation changes due to surface slope rather than changes in ice mass. Different methods have been applied to obtain surface elevation changes, using either along-track observations or crossover measurements (e.g. Slobbe et al., 2008; Gunter et al., 2009; Pritchard et al., 2009; Sørensen et al., 2011; Ewert et al., 2012).

Here we use the estimated rate of change of ice sheet elevation obtained from a newly developed technique that combines both crossover and along-track observations (Hoffmann, 2016). The method allows estimation of the local surface slope using a digital elevation model that has been derived from gridded estimates of ice height at ICESat crossover points. Over a crossover grid that geographically spans all campaign crossovers of a location, a static grid was created on which heights were interpolated at the epochs of all campaigns. The estimate of the elevation change over time is made by computing a weighted least-square regression of the height time series of each grid node and then computing a weighted mean value for all grid nodes to derive the crossover height rate. This not only allows to assess height rates at one location over time, but also to evaluate a digital elevation model directly from the data, which is used to estimate the slope at crossovers (Hoffman, 2016).

The slope estimates at the crossovers are then interpolated to remove the surface slope from the along-track measurements. Although the elevation change estimates from along-track measurements are naturally less precise than the rate estimates at crossovers, combining both methods significantly increases the accuracy of the slope correction, providing a measure to validate along-track estimates (Hoffman, 2016).

3.3. RACMO2/ANT

The RACMO2/ANT regional climate model, used to obtain SMB estimates, adopts the dynamical processes from the High Resolution Limited Area Model (HIRLAM) and the physical atmospheric processes from the European Centre for Medium-range Weather Forecasts (ECMWF) (Reijmer et al., 2005) and is forced by ERA-Interim reanalysis data at the lateral boundaries (e.g. Ligtenberg et al., 2011; Lenaerts and van den Broeke, 2012). The latest version, RACMO2.3 (Van Wessem et al., 2014), extends available model data from 1979-2012 (RACMO2.1) to 1979-2015 (RACMO2.3) and improves the temporal resolution from 6-hourly (RACMO2.1) to 3-hourly (RACMO2.3) (Ligtenberg per. comm., 2016). The horizontal resolution is 27 km and the vertical resolution 40 levels. Individual SMB components are provided including snowfall, evaporation/sublimation, and snowmelt, as well as snowdrift in RACMO2.3. Over

Antarctica RACMO2/ANT is coupled with a multilayer snow model, which estimates meltwater percolation, refreezing and runoff, as well as surface albedo and snowdrift (Van Wessem et al., 2014). The update in the physical parameters of RACMO2.3 results in a general increase in precipitation over the grounded East Antarctic Ice Sheet, evaluated using in-situ observations, ice-balance velocities and GRACE measurements and showing a general improvement of the SMB (Van Wessem et al., 2014).

3.4. Firn compaction

We developed a firn compaction model based on the firn densification model of Ligtenberg et al. (2011), using near surface climate provided by RACMO2.1. It is a one-dimensional, time-dependent model that estimates density and temperature individually for each layer and at each time step in a vertical firn column. The firn densification model of Ligtenberg et al. (2011) adds new snowfall instantly to the current top layer until the layer thickness exceeds ~15 cm (Ligtenberg, pers. comm., 2016), at which time it is divided in two layers. The properties of each layer are passed on to both layers. If a layer becomes too thin, due to compaction or surface melt, the layer is merged with the next layer and assigned the average properties of both layers. Our model has been simplified to improve the computational time. Rather than adding new snowfall instantly to the top layer, we compute the monthly sum of SMB and use the monthly averaged surface temperature to estimate the densification rate, density and new temperature to obtain the vertical velocity of the surface due to monthly firn compaction.

The model starts with a new firn layer created by the total SMB of one month and is built up by adding a new layer each month using monthly SMB values and mean surface temperatures. The surface snow density of each top layer is estimated using the proposed parameterisation of Kaspers et al. (2004), together with a proposed slope correction to improve the fit in Antarctica by Helsen et al. (2008):

$$\rho_s = -151.94 + 1.4266(73.6 + 1.06T + 0.0669A + 4.77W) \qquad \textbf{(3)}$$

where T is the average annual temperature (in K), A the average annual accumulation (in mm water equivalent (w.e.) yr^{-1}) and W the average annual wind speed 10m above the surface (in m s^{-1}). The densification rate is obtained using a dry snow densification expression proposed by Arthern et al. (2010):

$$\frac{d\rho}{dt} = CAg(\rho_i - \rho)e^{\left(\frac{-E_c}{RT}+\frac{-E_g}{RT_{av}}\right)} \qquad \textbf{(4)}$$

where C is the grain-growth constant (m s^2 kg^{-1}), independently calculated for densities below (C = 0.07) and above (C = 0.03) the critical density of 550 kg m^{-3}, A is the accumulation rate (mm w.e. yr^{-1}), g the gravitational acceleration, and ρ and ρ_i are the local density and the ice density (kg m^{-3}), respectively. The exponential term includes the activation energy constants (kJ mol^{-1}) for creep and for grain-growth, E_c and E_g, respectively, the gas constant R (J mol^{-1} K^{-1}) and the local temperature T, and annual average temperature T_{av} (K).

The process of liquid water percolation and refreezing is incorporated as a function of snow porosity P_s and density, as proposed by Coléou and Lesaffre (1998) (Ligtenberg et al., 2011; Kuipers Munneke et al., 2015):

$$L_W = 1.7 + 5.7\left(\frac{P_S}{1-P_S}\right) \qquad \textbf{(5)}$$

with the snow porosity:

$$P_S = 1 - \left(\frac{\rho}{\rho_i}\right) \qquad \textbf{(6)}$$

where ρ is the density of the layer and ρ_i the density of glacier ice.

The heat transport throughout the firn column is solved explicitly using the one-dimensional heat-

transfer equation (Cuffey and Paterson, 2010):

$$\frac{dT}{dt} = \kappa \frac{d^2T}{dz^2} \qquad \textbf{(7)}$$

with the thermal diffusivity κ and the depth *z*. Initially the heat-transfer equation consists of a term for heat conduction, advection and internal heating. However, initial heating is small within the firn layer and therefore neglected and the contribution of heat advection is taken into account by the downward motion of the ice flow (Cuffey and Paterson, 2010; Ligtenberg et al., 2011).

Finally, once the densification rate is estimated, the vertical velocity of the surface due to firn compaction, V_{fc}, can be assessed by integrating over the displacement of the compacted firn layers over the length of the firn column (Helsen et al., 2008):

$$V_{fc}(z,t) = \int_{z_i}^{z} \frac{1}{\rho(z)} \frac{d\rho(z)}{dt} dz \qquad \textbf{(8)}$$

where z is depth, ρ density and dρ(z)/dt the densification rate.

Ligtenberg et al. (2011) found that Equation (4) over-predicts the rate of densification for most regions in Antarctica, with the effect of the annual average accumulation being too large on the densification rate. They reintroduced an accumulation constant that previously had been proposed by Herron and Langway (1980) as α in A^{α} (below 550 kg m^{-3}) and β in A^{β} (above 550 kg m^{-3}), initially chosen between 0.5 and 1.1 but later assumed to be α, β =1 (Zwally and Li, 2002; Helsen et al., 2008). Ligtenberg et al. (2011) applied a modelled to observed ratio to correct for the accumulation dependence. We also found that Equation (4) over-predicts the rate of densification, depending on the rate of the average annual accumulation.

However, due to our use of monthly layers, the ratio proposed by Ligtenberg et al. (2011) is no longer valid and we introduce new α and β, depending on the accumulation rate (Table A1). The

values for α and β represent a best fit and have been obtained by investigating different values across several model runs. This means that the firn compaction model is adjusted to fit available observations and is therefore assumed to be correct and invariant of SMB model changes.

In Figure 2a we show the average annual rate of firn compaction across the study site and in Figure 2b the differences between our model and the model of Ligtenberg et al. (2011). Along the ice sheet margins and the Amery Ice Sheet our model overestimates their firn compaction rates by 5-10 cm yr^{-1}, while it underestimates rates by 7-12 cm yr^{-1} in most other areas further inland, with up to 15 cm yr^{-1} at two individual location near 28°E and between 68°E and 70°E. These differences are within our estimated uncertainty, based on the uncertainties provided for the modelled SMB from RACMO2 (Appendix A3).

4 Method to estimate the rate of change due to ice dynamic

A change in surface elevation, dH/dt, as measured by satellite altimetry is caused by a combination of processes that affect ice sheet thickness as well as the effect of GIA. The temporal change in surface height can be described as:

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$$\frac{dH^{ICESat}}{dt} = \frac{dH^{SMB}}{dt} + \frac{dH^{fc}}{dt} + \frac{dH^{ice}}{dt} + \frac{dH^{GIA}}{dt} \qquad (9)$$

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with the individual components representing elevation changes related to SMB (dH^{SMB}/dt), firn compaction (dH^{fc}/dt), ice dynamics (dH^{ice}/dt), and the elastic and viscoelastic response of the lithosphere combined under the term of GIA (dH^{GIA}/dt). While the process of firn compaction plays an important role in surface elevation changes it does not affect the overall mass balance of the ice sheet. Therefore, the general change in ice mass as detected by GRACE can be expressed as:

$$\frac{dM^{GRACE}}{dt} = \frac{dM^{SMB}}{dt} + \frac{dM^{ice}}{dt} + \frac{dM^{GIA}}{dt} \qquad (10)$$

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with the individual components representing a change in mass due to SMB (dM^{SMB}/dt), ice dynamics (dM^{ice}/dt), and GIA (dM^{GIA}/dt).

With the components that assemble dM^{SMB}/dt being represented by regional climate models simulating near surface climate in Antarctica, and dM^{GIA}/dt modelled by available GIA models, dM^{ice}/dt remains the only unknown in Equation 10. Therefore, an estimate of dM^{ice}/dt can be obtained by removing dM^{SMB}/dt and dM^{GIA}/dt from the GRACE observations:

$$\frac{dM^{ice}}{dt} = \frac{dM^{GRACE}}{dt} - \frac{dM^{SMB}}{dt} - \frac{dM^{GIA}}{dt} \quad . \qquad (11)$$

Similarly, the same approach can be used to obtain dH^{ice}/dt from altimetry:

$$\frac{dH^{ice}}{dt} = \frac{dH^{ICESat}}{dt} - \frac{dH^{SMB}}{dt} - \frac{dH^{fc}}{dt} - \frac{dH^{GIA}}{dt} \quad . \qquad (12)$$

The solutions to Equation 10 and 11 are the changes in ice mass, $\frac{dM^{ice}_{GRACE}}{dt}$, and surface elevation, $\frac{dH^{ice}_{ICESat}}{dt}$**, associated with changes in ice dynamics. We assume that changes within the firn layer have been taken into account by removing the rate of change due to SMB and firn compaction from the observations, and that the remaining signal is solely due to changes within the glacier ice. Therefore, w**e can convert to/from **the rate of change in** mass and **surface elevation** by dividing/multiplying by the density of glacier ice. Thus, observations from each satellite mission can provide an independent estimate of the ice dynamics.

We first correct both observational measurements, GRACE and ICESat, for GIA using three available GIA models: the W12a model of Whitehouse et al. (2012), the ICE-6G_C (VM5a) model of Peltier et al. (2015) and the recomputed version ICE6G_ANU of Purcell et al. (2016). Changes due to SMB are modelled using RACMO2.3/ANT, and the total trend **due to** SMB, for the period 2003-2009, is

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obtained using the monthly SMB (kg m^{-2} mth^{-1}). The change in dH^{SMB}/dt is acquired by dividing dM^{SMB}/dt by the density of surface snow (Eq. 3), and the rate of change due to firn compaction, dH^{fc}/dt, is taken into account by using our modelled firn compaction rates. Each month, the total SMB is computed and a monthly average firn compaction rate is removed from the SMB, before calculating the overall trend dH^{SMB}/dt over 2003-2009. Finally, the obtained $\frac{dH_{ICESat}^{ice}}{dt}$ rates can be converted to $\frac{dM_{ICESat}^{ice}}{dt}$ by multiplying by the density of glacier ice (~ 917 kg m^{-3}), while the $\frac{dM_{GRACE}^{ice}}{dt}$ rates are converted to $\frac{dH_{GRACE}^{ice}}{dt}$ by dividing by the density of glacier ice.

If ICESat and GRACE detect the same signal, the obtained $\frac{dM_{ICESat}^{ice}}{dt}$ estimates should correlate with $\frac{dM_{GRACE}^{ice}}{dt}$ and vice versa, $\frac{dH_{ICESat}^{ice}}{dt}$ with $\frac{dH_{GRACE}^{ice}}{dt}$. Moreover, modelling surface elevation changes ($\frac{dH^{Mod}}{dt}$) found by removing $\frac{dH_{GRACE}^{ice}}{dt}$ from the modelled dH^{SMB}/dt and dH^{fc}/dt estimates should approximate the ICESat observations:

$$\frac{dH^{Mod}}{dt} = \left(\frac{dH^{SMB}}{dt} - \frac{dH^{fc}}{dt}\right) - \frac{dH^{GIA}}{dt} - \frac{dH_{GRACE}^{ice}}{dt}. \qquad (13)$$

Conversely, $\frac{dH_{ICESat}^{ice}}{dt}$ not being equal to $\frac{dH_{GRACE}^{ice}}{dt}$ indicates that there must be an error, **which can be attributed either to errors in the data processing techniques, or the inability of the models to realistically simulate surface changes due to SMB, firn compaction and/or GIA.**

4 Results and discussion

The chosen region is part of a vast area in East Antarctica that shows an increase in mass, suggesting that the ice sheet is growing in this region. The signal the GRACE satellites detect includes changes in mass due to accumulation, ice discharge and GIA. In Figure 3 we show the observed change in mass measured by GRACE. Figure 3a **shows the map of the GRACE mass change signal and Figure 3b**

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shows a time series for a coastal location near 67°S 54°E for the entire operational period. In order to obtain the signal that is solely due to ice mass changes the contribution of GIA needs to be removed. In Figure 4 we show the GRACE signal corrected for GIA uplift rates using the ICE-6G_C (VM5) model by Peltier et al. (2015), W12a model by Whitehouse et al. (2012) and the recomputed version ICE6G_ANU of Purcell et al. (2016), respectively. Using ICE-6G_C (VM5) (Fig. 4a) significantly reduces the observed positive anomaly in Enderby Land, while applying W12a (Fig. 4b) and ICE6G_ANU (Fig. 4c) results in a smaller reduction of the mass anomaly, yielding a similar corrected GRACE signal. Due to the similarity between the W12a and ICE6G_ANU model the W12a model was chosen to correct the satellite observations for GIA, although the effect on the rate of change due to ice dynamic is insignificant between the models due to very small uplift rates across our study region. With the contribution of GIA removed, the signal should only comprise contributions from snowfall and ice discharge. The GIA-corrected GRACE observations suggest a positive anomaly between 30°E and 70°E and a substantial increase in mass between 2003-2009 (Fig. 4b).

The modelled trend in SMB and surface elevation due to SMB and firn compaction can now be removed from the GRACE and ICESat observations (Eq. 11 and Eq. 12), to obtain $\frac{dM^{ice}_{GRACE}}{dt}$ and $\frac{dH^{ice}_{ICESat}}{dt}$ and, subsequently, $\frac{dH^{ice}_{GRACE}}{dt}$ and $\frac{dM^{ice}_{ICESat}}{dt}$ by dividing (multiplying) by the density of glacier ice. We converted the rate of change of surface elevation due to the ice dynamic signal obtained from ICESat into spherical harmonics to be comparable with $\frac{dH^{ice}_{GRACE}}{dt}$. By doing this we represent the ice height information with the same spatial resolution as the mass change information, and impose the same potential leakage on to the altimetry observations. The estimated rate of change due to ice dynamics are shown in Figure 5, comparing estimates obtained using two different SMB models: RACMO2.1 and RACMO2.3.

We obtained similar rates of change due to ice dynamic by removing the modelled SMB estimates from both RACMO2 models and GIA uplift rates from GRACE and ICESat observations. Using SMB estimates from RACMO2.3 the ice dynamic estimates are significant smaller and primarily present between 30°E and 60°E with estimated rates between -0.08 to -0.13 m yr^{-1} obtained across the region. Using SMB estimates from RACMO2.1 yields a change due to ice dynamic of -0.08 m yr^{-1}

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and above along the entire ice sheet margin of our study region, **stretching across to 75°E**. **Generally, using RACMO2.3 the SMB estimates show a smaller difference between the obtained ice dynamic estimates obtained from GRACE and ICESat, improving results across the study area. However, regions remain that exhibit differences in the obtained ice dynamic signal of up to ± 0.05 m yr^{-1}**, (Fig. 5c and 5f). Significant changes **emerge** between **the rate of change due to ice dynamics obtained using** the former and latter RACMO2 versions, with an**, over the study region averaged,** RMS error of 0.019 m yr^{-1} and 0.021 m yr^{-1} for RACMO2.3 and RACMO2.1, respectively.

In both $\frac{dH^{ice}_{ICESat}}{dt}$ rates a positive trend is estimated across the centre of the region. This is the result of a slightly positive elevation trend that has been recorded by ICESat observations in region D (Fig. 6b). Finally, the total change in surface elevation is modelled, based on dH^{SMB}/dt, dH^{fc}/dt, **dH^{GIA}/dt** and $\frac{dH^{ice}_{GRACE}}{dt}$ (Fig. 6a). Using RACMO2.3, the result of the modelled rate of change of surface elevation reveals a similar pattern to the ICESat observations (Fig. 6b). In region A both the negative trend between 28°E and 32°E and the positive trend at 34°E is modelled. In region B a general negative trend is recorded along the ice margin with a positive trend near 46°E. Both signals appear in our modelled elevation trend, though at a smaller magnitude. Similarly for region C, which shows a general negative trend across the region, with the lowest trend near 51°E and a strong positive signal at 56°E. While the general negative trend is obtained in the model, the strong negative signal near 51°E is not present. The strong positive signal at 56°E is modelled, although it appears slightly over predicted, covering a larger region than seen in the ICESat observations. Across region D ICESat monitored an overall increase in elevation, especially near 70°E, together with a slight decrease in surface height along the margin between 58°E and 70°E and at the Mellor Glacier (Fig. 1) near 68°E. Similar to the ICESat observations the general positive trend across the region is modelled, together with the positive signal near 70°E, as well as a slight negative trend across the margin. However, the strong negative trend at the Mellor Glacier is lacking, though the region shows a slight negative trend. Although the modelled trend in surface elevation suggests similar behaviour to the altimetry observations, the signal generally appears damped compared to the ICESat observations. This is likely caused by the loss of spatial resolution through the use of degree 80 spherical harmonics (the resolution of the GRACE gravity fields) to

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remove the ice dynamic **signal**.

Uncertainties are estimated for the satellite observations and models **individually, and error propagation is used to obtain the uncertainty of the modelled ice dynamic estimates and modelled surface elevation changes.** The uncertainty estimated for the modelled surface elevation trend varies between near zero and ~6 cm yr^{-1} across the interior and along large parts of the ice sheet margins, and up to 12 cm yr^{-1} for the two locations with high SMB rates. **The uncertainty of the monthly GRACE solutions are derived following the method of Wahr et al. (2006) and are around 8 mm w.e. yr^{-1} (Fig. 7a), reducing towards the Polar Regions due to denser ground track coverage (Wahr et al., 2006). The uncertainties of the ICESat observations are below 0.05 m yr^{-1} in the interior, where a dense network of ground-tracks exists, and between 0.15 and 0.3 m yr^{-1} along the ice sheet margins due to greater distances between the ground-tracks and steeper slopes along the margins (Hoffmann, 2016) (Fig. 7b).**

For both RACMO2 models the overall uncertainty is given as 8% for the grounded ice sheet (Lenaerts et al., 2012; Van Wessem et al., 2014), resulting in an estimated uncertainty of less than 1 cm yr^{-1} in the interior and up to 6 cm yr^{-1} across the high SMB locations proposed in Enderby Land. The firn compaction model contains several error sources. In general, the complex physics of firn densification is still not fully understood, and the density of snow and firn is not well known, introducing large uncertainties into the computations (Sutterley et al., 2014). Error sources include the parameterisations to estimate surface snow density (Eq. 3) and the densification rate (Eq. 4), together with uncertainties within the forcing climate model RACMO2. As the firn compaction model is tuned to fit observations it is difficult to obtain realistic uncertainty estimates. However, following the idea of Helsen et al. (2008) we obtain our error estimate for the firn compaction model by assessing the propagation of the major error sources that affect firn compaction rates. This was done by applying a bias to the accumulation (8%) and temperature (10 K (Reijmer et al., 2005; Maris et al., 2012)), as well as to the surface snow density (±20 kg m^{-3} (Helsen et al., 2008)). The propagation of the errors is calculated to obtain the total uncertainty of the firn compaction model (Fig. 7c). Across most of the study site the uncertainty is estimated to be around ±2-3 cm yr^{-1}. However, at the two locations with the high SMB rates the

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uncertainty is significantly larger and is estimated to be up to 8 cm yr^{-1}. Uncertainties for GIA models are not provided, as the models are tuned to fit observations and the best-fitting ice sheet history and earth rheology values (e.g. Velicogna and Wahr, 2006). However, uncertainties within our study region are small due to small uplift rates and differences between the models of < 2mm yr^{-1}. Therefore, the error in the modeled GIA signals in our study region are likely to be small.
To estimate the uncertainty of the modelled ice dynamic and modelled surface elevation change, the propagation of errors of the particular error source is obtained (Fig. 7d and 7e). Depending on the incorporated satellite mission the uncertainty for the modelled rate of change due to ice dynamic is up to 6 cm yr^{-1} (GRACE, Fig. 7d) and up to 30 cm yr^{-1} (ICESat, Fig. 7e), due to the larger error of the ICESat observations. The uncertainty of the modelled elevation change is 0-12 cm yr^{-1} (Fig. 7f), with the greatest error source being the firn compaction model.

5 Conclusion

The rate of change due to ice dynamic**s** can be estimated independently from GRACE and satellite altimetry observations through the removal of GIA signals, SMB and, in the case of altimetry, firn compaction signals. Both approaches depend upon a separate SMB model, although in different ways since SMB causes a mass change in GRACE observations but a height change in altimetry observations. Therefore, any errors in the modelled SMB lead to differences in the ice dynamic **estimates** derived from GRACE versus altimetry. Thus, this approach provides a new and independent means of assessing the accuracy of SMB models. We showed that the differences between the old and new RACMO2 versions yield significantly different ice dynamic estimates, with RACMO2.3 producing smaller differences between the GRACE- and ICESat-derived estimates.
Although different GIA models affect GRACE and altimetry observations in different ways, changes in GIA models have a small effect on **the estimated rate of change due to** ice dynamic and so are not responsible for different estimates using the two satellite techniques. **Our data suggests** that the differences are not based on errors in the ICESat observations as most of the greatest differences occur in regions where ICESat uncertainties are low **(Fig. 7c)**, in particular the **large, negative difference**

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occurring inland within the study region (significantly different from zero at the 95% confidence level). Moreover, modelling the rate of change of surface elevation based on ice dynamic **estimates** obtained from GRACE observations and RACMO2.3 estimates positive and negative changes in elevation in the same regions as ICESat detects corresponding trends, though the rates appear slightly under-estimated compared to the altimetry observations. Therefore, it appears that the dominant driver in the differences of the modelled **rate of change due to** ice dynamic and surface elevation trends are the changes of the SMB rates within the RACMO2 model, with RACMO2.3 providing a more accurately modelled rate of change of surface elevation. Thus**, a comparison of estimated changes in ice dynamics** derived from GRACE and altimetry observations not only provides information about **dynamic mass changes,** but **may** also help to identify **regions** where models **fail to accurately simulate variations in SMB**.

Acknowledgement

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We would like to thank Stefan Ligtenberg for providing us with the RACMO2/ANT datasets and his extensive support and constructive help to establish our own firn compaction model.

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Appendix A. Supplement to data set and models
A1. Data sets and models
A1.1 GRACE
We use the monthly gravity field solutions CNES/GRGS RL03-v3, provided by the Groupe de Researches de Géodésie Spatiale (GRGS). The RL03 solutions have a spatial resolution of degree and order 80 (Lemoine et al., 2013) and have been chosen due to the stabilisation process that is applied to reduce noise in form of North-South striping. This is achieved by regularising the inversion for spherical harmonic coefficients (Bruinsma et al., 2010).
Temporary changes in the Earth's gravity field can be related to changes in surface mass due to the distribution of mass, as well as the elastic and viscoelastic (GIA) response of the lithosphere, the instantaneous and long term signal to changes in surface load (Wahr et al., 1998). We obtain mass anomalies by applying the equations that relate mass changes to gravity changes (Wahr et al., 1998) to obtain the change in mass due to SMB:

$$U^{w.e.}(\theta,\lambda,t) = R\sum_{n=2}^{N}\sum_{m=0}^{n} P_{nm}(\cos\theta)\frac{2}{1}$$

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and due to the viscoelastic deformation, or GIA:
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Brenner, A. C., DiMarzio, J. P. and Zwally, H. J.: Precision and Accuracy of Satellite Radar and Laser Altimeter Data Over the Continental Ice Sheets, Geoscience and Remote Sensing, IEEE Transactions on, 45(2), 321-331, 2007

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SMB (kg m^{-2} yr^{-1})	alpha	beta
<100	1.00	1.00
100-300	0.96	0.97
300-500	0.93	0.94
500-700	0.92	0.93
700-1000	0.90	0.86
1000-2500	0.88	0.86
2500-4000	0.87	0.84
>4000	0.87	0.54

Table A1:

Proposed values for the accumulation constants α and β used in our monthly firn compaction model. The constants are dependent on the accumulation rate and have been adapted to a best-fit.

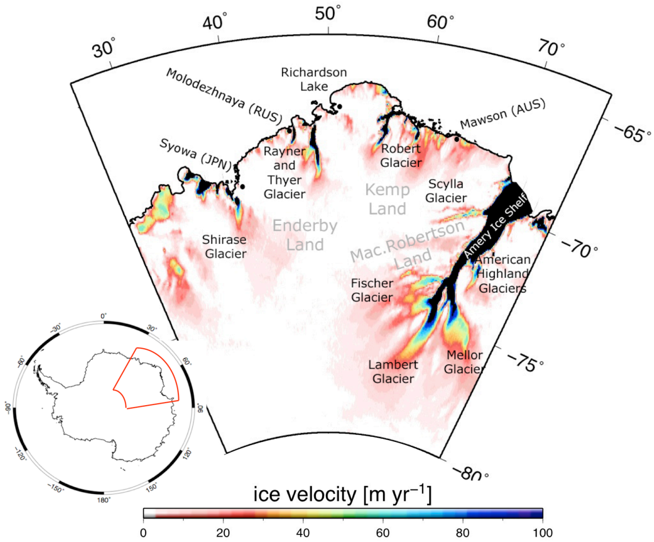

Figure 1: Regional map of our study area including Enderby Land, Kemp Land and Mac.Robertson Land. The map includes the locations of permanent research stations and major outlet glaciers. Ice velocity rates are plotted, sourced from the NASA MEaSUREs program (Rignot et al., 2011; Mouginot et al., 2012), to identify glaciers and regions with dynamic ice loss.

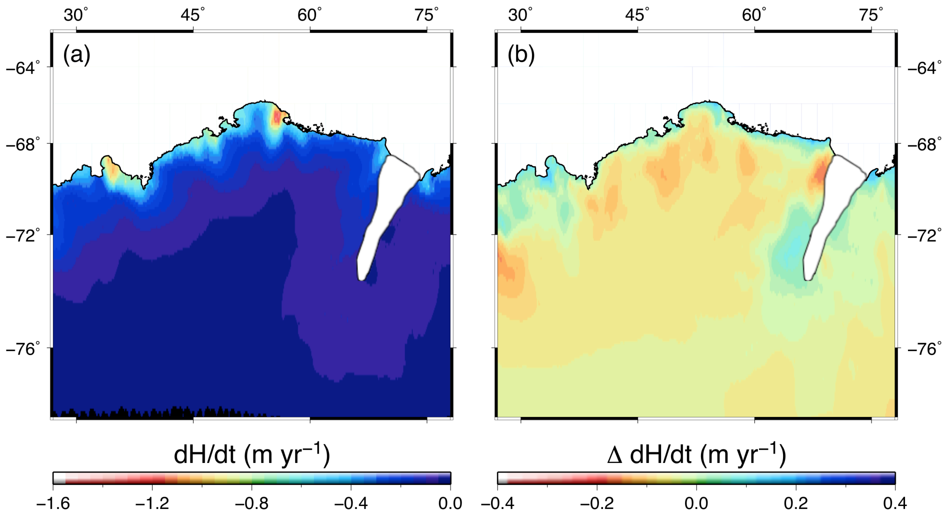

Figure 2: (a) Average annual vertical velocity rates due to firn compaction across the study site as obtained from our monthly firn compaction model, and (b) the differences between our model results and the firn densification model of Ligtenberg et al. (2011).

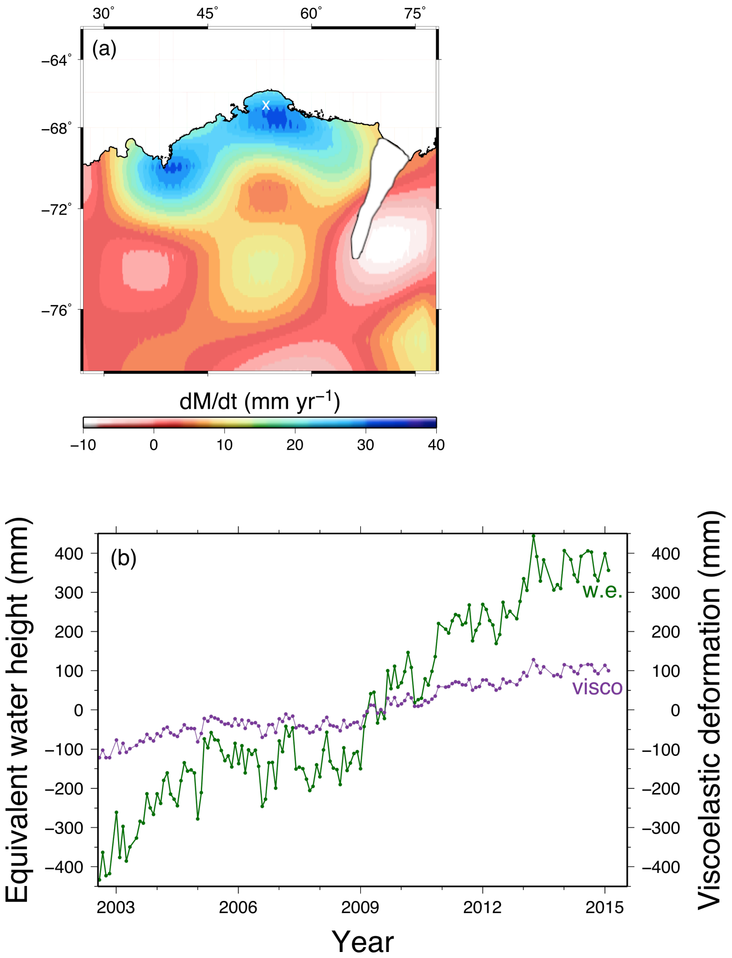

Figure 3: (a) Trend of the observed mass anomalies in Enderby Land monitored by GRACE over the time span of 2003-2009, uncorrected for GIA. The white cross illustrates the location of Richardson Lake, a former GPS station. (b) The time series shows a change in gravity at a chosen location in Enderby Land (67S 54E) over the total observational period. The green line illustrates the change assuming the gravitational change is caused by a surface mass load and is expressed in water equivalent (w.e.) (Eq.1), the purple line illustrates a change due to viscoelastic deformation (GIA) (Eq.2).

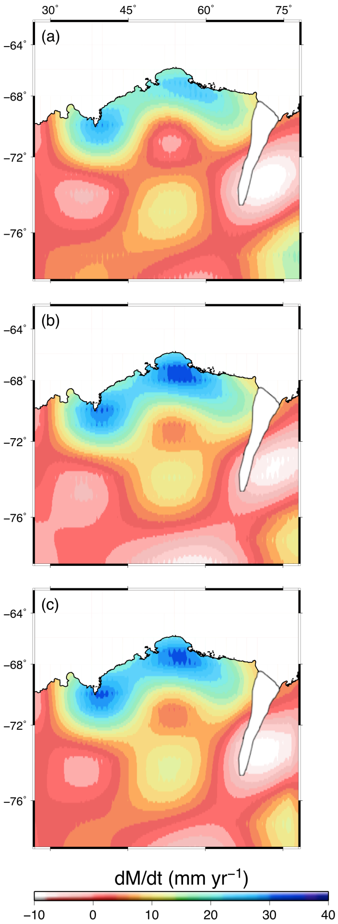

Figure 4: GRACE observations corrected for GIA uplift rates using (a) the ICE-6G_C(VM5) model by Peltier et al (2015), (b) the W12a model by Whitehouse et al. (2012), and (c) the ICE6G_ANU model by Purcell et al. (2016).

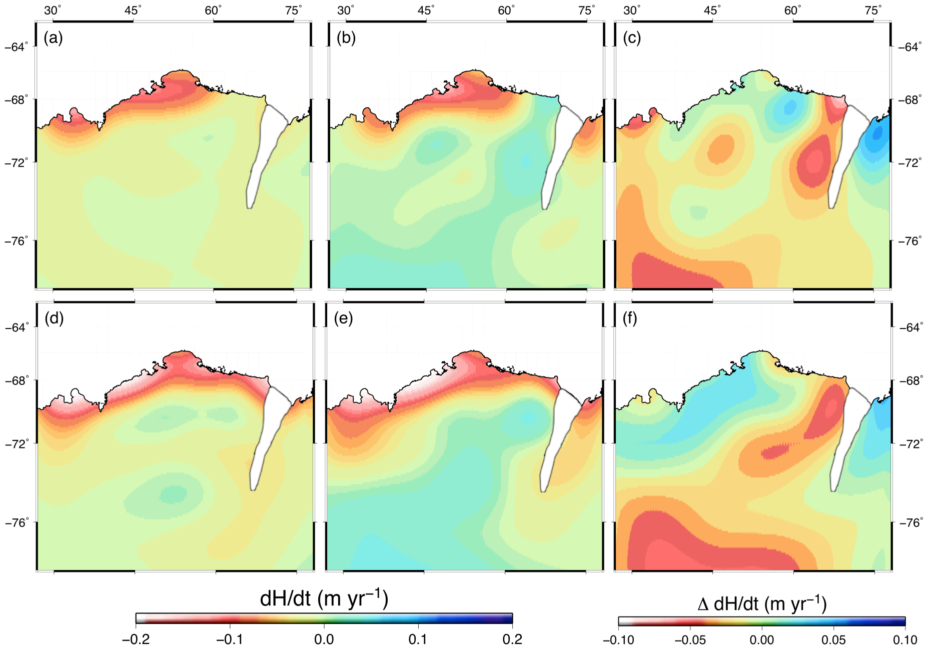

Figure 5: Comparison between the modelled ice dynamic rates obtained by employing SMB estimates from RACMO2.3 using (a) GRACE and (b) ICESat, and by employing SMB estimates from RACMO2.1 using (d) GRACE and (e) ICESat. (c) and (f) show the difference between ice dynamic rates obtained from GRACE minus ice dynamic rates obtained from ICESat for the employed SMB estimates obtained from RACMO2.3/ANT and RACMO2.1/ANT, respectively.

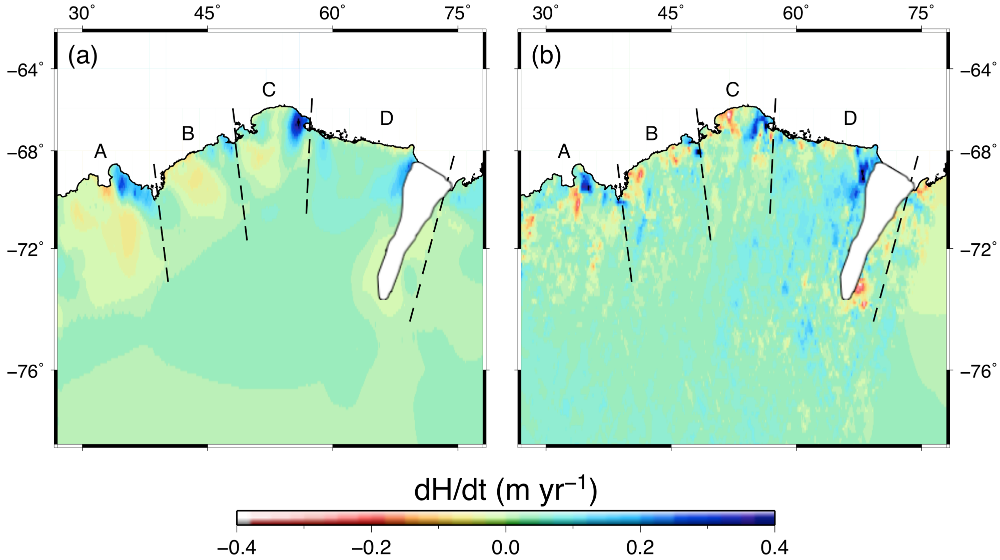

Figure 6: (a) our modelled rate of change of surface elevation retrieved by removing our estimated ice dynamic rates, obtained from GRACE, from the modelled trend in surface elevation (SMB-firn compaction) using RACMO2.3, compared to (b) the ICESat observations.

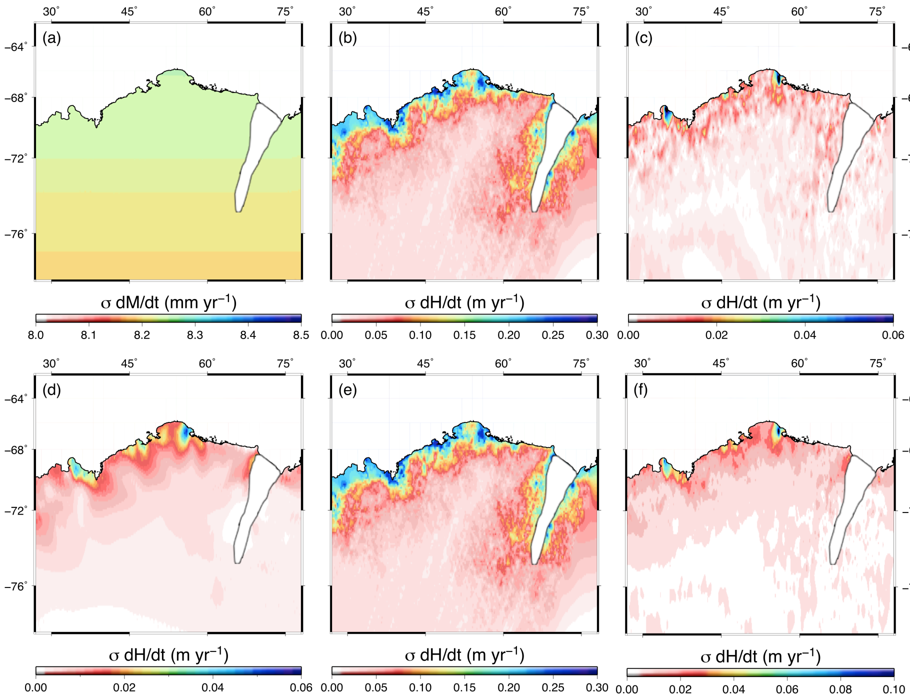

Figure 7: Uncertainties estimated for (a) GRACE, (b) ICESat, (c) our monthly firn compaction model, ice dynamic rates using RACMO2.3 obtained from (d) GRACE, (e) ICESat, and the modelled surface elevation trend for (f) RACMO2.3. The greatest uncertainty comes from the ICESat measurements, with up to 30 cm yr^{-1} at the margins, this results in greater uncertainties for the modelled ice dynamic rates obtained from the ICESat observations.

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) for our study period, and locally for a chosen

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(Fig. 2b)

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revealing a positive trend in SMB

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with a positive anomaly of ~32 ± 8 mm w.e. yr^{-1} near 40°E and 55°E, showing a significant

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rate estimates

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For both RACMO2 models the ice dynamic estimates are of somewhat similar rate for the two estimates obtained from GRACE and ICESat, with the greatest ice dynamic rates obtained between 30°E and 50°E.

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Appendix A. Supplement to data set and models

A1. Data sets and models

A1.1 GRACE

We use the monthly gravity field solutions CNES/GRGS RL03-v3, provided by the Groupe de Researches de Géodésie Spatiale (GRGS). The RL03 solutions have a spatial resolution of degree and order 80 (Lemoine et al., 2013) and have been chosen due to the stabilisation process that is applied to reduce noise in form of North-South striping. This is achieved by regularising the inversion for spherical harmonic coefficients (Bruinsma et al., 2010).
Temporary changes in the Earth's gravity field can be related to changes in surface mass due to the distribution of mass, as well as the elastic and viscoelastic (GIA) response of the lithosphere, the instantaneous and long term signal to changes in

surface load (Wahr et al., 1998). We obtain mass anomalies by applying the equations that relate mass changes to gravity changes (Wahr et al., 1998) to obtain the change in mass due to SMB:

$$U^{w.e.}(\theta,\lambda,t) = R\sum_{n=2}^{N}\sum_{m=0}^{n} P_{nm}(\cos\theta)\frac{2n+1}{1+k_n^{elast}}\left(\Delta C_{nm}(t)\cos m\lambda + \Delta S_{nm}(t)\sin m\lambda\right) \tag{A1}$$

and due to the viscoelastic deformation, or GIA:

$$U^{visco}(\theta,\lambda,t) = R\sum_{n=2}^{N}\sum_{m=0}^{n} P_{nm}(\cos\theta)\frac{h_n^{visco}}{k_n^{visco}}\left(\Delta C_{nm}(t)\cos m\lambda + \Delta S_{nm}(t)\sin m\lambda\right) \tag{A2}$$

where R is the Earth's radius, P_{nm} are the fully normalised Legendre functions, *n* and *m* are degree and order of the spherical harmonic coefficients, θ and λ are colatitude and longitude, and ΔC_{nm}, ΔS_{nm} are the spherical harmonic coefficients, at time *t*, of the GRACE anomaly fields. k_n and h_n are the elastic Love loading numbers (e.g. Pagiatakis, 1990) and the ratio of viscoelastic Love loading numbers (Purcell et al., 2011), depending on the degree.

A1.2. ICESat

Various methods are used to estimate surface elevation changes from ICESat observations, using either along-track measurements or measurements directly taken from the crossover location. Due to perturbations in the orbit, deviations of the repeated ground track occur and it is necessary to determine the surface topography to correct elevation changes due to surface slope rather than changes in ice mass. Different methods have been applied to obtain surface elevation changes, using either along-track observations or crossover measurements (e.g. Slobbe et al., 2008; Gunter et al., 2009; Pritchard et al., 2009; Sørensen et al., 2011; Ewert et al., 2012).
Here we use the estimated rate of change of ice sheet elevation obtained from a newly developed technique that combines both crossover and along-track observations (Hoffmann, 2016). The method allows estimation of the local surface slope using a digital elevation model that has been derived from gridded estimates of ice height at ICESat crossover points. Over a crossover grid, that geographically spans all campaign crossovers of a location, a static grid was created on which heights were interpolated at the epochs of all campaigns. The estimate of the elevation change over time is made by computing a weighted least-square regression of the height time series of each grid node and then computing a weighted mean value for all grid nodes to derive the crossover height rate. This not only allows to assess height rates at one location over time, but also to evaluate a digital elevation model directly from the data, which is used to estimate the slope at crossovers (Hoffman, 2016).

The slope estimates at the crossovers are then interpolated to remove the surface slope from the along-track measurements. Although the elevation change estimates from along-track measurements are naturally less precise than the rate estimates at crossovers, combining both methods significantly increases the accuracy of the slope correction, providing a measure to validate along-track estimates (Hoffman, 2016).

A1.3. RACMO2/ANT

The RACMO2/ANT regional climate model, used to obtain SMB estimates, adopts the dynamical processes from the High Resolution Limited Area Model (HIRLAM) and the physical atmospheric processes from the European Centre for Medium-range Weather Forecasts (ECMWF) (Reijmer et al., 2005) and is forced by ERA-Interim reanalysis data at the lateral boundaries (e.g. Ligtenberg et al., 2011; Lenaerts and van den Broeke, 2012). The latest version, RACMO2.3 (Van Wessem et al., 2014), extends available model data from 1979-2012 (RACMO2.1) to 1979-2015 (RACMO2.3) and improves the temporal resolution from 6-hourly (RACMO2.1) to 3-hourly (RACMO2.3) (Ligtenberg per. comm., 2016). The horizontal resolution is 27 km and the vertical resolution 40 levels. Individual SMB components are provided including snowfall, evaporation/sublimation, and snowmelt, as well as snowdrift in RACMO2.3. Over Antarctica RACMO2/ANT is coupled with a multilayer snow model, which estimates meltwater percolation, refreezing and runoff, as well as surface albedo and snowdrift (Van Wessem et al., 2014). The update in the physical parameters of RACMO2.3 results in a general increase in precipitation over the grounded East Antarctic Ice Sheet, evaluated using in-situ observations, ice-balance velocities and GRACE measurements and showing a general improvement of the SMB (Van Wessem et al., 2014).

A1.4. Firn compaction

We developed a firn compaction model based on the firn densification model of Ligtenberg e al. (2011), using near surface climate provided by RACMO2.1. It is a one-dimensional, time-dependent model that estimates density and temperature individually for each layer and at each time step in a vertical firn column. The firn densification model of Ligtenberg et al. (2011) adds new snowfall instantly to the current top layer until the layer thickness exceeds ~15 cm (Ligtenberg, pers. comm., 2016), at which time it is divided in two layers. The properties of each layer are passed on to both layers. If a layer becomes too thin, due to compaction or surface melt, the layer is merged with the next layer and assigned the average properties of both layers. Our model has been simplified to improve the computational time. Rather than adding new snowfall instantly to the top layer, we compute the monthly sum of SMB and use the monthly averaged surface temperature to estimate the densification rate, density and new temperature to obtain the vertical velocity of the surface due to monthly firn compaction.

The model starts with a new firn layer created by the total SMB of one month and is built up by adding a new layer each month using monthly SMB values and mean surface temperatures. The surface snow density of each top layer is estimated (Eq. 3) and the densification rate is obtained using a dry snow densification expression proposed by Arthern et al. (2010):

$$\frac{d\rho}{dt} = CAg(\rho_i - \rho)e^{\left(\frac{-E_c}{RT} + \frac{-E_g}{RT_{av}}\right)}$$
(A3)

where C is the grain-growth constant (m s^2 kg^{-1}), independently calculated for densities below (C = 0.07) and above (C = 0.03) the critical density of 550 kg m^{-3}, A is the accumulation rate (mm w.e. yr^{-1}), g the gravitational acceleration, and ρ and ρ_i are the local density and the ice density (kg m^{-3}), respectively. The exponential term includes the activation energy constants (kJ mol^{-1}) for creep and for grain-growth, E_c and E_g, respectively, the gas constant R (J mol^{-1} K^{-1}) and the local temperature T, and annual average temperature T_{av} (K).
The process of liquid water percolation and refreezing is incorporated as a function of snow porosity P_s and density, as proposed by Coléou and Lesaffre (1998) (Ligtenberg et al., 2011; Kuipers Munneke et al., 2015):

$$L_W = 1.7 + 5.7\left(\frac{P_S}{1 - P_S}\right)$$
(A4)

with the snow porosity:

$$P_S = 1 - \left(\frac{\rho}{\rho_i}\right)$$
(A5)

where ρ is the density of the layer and ρ_i the density of glacier ice.
The heat transport throughout the firn column is solved explicitly using the one-dimensional heat-transfer equation (Cuffey and Paterson, 2010):

$$\frac{dT}{dt} = \kappa \frac{d^2T}{dz^2}$$
(A6)

with the thermal diffusivity κ and the depth *z*. Initially the heat-transfer equation consists of a term for heat conduction, advection and internal heating. However, initial heating is small within the firn layer and therefore neglected and the contribution of heat advection is taken into account by the downward motion of the ice flow (Cuffey and Paterson, 2010; Ligtenberg et al., 2011).
Finally, once the densification rate is estimated, the vertical velocity of the surface due to firn compaction, V_{fc}, can be assessed by integrating over the displacement of

the compacted firn layers over the length of the firn column (Helsen et al., 2008):

(A7)

where z is depth, ρ density and dρ(z)/dt the densification rate.
Ligtenberg et al. (2011) found that Equation (A1) over-predicts the rate of densification for most regions in Antarctica, with the effect of the annual average accumulation being too large on the densification rate. They reintroduced an accumulation constant that previously had been proposed by Herron and Langway (1980) as α in A^{α} (below 550 kg m^{-3}) and β in A^{β} (above 550 kg m^{-3}), initially chosen between 0.5 and 1.1 but later assumed to be α, β =1 (Zwally and Li, 2002; Helsen et al., 2008). Ligtenberg et al. (2011) applied a modelled to observed ratio to correct for the accumulation dependence. We also found that Equation (A1) over-predicts the rate of densification, depending on the rate of the average annual accumulation. However, due to our use of monthly layers, the ratio proposed by Ligtenberg et al. (2011) is no longer valid and we introduce new α and β, depending on the accumulation rate (Table A1). The values for α and β represent a best fit and have been obtained by investigating different values across several model runs. This means that the firn compaction model is adjusted to fit available observations and is therefore assumed to be correct and invariant of SMB model changes.

In Figure A1a we show the average annual rate of firn compaction across the study site and in Figure A1b the differences between our model and the model of Ligtenberg et al. (2011). Along the ice sheet margins and the Amery Ice Sheet our model overestimates their firn compaction rates by 5-10 cm yr^{-1}, while it underestimates rates by 7-12 cm yr^{-1} in most other areas further inland, with up to 15 cm yr^{-1} at two individual location near 28°E and between 68°E and 70°E. These differences are within our estimated uncertainty, based on the uncertainties provided for the modelled SMB from RACMO2 (Appendix A3).