

Dear Reviewer 2,

Thank you very much for your constructive suggestions on how to improve our manuscript.

We have now addressed all your concerns and modified the manuscript accordingly. Please find below our point-by-point reply; the labels in brackets refer to the position in the text.

We also very much appreciated you commented printouts attached as a pdf supplement, which made it very easy for us to include your minor corrections.

Please let us know if you have further comments that need resolving.

Sincerely,

Ingo Sasgen & the REGINA team

This paper gives a detailed description of a group of linked geodetic datasets for Antarctica: ice surface height change from altimetry, gravity change from GRACE, and uplift rates from GPS. In addition, a set of viscoelastic response functions are provided, which can be used for joint inversion or other modeling of these data. The paper is comprehensive and thorough, and with a few minor points of confusion, gives a complete description of these datasets. The datasets themselves are useful, and although some version of each has been published before, there is good reason for people to use these latest and more complete versions. The authors do an excellent job in description and complete citation, and are to be commended for including the DOIs for the GPS sites that they use.

The paper seems to indicate that the code (in some cases) is also provided (and function names and so on are provided). But I do not see the code in the repository. I encourage the authors to include the code with the data. There was also mention in the paper of some other products such as the geodesic grid, which at present are not included in the repository. I see the value in including this, so the authors need to carefully check the repository for completeness. There also are some intermediate (calibration) data products that might be worth considering to add to the repository, in particular the time series for the firm correction and the SMB (as de-trended for this study).

The data, code and ancillary data were already available in the repository, however, hidden under the link “View data as html”. Thanks for pointing this out. We have now added links in each section of the data set directly to the respective zip-archive (links below). In addition, we explained in the conclusion to click on “View data as html”. [R2_010]

Altimetry: http://hs.pangaea.de/model/Sasgen-etal_2017/Ice_sheet_topographic_change.zip

Gravimetry: http://hs.pangaea.de/model/Sasgen-etal_2017/Geoid-height_change_from_GRACE_satellite.zip

GPS: http://hs.pangaea.de/model/Sasgen-etal_2017/In_situ_GPS_uplift_rates.zip

VE response functions: http://hs.pangaea.de/model/Sasgen-etal_2017/Viscoelastic_response_functions.zip

We marked a number of minor corrections on an annotated manuscript. Only the pages with corrections are included.

Corrections based on Supplementary files:
[R2_036] Re-arranged first sentence of Abstract

[R2_066] Included suggested corrections in Introduction
We included all minor corrections stated in the supplementary file

Introduction: Explain why the 2003-2009 time span was chosen for GRACE. It matches the altimetry, but a longer time span was used for GNSS. Also, emphasize more strongly the key assumption of a constant rate of change of all variables over the time period chosen, especially where datasets do not exactly coincide in time.

We added a paragraph emphasizing the assumption of the stationarity of the trend and explaining the time period chosen. Please note that GPS uplift rates for 2003-2009 and 2003-2013.7 are available in the data archive, for comparison [R2_020]

Line 158. Units mismatch: “rate” and “cm”. cm/year?

[R2:169] Corrected.

Section 2.3. Why not use a weighted average of the two datasets? I can guess that maybe when one of the datasets has a large uncertainty then it can be re-ally bad?

Yes, very large uncertainties may point to an unknown systematic bias, or a poor description by the fitting-plane model, which would degrade, by spatial averaging, the combined elevation rate [R2_181].

Could you either present the values or statistics of $(\text{data_type_chosen} - \text{data_type_not_chosen})/(\sqrt{\text{sum of sigmas}^2})$? The reader wonders how well the two altimetry sources agree.

This is a good idea. However, please understand that the work package associated with the altimetry processing has been closed for a while (2016). Therefore, we would like to divert the reader to the data sets in the Pangaea repository for doing these calculations independently, if necessary.

Section 2.4 Explain how the 27 km grid of firn corrections is re-sampled to the 10x10 km grid used for the altimetry. Standard bilinear interpolation, or something else? Also, line 213 refers to a 20 km grid. What is the relationship between this grid and the 10x10 grid? Are the 10x10 grid values decimated to 20 km? Block mean, block median? Or is the reference to 20 km an error in the text?

The interpolation scheme for ICESat and Envisat is bi-linear to the 10 km grid. For Envisat, we use linear interpolation to provide the 20 km in the archive [R2_172]. This is now specified in the text [R2_156]

Section 3. Why not just use JPL’s orbits and clocks? A short explanation is enough. I presume that the reason relates to ensuring consistency.

Orbits and clocks – at the time of set up and processing in late 2013 and early 2014, JPL’s repro 2.1 was not available so to include second order ionosphere and earth radiation pressure models, where orbit consistency is critical, we had to generate our own orbits. We also used VMF1 rather than JPL’s choice of GPT2 for repro2.1.

The sentence was added: “(at the time of processing JPL reprocessed orbits for these state-of-the-art options were not available)” [R2_279]

Line 281. Give the median values for white noise sigma scaling and the flicker noise amplitude.

The median values estimated for variable white noise scale factor and flicker noise amplitude were 1.6 and 13.4 mm respectively. Values added to text [R2_295]

In all the tables in Section 3, please clarify if the Argus et al. (2014) result are as originally presented, or if you attempted to correct for the small frame difference.

The rates are as given in the original paper, i.e. they are uncorrected. We have added the sentence “All rates from Argus et al. (2014) in Tables 2, 3 and 4 are shown as given in the original paper.” [R2_327]

Lines 617-618. This sentence says that the loading rate is held constant until dynamic equilibrium is reached, which as written suggests that the loading rate is changed after that time. How is it changed? Or is this simply an error in English? Based on other parts of the text, it seems that the loading rate was actually held constant for 2000 years (West Antarctica) or 15,000 years (East Antarctica), and that after this an extra time step is done with no loading change to give the purely viscoelastic response. In fact, the 2000 years and 15,000 years are argued to be longer times than needed to reach dynamic equilibrium, if I understood the authors properly.

Yes, it is correct: after reaching the equilibrium state, two extra time steps without a load change are done, yielding the purely viscoelastic response in terms of uplift rates and gravity rates. We included explanatory sentences [R2_658], also in the new figure in Appendix A.8

Line 658. Clarify whether you mean “to a value of $1/e$ ” (as written), or “by a factor of $1/e$ relative to the initial value”.

Changed. Also, it is $1/e^2$ (corrected).

For figure 7-11, it would be useful to complement the versions shown with a figure that shows the uplift rate as a function of time at the center of the load.

Fig. A.5 displays the ratio of geoid/surface displacement at load center, which is not exactly what you recommended, but most important for the joint inversion addressed in Paper II. We would like to recommend to the reader to download the viscoelastic response functions, which allow to display the surface displacement rate $u(t)$ and geoid rate at the load center (or other locations) over time.

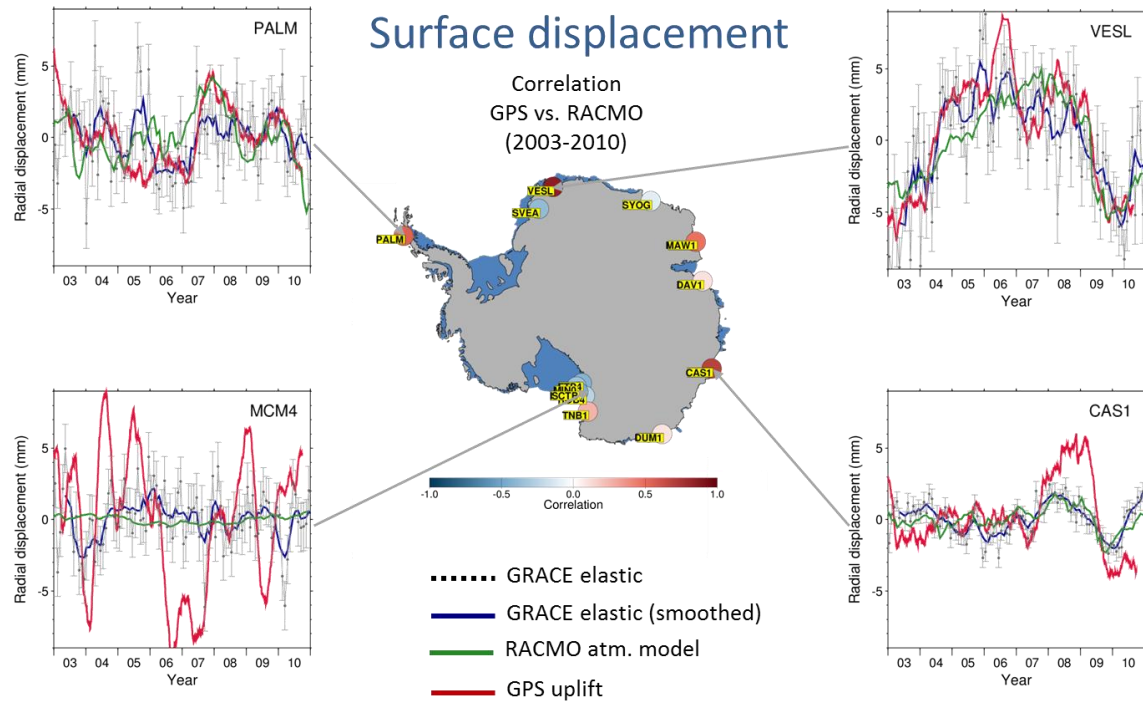
If there is any change in the loading rate during the calculation (see the comment on the ambiguity introduced by lines 617-618), then the loading rate should also be shown.

We think that the issue is clearer with the new explanatory text added [R2_658]. We have also modified the legend of the Fig. 7. And we have added a plot on the load evolution in Appendix A.7 [R2_1014]

SMB correction. Because the SMB correction for the GRACE has been detrended over the GRACE time period, it does not contribute to the rate. However, the SMB variations might contribute to the GPS uplift rates given that the time spans of these data vary. I wonder if this effect has been considered and whether it might be important? Sites with strongly non-linear time series were identified and excluded or handled separately, but there may be some subtle biases that remain. I think it would be useful to evaluate this impact, or at least to provide the de-trended SMB grid time series so that it could be assessed.

[R2_867] In theory, remaining trends in the data sets due to (unknown) SMB fluctuations should be separated and removed by the joint inversion. However, as you have stated above, inconsistencies arise when different time spans are used for the different data sets. This concerns not only surface-mass changes, but co- and post-seismic deformation at single stations (more in Sasgen et al. 2017, GJI). As we have point out, spatial coverage of the GPS data was preferred over consistency with

the other data sets. However, in an earlier study phase, we have assessed the SMB variability and the associated apparent trend for the GPS stations, which is now included in the paper as Appendix A.7. We have also compared GPS recorded deformations to predictions with GRACE and RACMO2 (Figure below). For some stations, there is a good agreement (e.g. VESL and PALM), but it is also visible that a lot of the interannual variations could have a different source, likely instrumental uncertainties (snow on or in the antenna, e.g. MCM4, secondary reflection of the signals, etc.). Therefore, we decided to place our focus on manually evaluating the GPS time series for robustness.



**Altimetry, gravimetry, GPS and viscoelastic modelling data for the joint inversion for
glacial isostatic adjustment in Antarctica (ESA STSE Project REGINA)**

3

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Keywords:

Global change from geodesy, Gravity anomalies and Earth structure, Loading of the Earth,
33 Glaciology, Antarctica, Joint inversion

ABSTRACT

[R2_036] The poorly known correction for the ongoing deformation of the solid Earth caused by glacial isostatic adjustment (GIA), is a major uncertainty in determining the mass balance of the Antarctic ice sheet from measurements of satellite gravimetry, and to a lesser extent satellite altimetry, ~~is the poorly known correction for the ongoing deformation of the solid Earth caused by glacial isostatic adjustment (GIA).~~ In the past decade, much progress has been made in consistently modelling the ice sheet and solid Earth interactions; however, forward-modelling solutions of GIA in Antarctica remain uncertain due to the sparsity of constraints on the ice sheet evolution, as well as the Earth's rheological properties. An alternative approach towards estimating GIA is the joint inversion of multiple satellite data – namely, satellite gravimetry, satellite altimetry and GPS, which reflect, with different sensitivities, trends of recent glacial changes and GIA. Crucial to the success of this approach is the accuracy of the space-geodetic data sets. Here, we present reprocessed rates of surface-ice elevation change (Envisat/ICESat; 2003-2009), gravity field change (GRACE; 2003-2009) and bedrock uplift (GPS; 1995-2013.7). The data analysis is complemented by the forward-modelling of viscoelastic response functions to disc load forcing, allowing us to relate GIA-induced surface displacements with gravity changes for different rheological parameters of the solid Earth. The data and modelling results presented here are available in the Pangaea; <https://doi.pangaea.de/10.1594/PANGAEA.875745>. The data sets are the input streams for the joint inversion estimate of present-day ice-mass change and GIA, focusing on Antarctica. However, the methods, code and data provided in this paper are applicable to solve other problems, such as volume balances of the Antarctic ice sheet, or to other geographical regions, in the case of the viscoelastic response functions. This paper presents the first of two

contributions summarizing the work carried out within a European Space Agency funded study,
REGINA.

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1. INTRODUCTION

[R2_066] Glacial isostatic adjustment (GIA), the viscoelastic deformation of the solid Earth in response to climate-driven ice and water mass redistribution on its surface, is poorly constrained in Antarctica. The primary reason is the sparseness of geological evidence ~~of~~for the past ice sheet geometry and local relative sea-level change. These are important constraints on the exerted glacial forcing and on the viscoelastic structure of the lithosphere and of the mantle, ~~respectively~~, which ~~concoctedly together~~ determine the signature of GIA (e.g. Peltier, 2004; Ivins and James 2005; Whitehouse et al. 2012; van der Wal et al., 2015). The predictions of GIA in Antarctica remain ambiguous (Shepherd et al. 2012, suppl.) and cause a large uncertainty in gravimetric mass balance estimates of the ice sheet of the order of the estimate itself (Martín-Español et al. 2016b). Measurements of bedrock uplift by GPS ~~have shown to be~~ are inconsistent with the predictions of existing GIA. [R1_076] In many regions, uplift rates and thus mass increase due to GIA is over-predicted (Bevis et al. 2009), biasing estimates of present-day Antarctic ice-mass loss from GRACE to more negative values. However, for regions with a weak Earth structure, large uplift signals are recorded by GPS (e.g. Groh et al. 2012), which are likely caused by load changes within the past few thousand year, and often not accurately represented in GIA predictions (Wolstencroft et al. 2015).

Much progress has been made in reconstructing the ice sheet evolution from geomorphological evidence (Bentley et al. 2014) and inferring the underlying Earth structure from seismic observations (An et al. 2015; Heeszel et al. 2016). However, an independent approach to constraining GIA is to make use of the different sensitivities of the various types of satellite data to recent glacial changes and GIA, respectively. ~~And thus to separate~~ Separating both signals in a joint inversion approach has been pursued by e.g. Wahr et al. 2000; Riva et al.

2009; Wu et al. 2010; Gunter et al. 2014, Martín-Español et al. 2016a. Another approach used regional patterns of GIA from forward modelling and adjusted them to GIA uplift rates in Antarctica (Sasgen et al. 2013).

In this paper, we present methods and data inputs in preparation of solving the joint inversion problem for GIA in Antarctica. As the GIA process is gradual, causing an approximately constant rate of change within a decade, we first process the satellite data to recover optimal temporal linear trends. [R2_020] We focus on the trends derived for the time period 2003-2009 in which GRACE and ICESat operated simultaneously. Note that the stationarity of the trend is a key assumption underlying our approach, when including GPS rates covering a longer time span (1995-2013.7). However, limiting the GPS data to the time span 2003-2009 leads to a significant reduction of the number of stations for which reliable trends can be estimated, and, hence, a loss of spatial coverage. For comparison, the reader is advised to the data archive, in which GPS uplift rates for the time periods 2003-2009 and 2003-2013.7 are made available.

[R1_099] In this paper, we present refined ~~We refine existing~~ procedures for estimating trends for of the data sets on surface-ice elevation changes, surface displacement and gravity field changes. The rates of surface-ice elevation changes from Envisat and ICESat satellite altimetry are improved by (Section 2), by combining both data sets based on their respective uncertainties, increasing the spatial coverage and accuracy of the elevation rates (Section 2).; ~~bedrock-Bedrock~~ displacement from in situ networks of GPS stations in Antarctica are improved in coverage by allowing for campaign-based data and carefully assessing the uncertainty of the trend with a noise model (Section 3). Compared to the rates in Thomas et al. (2011) also more stations and longer time series are included, and ~~The~~ gravity field changes

111 from GRACE are refined compared to previous work by optimizing the de-stripping filtering for
the region of Antarctica (Section 4). [R1_111] The processing aims at fulfilling the requirement
of the joint inversion to combine input data based on the same time period (not possible for
114 GPS without having to ignore a large number of stations) and covering entire Antarctica,
accompanied by a realistic description of the uncertainties.

We also present forward modelling results of viscoelastic response functions to disc load
117 forcing for the range of Earth structures likely to prevail in Antarctica (Section 5). [R1_105]
The viscoelastic response functions allow us to combine the surface displacement and gravity
changes based on the physical description of the Earth's viscoelastic response for a specified
120 Earth structure. In addition, the response functions enable us to combine data sets of different
spatial resolutions, as this is the case for GPS, GRACE and altimetry.

The determination of viscoelastic response functions is a classic topic in solid Earth
123 modelling (e.g. Peltier & Andrews, 1976), though uncommon in the application to joint ~~invasion~~
inversion studies of satellite data. Although this paper focusses on Antarctica, the response
functions and data processing techniques presented here are applicable to other regions. The
126 response kernels represent a wide range of Earth structures and can be used for the separation
of superimposed present-day (elastic) and past (viscoelastic) signatures of mass change in other
regions with a similar Earth, for example hydrological storage changes and GIA in North
129 America and Alaska. The response functions give insight into the temporal and spatial scales of
deformation expected for Antarctica, and are crucial when combining the input data streams.

The data sets and modelling results presented in this paper are accessible in the Pangeae
132 archive, <https://doi.pangaea.de/10.1594/PANGAEA.87574> – subsections provide user guidance
and point to data and code stored in the archive. As mentioned above, the data sets and

modelling results are of value to address other research questions as well. For example, the GPS
135 rates provided are useful for the validation of forward modelling GIA solutions, the GRACE
gravity rates can be used for mass balance studies, and altimetry data 2003-2009 can be
extended with the ongoing CryoSat-2 mission to infer volumetric mass balances, also over the
138 ice shelves. The viscoelastic response functions are based on Earth model parameters
potentially suitable to other geographical regions, as well; they are useful for similar studies
combining different data sets of geodetic observables, surface deformation, gravity field
141 change, and topographic change in glaciated areas.

The actual method of the joint inversion is described in a second contribution of the
REGINA project team (Sasgen et al. ~~submitted~~2017). In this second paper, the resulting GIA
144 estimate is also compared to previous studies. ~~The processing of the data issued here was
enabled by the European Space Agency within the CryoSat+ Support To Science Element Study
REGINA.~~

2. ALTIMETRY DATA ANALYSIS

ICESat elevation rate determination

We use along-track altimetry measurements from *ICESat 633 Level 2*, providing high-resolution elevation change observations for the period February 2003 until October 2009. Two corrections are applied to this data set: the range determination from Transmit-Pulse Reference-Point Selection (Centroid vs. Gaussian) (Borsa et al. 2014) available from the National Snow and Ice Data Center (NSIDC), and the inter-campaign correction (Hofton et al. 2013). The Centroid-Gaussian correction is a well-established correction and has been incorporated to the latest ICESat release (634). Concerning the ICESat Intercampaign Bias (ICB) correction, uncertainties are available at Hofton et al (2013). Furthermore, several studies have determined this correction from different methodologies. For a summary of published ICESat ICB corrections see Scambos & Shumman (2016). [R1_157] Note that these corrections are part of a widely accepted procedure and their effect on the elevation rates and uncertainties caused by varying the processing choices have not been evaluated. Because the ICESat tracks do not usually overlap, a regression approach is used in which topographic slope (both across-track and along-track) and the rate of surface-elevation change $\mathbf{y}_{\text{ICESat}}^h$, are simultaneously estimated using the ‘plane’ method (Howat et al. 2008⁺) over areas spanning 700 m long and few hundred meters wide. A regression is only performed if a plane has at least 10 points from four different tracks that span at least one year. Regression was carried out twice; first, individual elevation measurements with corresponding residuals outside the range of two standard deviations were detected, then, the regression was repeated omitting these outliers. The standard deviation of the regression coefficient, here taken as the uncertainty of the elevation rate, σ^h (here, ICESat) is calculated by the propagation of the [R1_168] residuals of the -uncertainties of the input data

and the estimated topographic heights,

$$\hat{s}_{\text{ICESat}} = \sqrt{\frac{\sum e_i^2 / (n-2)}{\sum (x_i - \bar{x})^2}}, \quad (1)$$

to the trend parameter (see Eq. 1 in Hurkmans et al. 2012), where $\mathbf{e}$ is the vector of residuals, n is the sample size ($i = 1, 2, \dots, n$), and $\mathbf{x}$ is the vector of input elevations with mean $\bar{x}$. This standard deviation (σ_{ICESat}) takes into account the sample size and the variance of both input data and residuals of the regression (Hurkmans et al. 2012). The residuals of the regression are used as they quantify the approximation of fitting the data with a plane. The exact ICESat observation periods are shown in the Appendix (A.1, Table A.1). Then, the elevation rate and its uncertainty are interpolated (bi-linear) [R2_156] to a common 10×10 km grid in polar-stereographic projection (central latitude 71°S ; central longitude 0°W , and origin at the South Pole, WGS-84 reference ellipsoid).

2.2

Envisat elevation rate determination

We use a time series of elevation changes derived from along-track Envisat radar altimetry data for the interval January 2003 to October 2009 (coeval to ICESat time span). Elevation rates y_{Envisat}^h are obtained at points every 1 km along track, by binning all the echoes within a 500 m radius. Then, a 10-parameter least squares model is fitted in order to correct for the across-track topography and changes in snowpack properties. The least square model is defined in Flament and Remy (2012). The estimated parameters include parameters determined for the backscatter, leading-edge width and tailing-edge slope, the mean altitude, quadratic surface slope parameters to define surface curvature and a linear time trend. A digital elevation model was not used for the correction of the topographic slope. For processing reasons, the temporal

192 resolution is re-sampled from 35 days to monthly periods for each grid cell, before estimating
the elevation rates. This has a minor effect on the elevation rate estimate ([R2_169] smaller
than ± 1 cm/yr) and reduces the standard deviation by about 14 %. As ~~for~~ with ICESat, the
195 elevation rate is interpolated to a common 10×10 km polar stereographic grid (and 20×20
km for download in the archive [R2_172]), and the standard deviations of the rates within each
grid cell are taken as an estimate of the measurement uncertainty, σ_{Envisat} according to Eq. 1.

198 *Combination of Envisat and ICESat*

2.3 We produce a combined rate of surface-elevation change product from the ICESat and
Envisat datasets for the Antarctic ice sheet, y^h . The aim is to take advantage of the high spatial
201 resolution of ICESat data and the high temporal resolution and high-track density of the Envisat
data.

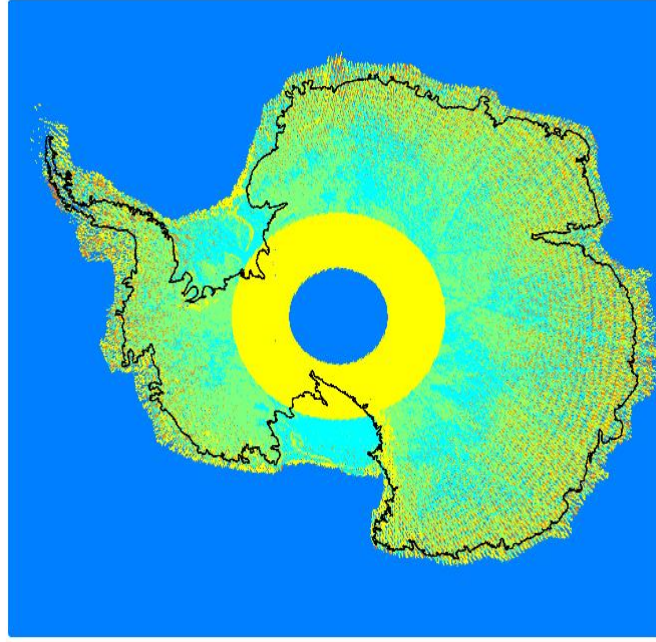


Figure 1. Mask for the combination of Envisat/ICESat. ICESat but not Envisat available (yellow),

$\sigma_{ICESat} \leq \sigma_{Envisat}$ (green), $\sigma_{ICESat} > \sigma_{Envisat}$ (turquoise), Envisat but not ICESat available (orange), and no data (blue). No interpolation is used.

We combine the two altimetry datasets based on their common 10×10 km polar-stereographic grid. At each location, the elevation rate with the smallest standard deviation is chosen from either Envisat or ICESat datasets. [\[R2_181\]](#) We prefer this masking procedure instead of a weighted average, in order to avoid introducing possible biases associated with gridded elevation rates of very high uncertainty.

Fig. 1 shows the resulting mask underlying the combination. It is evident that some grid points are only represented by either ICESat or Envisat. Most prominent is the narrowing of the polar gap with ICESat data, resulting from the 81.5°S latitude limit for Envisat compared to 86°S for ICESat due to satellite orbit inclination. On the Antarctic Peninsula, Envisat picks up some points that are not present due to a sparser track coverage in the ICESat data set. As expected, ICESat outperforms Envisat in terms of uncertainty of the elevation rate over steep

topographic slopes and along the ice sheet margins. This is due to the smaller footprint of the laser altimeter, its higher accuracy and lower slope-dependent uncertainty (e.g. Brenner 2007).

216 On some flat areas and over some faulty ground tracks, where ICESat data measurements are scarce, however, Envisat provides better temporal and spatial coverage leading to better accuracy of the resulting elevation rates. The resulting combined data set of surface-elevation
219 rates and its uncertainties are shown in Fig. 2.

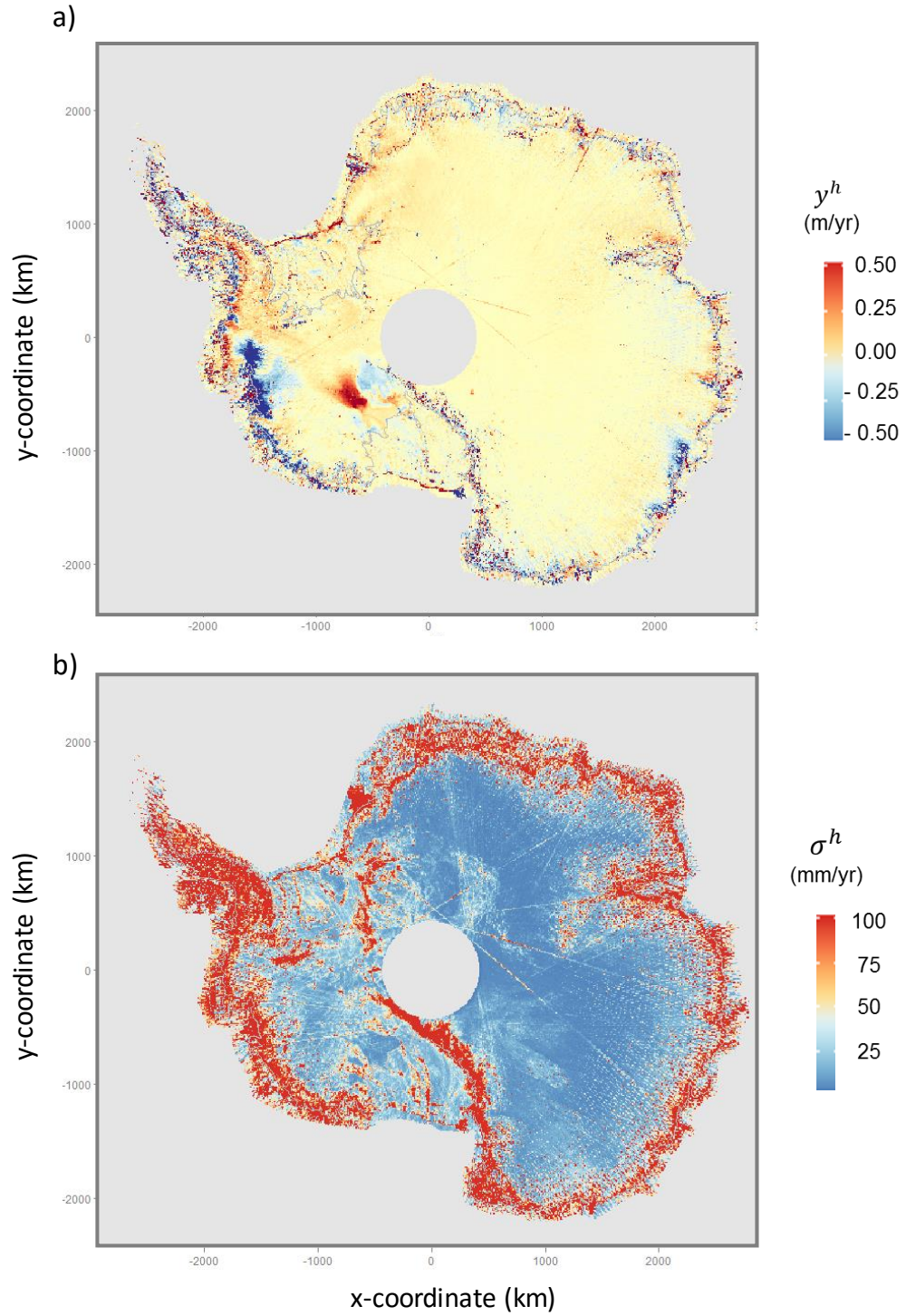


Figure 2: a) Rate of surface-ice elevation change y_h^h and b) associated uncertainties σ_h^h derived from Envisat/ICESat combined dataset for the time interval 2003-2009. No interpolation is used; grid points without values are empty (shaded grey).

Firn correction

222 The elevation rates derived from ICESat and Envisat are corrected for changes in the firn layer thickness using the firn compaction model of Ligtenberg (2011), which is driven by the regional atmosphere and climate model RACMO2/ANT (Lenaerts, 2010). We determine the 2.4
225 firn compaction for January 2003 to October 2009, with respect to the mean of the years 1979 to 2002 and estimate a temporal linear trend, h_{comp} . The model output is re-gridded onto the 10×10 km common grid using nearest neighbor interpolation. The standard deviation of the
228 re-gridding is less than 1 cm/yr, causing a maximum change of 2 % of the firn compaction rate. Note that the firn compaction model has a spatial resolution of 27 km, potentially neglecting finer-scale processes relevant for the altimetry data. Clearly, the re-gridding uncertainty stated
231 above is merely a minimum estimate, neglecting, for example, uncertainties in the calibration or the atmospheric forcing of the firn compaction model.

The data were re-sampled from every two days to monthly mean time periods for every
234 grid cell before estimating elevation rates. As ~~for~~with the Envisat and ICESat data, no seasonal terms ~~are~~were co-estimated and removed (i.e. annual and semi-annual). We do not apply an *a priori* correction for surface-mass balance (SMB) trends, in accordance with the GRACE
237 processing (Section 5), which requires defining a climatological reference period. Note that applying the commonly used reference period (1979 to present) leads to spurious accumulation anomalies in the altimetry data (see Appendix A.2, Fig. A.1). The derivation of an adequate
240 climatological reference epoch in the RACMO2/ANT simulations is in itself challenging and beyond the scope of this paper.

The total uncertainty of the rate of elevation change from satellite altimetry is calculated
243 by

$$\sigma_h = \sqrt{\sigma_{\text{Envisat/ICESat}}^2 + \sigma_{\text{Firm}}^2}, (2)$$

where the standard deviation of the firm correction, σ_{Firm} is the formal regression uncertainty (neglecting model uncertainties, as these are not available), and we assume the error sources to be uncorrelated. [R1_243] It is recognized that neglecting uncertainties of the firm model leads to underestimated values of σ_h . However, the magnitude of the firm correction itself is small (see Appendix A.2) compared to the observational uncertainties, and the associated underestimation of σ_h is likely to be small.

Data availability

2.5

Annual elevation trends from a combination of Envisat and ICESat data are provided for the time period between February 2003 and October 2009. Trends have been corrected for firm densification processes using RACMO2/ANT. Elevation trends are provided in a 20 km polar stereographic grid (central meridian 0° , standard parallel 71° S) with respect to the WGS84 geoid. X and Y are given in km, and the elevation rate and its standard deviation are given in m/yr.

The altimetry data and related ancillary data are directly accessible in the Pangaea

2.5.1

repository: [R2_010]

http://hs.pangaea.de/model/Sasgen-et al_2017/Ice_sheet_topographic_change.zip

ICESat elevation trend for the time period between February 2003 and October 2009.

The dataset is provided in a 10 km grid in polar stereographic projection (central meridian 0° da standard parallel 71° S) with respect to the WGS84 geoid. X and Y are given in km, and

the elevation rate and its standard deviation are given in m/yr.

Envisat elevation trend for the time period between February 2003 and October

267 2009.

The dataset is provided in a 10 km grid in polar stereographic projection (central meridian
2.5.2

0°, standard parallel 71° S) with respect to the WGS84 geoid. X and Y are given in km, and the

270 elevation rate and its standard deviation are given in m/yr.

ICESat & Envisat combination for time period between February 2003 and October

2.5.3 2009.

273 Elevation changes have been corrected for firn densification processes using a ~~FDM~~firn-

densification model. The dataset is provided in a 10 km grid in polar stereographic projection

(central meridian 0°, ~~to~~ standard parallel 71° S) with respect to the WGS84 geoid. X and Y are

276 given in km, and the elevation rate and its standard deviation are given in m/yr.

2.5.4

Annual elevation trends from CryoSat-2 derived from a single trend covering the

time period 2010-2013.

279 An acceleration term in areas with dynamic thinning was added to the linear trend to obtain

annual rates. Elevation trends are provided at 10 km resolution in a polar stereographic grid

(central meridian 0°, standard parallel 71° S) with respect to the WGS84 geoid. X and Y are
2.5.5

282 given in km and the elevation rate and its standard deviation are given in m/yr.

Elevation changes from firn model

Annual firn densification rates over 2003-2013 rates obtained from RACMO2.3. Data is

285 provided in a 27 km polar stereographic grid (central meridian 0°, standard parallel 71° S) with

respect to the WGS84 geoid. X and Y are given in km and the annual firn densification rates

in m/yr.

288

Snow / ice density map

291

[R1_284] To perform the conversion of volume change to mass change, a ~~The density map~~
~~for volume to mass conversion is~~ density map is provided in 20 km resolution in a polar
2.5.6 stereographic grid (central meridian 0° , standard parallel 71° S) with respect to the WGS84
geoid. X and Y are given in km and density in kgm/m^3 . We provide the data set at this point for
completeness; more details on the generation of this density map is given in Sasgen et al. (2017).

294

ICESat/Envisat combination mask

2.5.7 Mask used for combining ICESat and Envisat in a 10 km resolution and polar stereographic
coordinates.

297

X and Y are coordinates in km and the id represents whether ICESat or Envisat has been
used to construct the elevation change combination.

4: only Envisat was available

300

3: only ICESat was available

2: ICESat lower errors

1: Envisat lower errors

303

3. GPS UPLIFT RATE ESTIMATION & CLUSTERING

306

The aim of the GPS time series analysis is to derive uplift rates, y_u that represent the
~~geophysical~~ vertical ground motion at the sites as accurately and robustly as possible. We derive
uplift rates based on GPS records from a total of 118 Antarctic sites. Data were processed from
1995 day of year (doy) 002 to 2013 doy 257 (1995.0-2013.7) but data at individual sites are of
varying length and quality. The processing and uplift rate and uncertainty estimation
methodology are documented in detail in Petrie et al. (in prep. a, b), but a short summary is
309 given here for convenience. It resembles that of Thomas et al. (2011), but with more recent

processing software (GIPSY 6.2) and model updates (including second order ionospheric and
312 earth radiation models): an initial satellite orbit and clock estimation step is performed, using a
carefully selected balanced stable global network of GPS sites (at the time of processing JPL
reprocessed orbits for these state-of-the-art options were not available) [R2_279]. The orbits
315 and clocks are then used to perform precise point positioning (PPP) processing of all the
available Antarctic sites of interest. A mini-ensemble was created to investigate systematic
processing uncertainties and ~~manual investigation was performed of the~~ effects of possible
318 systematic errors in the time series on uplift rates. The mini-ensemble investigation showed that
decisions taken when analyzing time series tended to have larger effects on uplift rates and
uncertainties than the effects of small processing strategy changes. Outliers and systematic
321 errors, such as offsets due to equipment changes or other causes, were removed where possible.
Due to the varying characteristics of the time series it was not possible to use the same approach
at all sites. The strategy was as follows (and is summarized in Appendix A.3, Fig. A.3). For
324 sites with over 2000 days of data, uplift rates and associated uncertainties were estimated using
the CATS software (Williams 2008). We co-estimated a white-noise scale factor for the formal
uncertainties, and a power-law noise amplitude with the index fixed to -1 (flicker noise), along
327 with the temporal linear trend (rate), seasonal (annual and semi-annual) parameters, and sizes
of the offsets (at the specified epochs).

The median values of the white-noise scale (1.6) factor and the power law noise amplitude
330 (13.4 mm), [R2_295] derived from these long time series, were then used to propagate rates and
uncertainties for the shorter time series, for which CATS cannot produce reliable estimates of
the error model. For the propagation, the time series with fewer than 2000 epochs are
333 additionally subdivided into two categories; continuous sites (≥ 2.5 yr), for which periodic

parameters are estimated in the propagation of uncertainties, and very short continuous sites (< 2.5 yr) and campaign sites for which periodic parameters are not estimated. For each campaign, 1 mm of noise was added when propagating the uncertainties, to allow for tiny differences when re-setting up equipment.

Finally, for each site, the uplift rate y^u and its uncertainty σ^u are assessed by manually removing portions of the time series (for example deleting campaigns in turn). If the rate changes by an amount larger than the propagated uncertainty for the site, the uncertainty is assigned as $\pm$ the maximum difference in rate, and the rate is adjusted, if necessary, to the values of the most likely part of the range. Sites with only two campaigns were assigned an uncertainty of ± 100 mm/yr, unless there was further evidence for or against the existence of systematic errors.

Table 1 summarizes the rate estimation methods and the number of sites for each. For further details and full information on individual rates and time series, see Petrie et al. (in prep a) for a full description of the processing and ensemble evaluation, and Petrie et al. (in prep b) for details of time series analysis and rate and uncertainty estimation. ~~Table 1 shows the numbers of sites at which each approach was taken.~~ Further work was undertaken to combine or ‘cluster’ the rates regionally for inclusion in the estimation process – see the REGINA Paper II (Sasgen et al. 2017), ~~Section 3.2.2 for details and Table Appendix A.4 for more details.~~

Table 1: Number of sites for each GPS uplift rate and uncertainty estimation method.

Rate and uncertainty estimation method	Number of sites (118 total)
CATS rate and uncertainty ('cats, cats')	18
CATS rate, manually increased uncertainty ('cats, eman')	2
Propagated rate and uncertainty ('prop, prop')	28
Propagated rate and manually increased uncertainty ('prop, eman')	50
Manually adjusted rate and manually increased uncertainty ('rman, eman')	20

Table 2. Uplift rates y^u and associated uncertainties σ^u (mm/yr) for selected GPS sites with more than 2000 epochs of data, compared to data published by Thomas et al. (2011) and Argus et al. (2014). Temporal components and noise characteristics are derived using the CATS software (Williams 2008), i.e. 'cats, cats' method.

Site	REGINA		Thomas et al. (2011)		Argus et al. (2014)	
	y^u	σ^u	y^u	σ^u	y^u	σ^u
cas1	1.5	0.2	1.2	0.4	1.7	0.8
crar	0.7	0.4	1.0	0.7	1.0	0.6
dum1	-0.3	0.3	-0.8	0.5	-0.2	0.8
maw1	-0.4	0.2	0.1	0.4	0.2	0.6
mcm4	0.8	0.2	0.7	0.4		
sctb	0.9	0.5	0.6	1.1		
syog	1.1	0.2	2.3	0.4	0.6	0.8
tnb1	0.1	0.5	-0.2	0.8	-0.4	1.0
vesl	0.4	0.3	1.1	0.5	1.5	0.8
McMurdo*					1.0	0.6

*Sites: crar-sctb-mcm4-mcmd

3

Comparison with existing results

354

Next, we briefly compare the uplift rates at individual sites (data span 1995.0-2013.7)

derived from the GPS processing described above with those available from three previous studies: Thomas et al. (2011) (data span 1995.0-2011.0), Argus et al. (2014) (data span 1994-
357 2012) and the more geographically limited set of Wolstencroft et al. (2015) (data span 2006-
late 2013, focused on Palmer Land). It should be noted that the REGINA and Wolstencroft et al. (2015) rates are in ITRF2008, the Thomas et al. (2011) rates are in ITRF2005 (which has
360 negligible scale or translation differences to ITRF2008), and the Argus et al. (2014) rates are in a reference frame specific to the paper which they note yields 0.5 mm/yr more uplift than ITRF2008 at high southern latitudes. All rates from Argus et al. (2014) in Tables 2, 3 and 4 are shown as given in the original paper. [R2_327].

Due to the large number of Antarctic sites, in total 118, we focus the comparison on the uplift rates and uncertainties derived by the methods ‘cats, cats’ (Table 2) and ‘prop, prop’ (Table 3). Uplift rates resulting from our study are provided in Appendix A.4 for all sites (Table A.2). Tables A.3 shows comparisons with the values of Thomas et al. (2011) and Argus et al. (2014) for ‘prop, eman’ sites not shown in the main text. All uplift rates, y^u , are in mm/yr, with uncertainties reflecting 1-sigma standard deviations, σ^u . Sites with particularly complex non-linear time series such as those at O’Higgins (ohi2, ohig) and Palmer (palm) in the Antarctic Peninsula are omitted here, as comparison with different studies is potentially misleading due to the effects of different measurement time periods. Table 2 shows data for selected sites with

Table 3. Uplift rates y^u and associated uncertainties σ^u (mm/yr) for selected GPS sites with fewer than 2000 epochs for data, compared to data published by Thomas et al. (2011) and Argus et al. (2014). Noise characteristics are derived median values from CATS software results for longer station records and propagated in the parameter estimation (‘prop, prop’ method). See Appendix A.4, Table A.2 for a full list of rates from this study.

Site	REGINA		Thomas et al. (2011)		Argus et al. (2014)	
	y^u	σ^u	y^u	σ^u	y^u	σ^u
belg	-1.4	0.7	3.0	1.5	0.8	2.4
dupt	11.5	1.1			12.4	2.5
fonp	13.5	1.8			14.8	3.4
frei	-4.4	0.7			-2.9	1.4
hugo	0.9	1.3			1.7	3.6
robi	8.7	1.5			8.7	3.2
roth	5.5	1.4			5.4	1.4
svea	1.3	1.1	2.1	2.0	1.7	2.9
vnad	4.4	1.1			5.2	2.5

long time series, where uplift rate and uncertainty were derived using the CATS software
375 (Williams 2008). Uplift rates at the majority of the GPS sites agree within uncertainty, except
syog (Syowa), where the REGINA value is between that from the other two studies. The
uncertainty limits for the REGINA value and the Argus et al. (2014) just meet at 0.9 mm/yr,
378 even when allowing for the ~ 0.5 mm difference in reference frames, but the Thomas et al.
(2011) value does not. This may be due to the fact that Thomas et al. (2011) estimate two offsets
in the series. Table 3 shows uplift rate comparisons for sites where the ‘prop,prop’ method was
381 used; the noise characteristics are derived from median values from CATS software results for
longer site records and then propagated in the parameter estimation in which annual and semi-
annual parameters were also estimated along with the trend. Again, the rates agree within
384 uncertainty, except for site belg where there is a disagreement with Thomas et al. (2011). This
may be due to their shorter data span. Table 4 shows comparisons for sites where the REGINA
rates and uncertainties have been manually evaluated based on the spread of rates obtained by
387 sub-sampling the time series (‘rman’ method). There is a large difference (over 10 mm/yr) in
the values at capf (Cape Framnes) between the REGINA value (4.0 ± 1.4 mm/yr) and the Argus
et al. (2014) value (15.0 ± 4.2 mm/yr). Interestingly, the Wolstencroft et al. (2015) rate values
390 for bean, gmez, lntk, mkib, and trve are all systematically higher than the REGINA values, by
an average of just over 3 mm/yr, and the uncertainties we assigned are also several times larger.

[R1_370] The systematic differences between Wolstencroft et al. (2015) and the REGINA
393 values for Palmer Land are currently unexplained and a matter of ongoing investigation. For
~~more detailed analysis of rates and time series at individual sites, see Petrie et al. (in prep b).~~

Table 4. Uplift rates y^u and associated uncertainties σ^u (mm/yr) for selected sites where uplift rates are manually evaluated based on the spread of rates obtained by sub-sampling the time series ('rman' method), compared to data published by Thomas et al. (2011), Argus et al. (2014), Wolstencroft et al. (2015). See also 'rman' sites in Table Appendix A.4, Table A.2.

Site	REGINA		Thomas et al. (2011)		Argus et al. (2014)		Wolstencroft et al. (2015)	
	y^u	σ^u	y^u	σ^u	y^u	σ^u	y^u	σ^u
bren	3.1	1.1	3.9	1.6	2.1	3.7	3.2	0.8
capf	4.0	1.4			15.0	4.2		
dav1	-1.6	0.6	-0.9	0.5	-0.8	1.0		
mait	0.4	1.1	0.1	0.6	1.3	0.7		
mbl3	1.3	17.9	0.1	2.0				
bean	2.1	4.3					7.5	1.2
gmez	1.5	4.8					5.7	0.8
lntk	4.6	3.1					6.0	0.7
mkib	4.7	2.6					6.9	0.5
trve	2.5	5.6					4.7	0.6

Data availability

The GPS data and related code are directly accessible in the Pangaea repository,

http://hs.pangaea.de/model/Sasgen-etal_2017/In_situ_GPS_uplift_rates.zip

3.2

Bedrock uplift rates

Bedrock uplift rates derived for the REGINA project are available in the text file

“REGINA_rates_full.txt”, as presented in Table A.2 and A.3 of the Appendix A.4. The files

“REGINA_rates_03-13.txt” and “REGINA_rates_03-09.txt” contain subsets of the data, with

the temporal coverage limited to 2003-2013.5 and 2003-2009, respectively. The files are

organized as follows:

Lon [°], Lat [°], uplift rate [mm/yr], uncertainty of the uplift rate [mm/yr], GPS site ID

These *.txt files are the input to the clustering script described below. No elastic correction

has been applied.

3.2.2

Clustering script

In addition to the uplift rates for individual GPS sites, we provide a *bash* script “cluster.sh”

for clustering the heterogeneous data according to their geographic locations, for a pre-defined

threshold value. The idea is to reduce stochastic and geophysical noise of neighboring stations

in order to obtain uplift rates that are better regional representations for the length scale

recovered with GRACE (ca. 200 km). In an iterative procedure, the script selects neighboring

sites within a threshold ranging from 10-220 km [R1_361] and calculates the weighted average

of the uplift rates and a simple uplift average [R1_361] of the stations locations. Input to the

script are the REGINA rate files, specified in the previous Section 3.2.1. Further details and the

application to the GPS data set can be found in REGINA paper II (Sasgen et al., submitted2017)

[R1_361]. Note that the script relies on the open-source program suite Generic Mapping Tools,

<http://gmt.soest.hawaii.edu/> (Wessel et al. 2013). Similar clustering can be achieved with the function *kmeans* in Matlab® or its open-source alternative GNU Octave.

GPS time series

The GPS time series were created as part of the RATES project, not solely the REGINA study. ~~They will be made available along with the detailed descriptions in Petrie et al. (in prep b). The time series of vertical bedrock displacement will then be accessible here: [LINK]. The~~ data can be obtained upon request from co-author Dr. Elizabeth Petrie.

4. GRAVIMETRY DATA ANALYSIS

We investigate the Release 5 (RL05) GRACE coefficients of the Centre for Space Research (CSR; Bettadpur, 2012) and the German Research Centre for Geosciences (GFZ; Dahle, 2013), provided up to spherical-harmonic degree and order $j_{max}=96$ and 90 respectively in the Science Data System (SDS). For reasons of comparison, we adopt $j_{max}=90$ for both GRACE solutions. A temporal linear trend in the ocean bottom pressure variations modeled by the atmospheric and oceanic background models (GAD) was re-added to the monthly solutions, according the GRACE Science and Data System recommendation (Dobslaw et al. 2013). The GRACE coefficients C_{20} were replaced by estimates from Satellite Laser Ranging (SLR) provided by Cheng et al. (2013). In our analysis we apply the cut-off degrees $j_{max}=50$, which has been commonly used, as well as $j_{max} = 90$, which is considered experimental in terms of the remaining signal content.

The determination of the rate of the gravity field change over Antarctica follows the scheme sketched in Fig. 3. The rate of the gravity field change, expressed as equivalent water height variations, is estimated in the spatial domain by adjusting a six-parameter function consisting of a constant, a temporal linear trend and annual and semi-annual harmonic amplitudes. A

quadratic term was not co-estimated due to the project's focus on the rates (i.e. temporal linear

444 trends). It should be stated that including a quadratic term would slightly reduce the residual
uncertainties, particularly in the Amundsen Sea Sector, where an ice-dynamic [R1_476]
acceleration of mass balance rates occurs that is not accounted for by ~~interannual-non-linear~~
447 [R1_476] SMB variations of the ice sheet (see Section 4.2).

The post-processing of the GRACE coefficients follows three main steps:

Step 1: Optimization of de-stripping filter

450 Due to effects like the propagation of measurement noise and temporal aliasing, a large
proportion of the variations contained in the monthly solutions is related to noise. The noise of
the monthly solutions is lowest close to the pole and exhibits a characteristic north-south

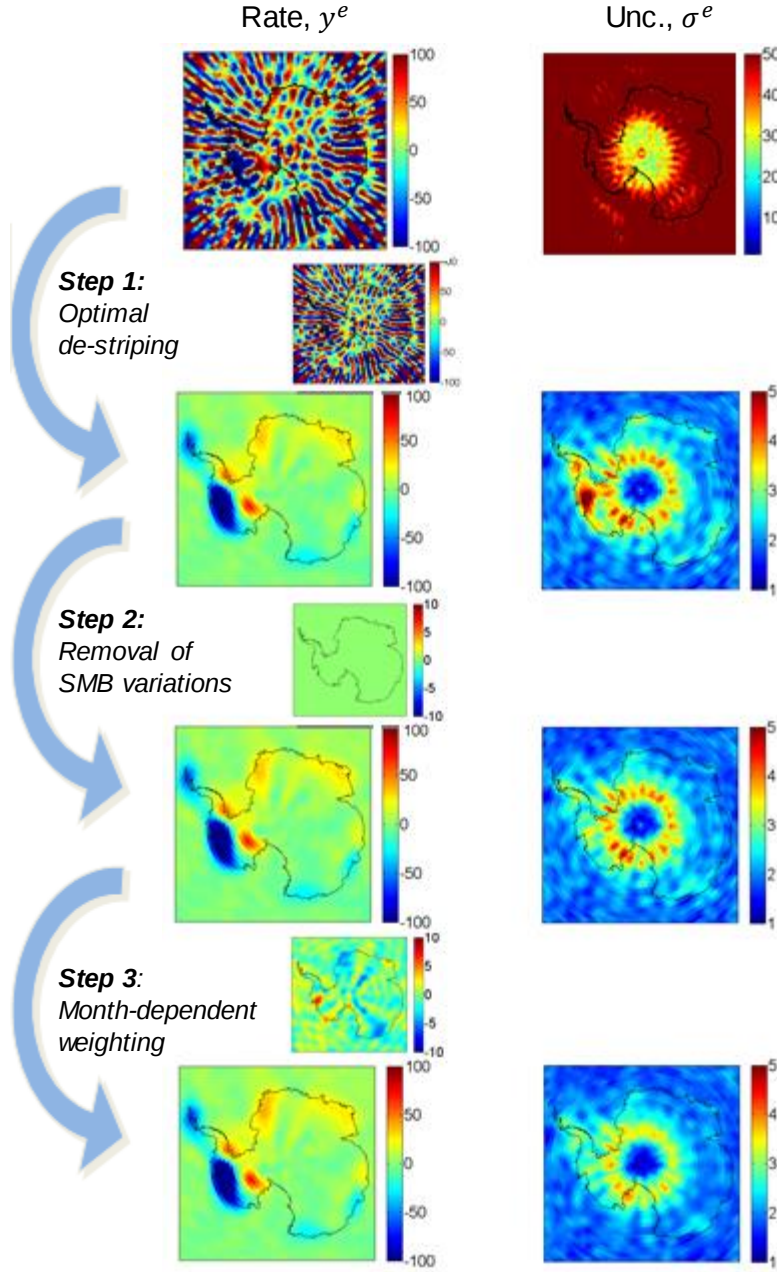


Figure 3. Post-processing steps applied to the GRACE gravity fields; shown is the impact on the gravity field rate y^g (left) and the associated RMS uncertainty σ^g (right). Small maps show change in the gravity field rate between two subsequent steps. Color scale is mm w.e./yr. GRACE data is GFZ RL05a.

oriented stripe pattern. This is visible in the gravity field rate and the propagated Root-Mean-Square (RMS) uncertainties shown in Fig. 3. In order to remove the stripe pattern, we apply the

de-correlation filter of Swenson & Wahr (2006) (hereinafter, “Swenson filter”) specifically
456 tuned to optimize the recovery of the gravity field rate over the region of Antarctica, which is
detailed in Section 4.1. Fig. 3 shows that the de-stripping procedure reduces the RMS uncertainty
of the rate by approximately one order of magnitude.

459 Step 2: Reduction of ~~interannual-non-linear~~ mass variations [R1_476]

For isolating gravity field rates, the second step in the processing is the reduction of de-
trended variations of the surface mass balance, caused by accumulation events. The data set
462 used for this purposes is the RACMO2/ANT (Lenaerts et al 2012) converted into monthly sets
of spherical harmonic coefficients. The reduction of these ~~interannual-accumulation~~ variations
[R1_476] does not change the temporal linear trend, but it reduces RMS uncertainties especially
465 in coastal regions (Fig. 3). Details are provided in Section 4.2.

Step 3: Month-dependent weighting

The performance of the GRACE satellite system was weaker in the early mission phase
468 due to issues with the star cameras of the satellites (*C. Dahle, GFZ, pers. comm.*; Fig. 5). A rate
estimate with uniform weighting of all months does not account for these variations. Therefore,
in the last step, month-dependent uncertainties are estimated and applied as weights during the
471 linear regression of the temporal linear trend. This slightly changes both the resulting rate
estimate, as well as its RMS uncertainties. Details are provided in Section 4.3.

Finally, after post-processing and evaluation of the gravity field rate (Section 4.4), we select
474 the GRACE release and cut-off degree providing the lowest uncertainty level (Section 4.5) as
reference input for our joint inversion for present-day ice-mass change detailed in REGINA
paper II (Sasgen et al. 2017).

The Swenson filter has been proven to effectively reduce the typical north-south correlated error structures of GRACE monthly solutions. The filter is based on the observation that these 4.1 structures correspond to correlated patterns in the spherical harmonic domain, namely correlations within the coefficients of the same order and even degree, or respectively, odd degrees (Swenson & Wahr, 2006). The standard way of fitting and removing these patterns is by adjusting polynomials to the respective sequences of spherical harmonic coefficients, independently for individual months. Parameters to choose are the degree of the polynomial n_{pol} and the minimum order m_{start} starting from which this procedure is applied. In principle, a higher degree polynomial reduces the variability of coefficients of even / odd degree, and results, also at lower minimum order, in stronger filtering – however, the behavior of the filter may differ for regional applications, as discussed below. Note that tuning of other parameters has been presented, e.g. the window width (Duan et al. 2009) or the degree range to which the filter is applied. Chambers and Bonin (2012) have assessed these parameter options with regard to the new GRACE RL05 solutions and global oceanic signals. Here, we perform a detailed analysis of the choice of the Swenson filter parameters in order to optimize the signal-to-noise characteristics of the rate of the gravity-field change over Antarctica. ~~The resulting gravity field rates are later used in the joint inversion for present-day ice mass change and GIA described in REGINA Part II.~~

We assess signal corruption by applying the filter to a synthetic test signal, which is based on high-resolution elevation rates from satellite altimetry and reflects the prevailing signatures of present-day ice-change with sufficient realism. For each choice of filter parameters, the signal corruption is assessed as the RMS difference between the original and the filtered

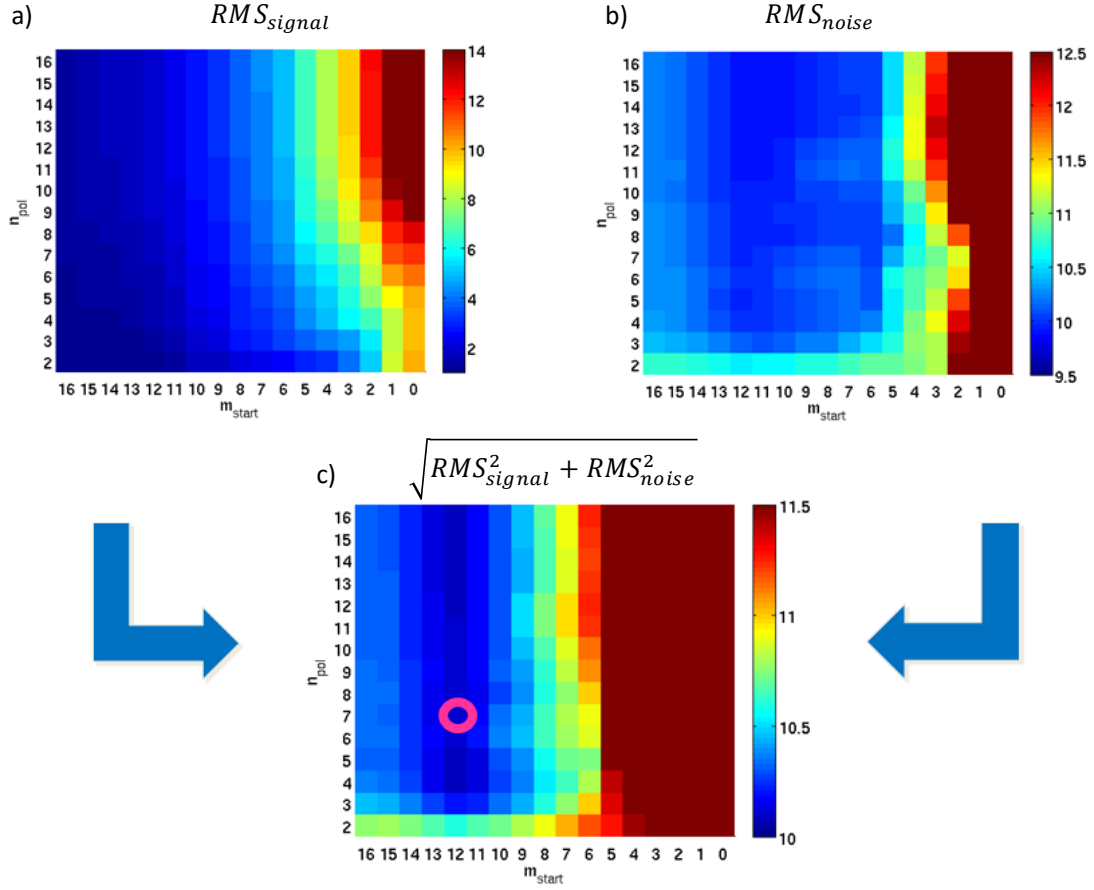


Figure 4. Effect of Swenson filter parameters m_{start} and n_{pol} on a) signal corruption, b) noise reduction and c) combined effect on signal and noise. RMS residuals are shown for the gravity field rates in (mm w.e./yr). The optimal choice of filter parameters $m_{start} = 12$ and $n_{pol} = 7$ is indicated as circle. Results are shown for GFZ RL05a with $j_{max} = 90$.

synthetic signal, RMS_{signal} . The RMS is evaluated in terms of water-equivalent height per year

501 for the signal components within the region south of 60°S latitude.

For assessing the noise and noise reduction in the filtered fields, we face the task of separating the noise from the geophysical signals in the gravity field rates derived from

504 GRACE. Here we attempt such a separation by reducing *a priori* information on the rate of ice mass change from the GRACE fields and considering the residual as an upper bound representation of noise. The *a priori* information is, again, based on elevation rates. For the

noise assessment we then take the RMS of the residual rates in terms of water equivalent height per year, RMS_{noise} , again for the region south of 60°S latitude. Since the residual gravity field rates may still contain some geophysical signal, we consider this noise estimate as an upper bound for the true GRACE uncertainties. It should be stated that, after the Swenson filtering, an additional Gaussian filtering is applied to the signal and noise models with a 200 km filter width, which was determined to be the optimal smoothing half-width for the signal-to-noise ratio in the GRACE spectra by Wiener optimal filtering (Sasgen et al. 2006) as reflected in the degree-amplitude spectrum.

Fig. 4 shows the assessed signal corruption and noise reduction as a function of the two Swenson filter parameter choices, the polynomial degree n_{pol} and the minimum order m_{start} . The results are shown for the gravity field expanded to degree and order $j_{max} = 90$ of the GFZ RL05a coefficients, even though using $j_{max} = 50$ and CSR RL05 yields similar results. As expected, the signal corruption, RMS_{signal} increases with increasing strength of the Swenson filter, that is with increasing n_{pol} and the decreasing minimum order m_{start} . In terms of noise reduction, we see as expected that stronger filtering (increasing n_{pol} ; decreasing m_{start}) decreases the RMS_{noise} (Fig. 4). However, we find that only for the range of filter parameters with $m_{start} \geq 10$. For filter parameters $m_{start} < 10$ this pattern is reversed, and RMS_{noise} . A closer analysis indicates that the consideration of the low orders into the Swenson filtering transfers energy (both from signal and noise) from low-to-mid latitudes to the Polar Regions, which leads to a considerable signal corruption over the region of interest, that is only avoided. We avoid this degradation by limiting the range of filter parameters in this the subsequent regional analysis optimization of the gravity trends to $m_{start} \geq 10$.

To define the optimal filter parameters a quadratic sum of the signal corruption and noise

reduction is computed, allowing us to balance both effects, the optimal values are $m_{start} =$

12 and $n_{pol} = 7$ as indicated in Fig. 4c. These filter parameters are subsequently used. For

comparison it is stated that Chambers & Bonin, 2012 ~~find~~found $m_{start} = 15$ and $n_{pol} = 4$

~~as to be~~ optimal for oceanic applications. [R1_511] Note that the signal corruption is assessed

only to optimize the de-stripping filter. Possible signal degradation due to de-stripping is not

included in the uncertainty estimate of the optimally filtered GRACE trends. However, signal

loss due to the additional smoothing with a 200 km Gaussian filter is accounted for by applying

the same filter to the viscoelastic response functions, as well as the altimetry-based input fields

(Appendix A.5).

Reduction of ~~interannual-non-linear~~ mass variations

4.2

~~Interannual variations are a major constituent of t~~The temporal variations of the Antarctic gravity field show a strong year-to-year fluctuation, apart from the linear trend (Wouters et al.

2014). [R1_476]. A large portion of the non-linear signal in geodetic mass and volume time

series is well explained by modelled SMB fluctuations (Sasgen et al. 2010; Horwath et al.

2012). Towards the ultimate goal of isolating the linear GIA signal from time series of mass

change, we removed non-linear effects of modelled SMB variations from the GRACE time

series; for this we calculate the *monthly cumulative SMB anomalies* with respect to the time

period 1979 to 2012 obtained from RACMO2/ANT (Lenaerts et al. 2012).

We then transfer the monthly cumulative SMB anomalies in terms of their water-equivalent

height change into the spherical harmonic domain and subtract them from the monthly GRACE

coefficients. In principle, the reduction of the SMB variations from the GRACE time interval

has two effects: first, it may change the overall gravity field rate derived from GRACE,

depending on the assumption of the SMB reference period. Ideally, the reference period reflects

a state of the ice sheet in which input by SMB equals the outflow by ice discharge, and SMB anomalies estimated for today reflect the SMB component of the mass imbalance. However, any bias in the SMB in the reference period leads to an artificial trend in the ice sheet mass balance attributed to SMB. This is an undesired effect, and to avoid it we de-trend the cumulative SMB time series for the time interval coeval to the GRACE analysis (February 2003 to October 2009), before subtracting it from the GRACE gravity fields, ~~rates derived from yielding zero difference in the gravity field rates GRACE (zero difference for before and after processing~~ *Step 2*, (Fig. 3) [R1_402]. The second effect is the reduction of the post-fit RMS residual for this known temporal signal variation. After reducing the SMB variations, the propagated RMS uncertainty of the derived gravity field rate becomes closer to the uncertainty level of the GRACE monthly solutions (Fig. 3).

The quality of GRACE monthly solutions changes with time, for example due to changing orbital sampling patterns (Swenson & Wahr 2006). Fig. 5 shows the temporal evolution of RMS

4.3

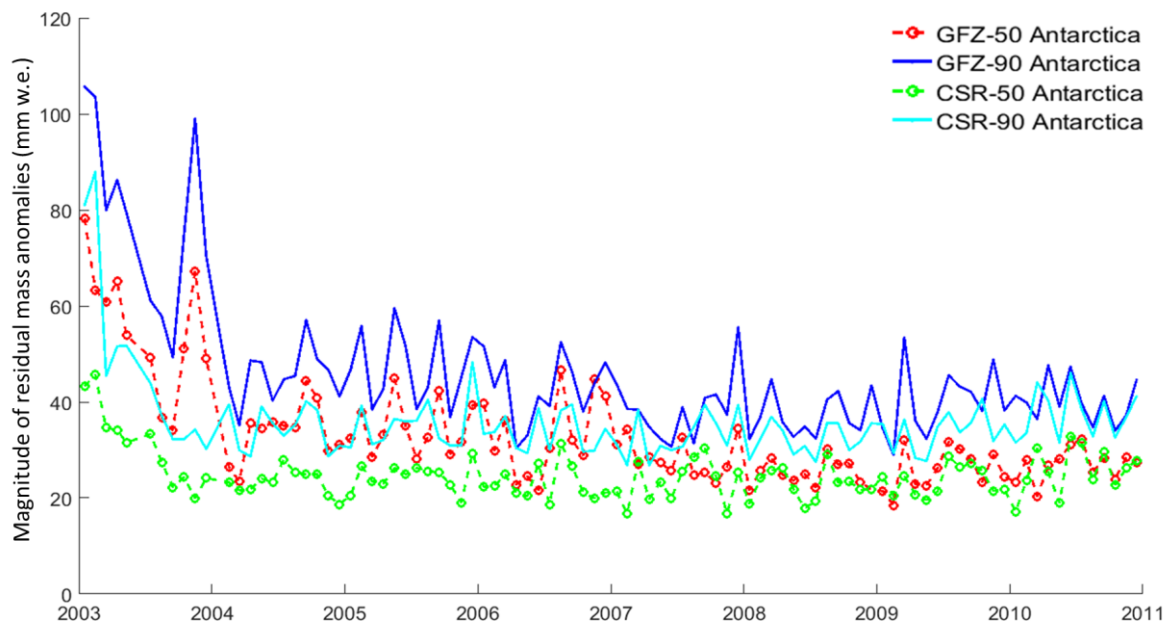


Figure 5 [R1_495]. RMS-Residual mass anomaly uncertainty of monthly GRACE gravity fields for 2003-2011, averaged over the Antarctic region south of -60°S latitude.

Shown are results for GFZ RL05a and CSR RL05 and $j_{\max} = 50$ and $j_{\max} = 90$.

uncertainties of the monthly GRACE gravity fields in the Antarctic region: sShown are residual mass anomalies, integrated over Antarctica, after the grid-based removal of the temporal linear trend and annual oscillation components and applying the filtering described in Step 1 and removing the SMB fluctuations in Step 2 [R1_495]. Note that an annual oscillation component is included to remove possible seasonal fluctuations in SMB not captured by the regional

climate model [R1_495]. However, omitting the annual oscillation component yields similar results. The residual monthly mass anomalies are attributed to noise and are used To improve the accuracy of the estimate of the gravity field rate, we include monthly uncertainties as weights in our least-squares linear regression, applied as Step 3 of the GRACE processing. Fig. 5 shows that these uncertainties are higher during early 2003. Applying the monthly dependent weighting has the effect of reducing the influence of the first months of the year 2003 on the estimated gravity field rate, which is similar to shortening the time series, given the relatively large uncertainties. Also As expected, the post-fit RMS uncertainty associated with the rate reduces, if the early months of the year 2003 are excluded [R1_509]., indicating that down-weighting the months from early 2003 is more beneficial than retaining a longer time series. Altogether, the month-dependent weighting reduces the magnitude of stripe patterns characteristic for the uncertainty of GRACE monthly solutions, and yields a more accurate realistic representation estimate of the of propagated RMS uncertainty associated with the gravity field rates (Fig. 3) [R1_514].

588 Fig. 6 shows the estimated RMS uncertainty of the gravity field rate over Antarctica, after
 post-processing. It is evident that the largest uncertainties are located in a ring south of -80°S
 4,4 latitude. This is explained by the design of the Swenson filter; little or no noise reduction is
 591 achieved close to the poles, as the gravity field is represented by near-zonal coefficients, which

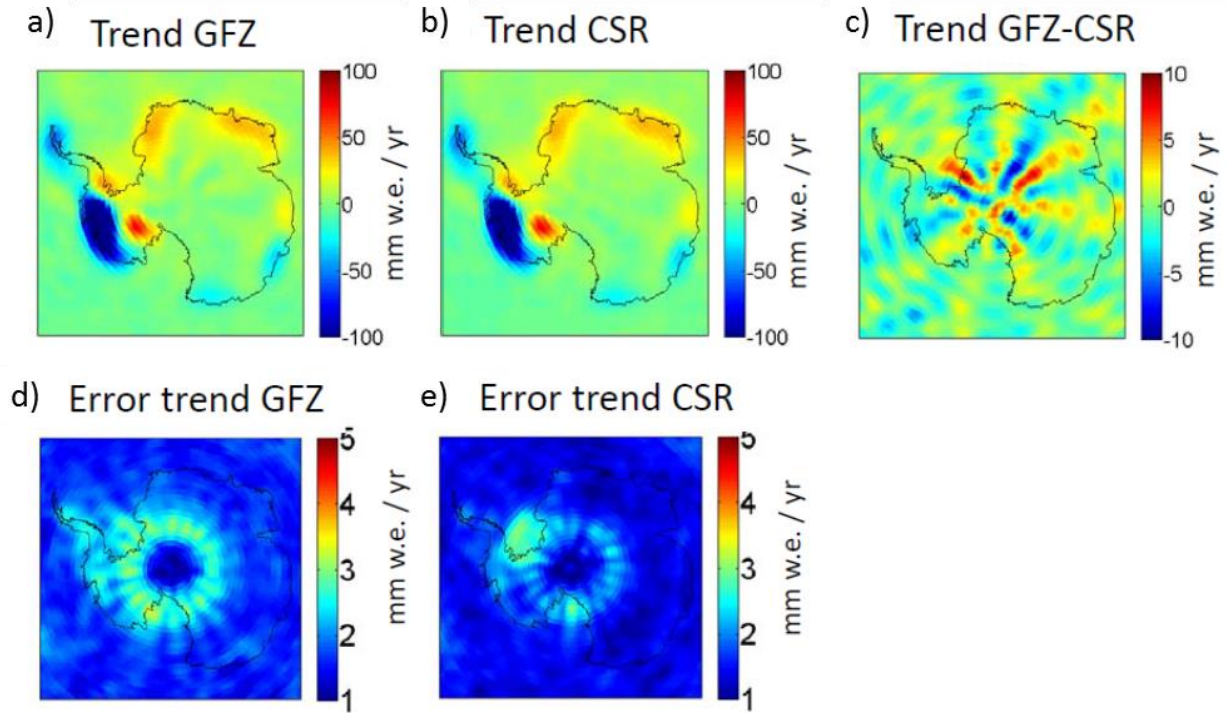


Figure 6. Linear trend in the GRACE gravity fields for the years 2003-2009; a) GFZ RL05a, b) CSR RL05, c) difference between rates from GFZ RL05 and CSR RL05, propagated d) RMS uncertainty for GFZ RL05a and e) RMS uncertainty for CSR RL05.

pass the filter mostly unchanged ($m_{start} = 12$). ~~It is observed that e~~Extending the kernel of the Swenson filter to these near-zonal coefficients ($m_{start} \leq 10$) creates high signal corruption
 594 and is not suitable for the optimal rate estimate over Antarctica (see Section 4.1). Larger uncertainties are also estimated for the Ronne and Ross ice shelf areas, which are most likely a consequence of incomplete removal of the ocean tide signal during the GRACE de-aliasing

procedure (Dobslaw et al. 2013). It should also be stated that the RMS uncertainty estimate does not include possible systematic errors in the GRACE solutions, e.g. due to a long-term drift behavior of the observing system.

Selection of GRACE release

Our evaluation of the monthly GRACE uncertainties (Fig. 5), as well as the propagated RMS uncertainty of the temporal linear trend (Fig. 6) indicates that the lowest noise level for the Antarctic gravity field rate (February 2003 to October 2009) is currently achieved with GRACE coefficients of CSR RL05, expanded to $j_{max} = 50$. We therefore refrain from including coefficients with $j_{max} > 50$ in order not to ~~comprise~~ compromise the rate estimates by unnecessarily increasing the noise level (see Appendix A.5, Fig. A.3). We adopt CSR RL05 with $j_{max} = 50$ as our preferred solutions for the representation of the gravity field rates over Antarctica, even though GFZ RL05 with $j_{max} = 50$ yields very similar rates (Fig. 6). This choice is supported by the joint inversion, as CSR RL05 with $j_{max} = 50$ provides the highest level of consistency (lowest residual misfit) with the altimetry and GPS data sets (see REGINA Part II, Sasgen et al. 2017, Supplementary Information, Section S.3), which we interpret to ~~indicate as a~~ minimum of spurious signals in the trends. To account for the uncertainty related to our choice of the solution, we consider not only RMS uncertainties of the GRACE rates but also solution differences, in the uncertainty of the final GIA estimate (Fig. 6). The solution difference represent the absolute deviation between trends from GFZ RL05 and CSR RL05 (February 2003 to October 2009, cut-off degree $j_{max} = 50$). These are then summed up squared with the propagated RMS uncertainties. It is acknowledged that the solution differences contain systematic noise arising from the GRACE processing; the pattern and magnitude may change over time. However, they provide a measure how much the results will change, if a GRACE

release alternative to CSR RL05 is considered. The difference between GRACE rates filtered
621 with Gaussian smoothing of 200 km and the optimized Swenson filter together with Gaussian
smoothing of 200 km is shown in the Appendix A.5, Fig. A.4.

Data availability

624 The gravity data and related code are directly accessible in the Pangaea repository,
4.6 [http://hs.pangaea.de/model/Sasgen-et al_2017/Geoid-
height_change_from GRACE satellite.zip](http://hs.pangaea.de/model/Sasgen-et al_2017/Geoid-height_change_from_GRACE_satellite.zip)

Stokes coefficients of gravity field change

627
4.6.1 The monthly GRACE gravity field solutions from the Data System Centers GFZ and CSR
are available under <ftp://podaac.jpl.nasa.gov/allData/grace/L2/> or <http://isdc.gfz-potsdam.de/> as
630 spherical harmonic (SH) expansion coefficients of the gravitations potential (Stokes
confidents). More information is available in Bettadpur (2012). The data archive contains
temporal linear trends of the fully normalized Stokes coefficients in the ‘geodetic norm’
633 (Heiskanen & Moritz, 1967), complete to degree and order 90, inferred from these time series
according to Section 4.7. We provide data for GFZ RL05 and CSR RL05, for the time period
2003-2009 and 2003-2013, and for various combinations of filtering. The coefficients are
636 organized as:
4.6.2

[Degree j], [Order m], [c_{jm}], [s_{jm}]

Code for de-striping filtering

639 The Matlab® function “KFF_filt” performs decorrelation filtering for sets of spherical
harmonic coefficients, typically from GRACE gravity field solutions, after the idea of Swenson
& Wahr (2006). An open-source alternative to Matlab® is GNU Octave

642 <https://www.gnu.org/software/octave/>. The function is called as `KFF_filt = swenson_filter_2(KFF, ord_min, deg_poly, factorvec, maxdeg)`, where variables `ord_min` and `deg_poly` equal m_{start} and n_{pol} , respectively, in Section 4. KFF contains the sets of spherical
645 harmonic coefficients in the 'triangular' format (not memory-efficient but intuitive). For example, for a set of coefficients with maximum degree $j_{max} = 3$ and maximum order $m_{max} = 3$, the set of coefficients is stored in a $j_{max} \times m_{max}$ matrix in the following way:

```
648 % KFF = [0 0 0 c_00 0 0 0;
% 0 0 s_11 c_10 c_11 0 0;
% 0 s_22 s_21 c_20 c_21 c_22 0;
651 % s_33 s_32 s_31 c_30 c_31 c_32 c_33]
```

5. VISCOELASTIC MODELLING

654 The Earth structure of Antarctica is characterized by a strong dichotomy between east and west, separated along the Transantarctic Mountains (e.g. Morelli & Danesi, 2004). Recent seismic studies have produced refined maps of crustal thicknesses also showing slower upper-
657 mantle seismic velocities in West Antarctica, indicating a thin elastic lithosphere and reduced mantle viscosity (An et al. 2015; Heeszel et al. 2016). Moreover, yield strength envelopes of the Earth's crust and mantle suggest the possibility of a viscously deforming layer (DL) in the
660 lower part of the crustal lithosphere (Ranalli & Murphy, 1987), a few tens of km thick and with viscosities as low as 10^{17} Pa s (Schotman et al., 2008). High geothermal heat flux is in agreement with the seismic inferences of a thin elastic lithosphere and low mantle viscosity,
663 and would favor the presence of such a DL also in West Antarctica (Shapiro & Ritzwoller 2004; Schroeder et al. 2014).

The choice of the viscoelastic modelling approach used to determine load-induced surface displacements and gravitational perturbations is governed by three main requirements; i) to accommodate lateral variations in Earth viscosity, ii) to allow for Earth structures with thin elastic lithosphere and low viscosity layers, in particular including a DL, and iii) to provide viscoelastic response functions for the joint inversion of the satellite data described in REGINA paper II (Sasgen et al.-2017submitted). With regard to point iii) it should be mentioned that the viscoelastic response functions provide a geophysical meaningful way to relate surface displacement and gravity field changes, considering also dynamic density changes within the Earth's interior . Moreover, it allows us to consider the changes in the ratio of surface-displacement and gravity field changes caused by the Earth structure, in particular, the lithosphere thickness. Another advantage is that different filtering can be applied to the viscoelastic response functions in order to match the filtering of the input data set, avoiding the introduction related biases (Appendix A.5)[R1_653].

To meet these requirements, we adopt the time-domain approach (Martinec 2000) for calculating viscoelastic response functions of a Maxwell continuum to the forcing exerted by normalized disc-loads of constant radius. Then, the magnitudes and spatial distribution of the surface loads are adjusted according to the satellite data to obtain the full GIA signal for Antarctica. The forward modelling of viscoelastic response functions is a classic topic in solid Earth modelling (e.g. Peltier & Andrews, 1976), however, their application to inverting multiple-satellite observations for present and past ice sheet mass changes is new and applicable to other regions, such as Greenland or Alaska.

The viscoelastic response function approach allows for high spatial resolution at low computational cost in the numerical discretization of the Earth structure as well as in the

representation of the load and the response. In addition, we can accommodate a high temporal resolution, which is required when considering low viscosities and associated relaxation times of only a few decades. The spherical harmonic cut-off degree for the simulations shown in the following is $j_{max} = 2048$ (ca. 10 km).

Load model parameters

The load function $\sigma(t, \vartheta)$ is disc shaped with a constant radius of ca. 63 km. The radius of 63 km matches the mean radius of the discs south of 60°S of the geodesic grid (here, ICON 1.2 grid, status 2007, e.g. Wan et al., 2013), which underlie the joint inversion of the altimetry, gravimetry and GPS observations (see REGINA paper II, Sasgen et al. [submitted2017](#)). The resolution of the geodesic grid is chosen to allow for an adequate representation of the load and viscoelastic response with regard to the input data sets, while minimizing the computational cost. The disc load experiment consists of a linear increase in the ice thickness at a rate of 0.5 m/yr continuing until a new dynamic equilibrium state between load and response is reached. [R2_658] After the application of the constant loading rate, two extra time steps are done with no loading change to give the purely viscoelastic response. For West Antarctica, the loading rate is held constant for 2000 years, for East Antarctica it is 15,000 years, which are longer times than needed to reach dynamic equilibrium (see Appendix A.8). -With reference to the assumed ice density of 910 kg/m³, this thickness increase corresponds to a mass gain of ca. 5.6 Gt/yr. Then, to obtain the signal component of the viscous Earth response only, the elastic response and the direct gravitational attraction of the load are subtracted.

The experiment is designed as an *increasing* load, for example representative for the ceasing motion of the Kamb Ice Stream (Ice Stream C; Retzlaff & Bentley, 1993), West Antarctica. Due to linearity of the viscoelastic field equations, it is not necessary to calculate

separately the equivalent *unloading* experiment, $-\sigma(t, \vartheta)$, for example corresponding to the past and present glacier retreat of the Amundsen Sea Sector, West Antarctica (Bentley et al. 2014 and Rignot et al. 2014, respectively). Among others, the combined inversion of the altimetry, gravity and GPS data (REGINA paper II, Sasgen et al. 2017) solves for the magnitude and the sign of the load, allowing for ice advance as well as ice retreat.

Earth model parameters

We set up an ensemble of 58 simulations representing different parameterizations of the viscosity structure (Table 5), split into West Antarctica (56 simulations) and East Antarctica (2 simulations). The ensemble approximately covers the range of values of the viscosity and lithosphere thickness inferred from Priestley & McKenzie (2013) [R1_696]. For West Antarctica, varied parameters are the lithosphere thickness, h_L (30 to 90 km in steps of 10 km), the asthenosphere viscosity (1×10^{18} Pa s to 3×10^{19} Pa s in four steps), and the presence of a ductile lower crust, DL, with 10^{18} Pa s. For East Antarctica, we employ parameter combinations appropriate for its cratonic origin with h_L of 150 km and 200 km, and an asthenosphere viscosity equivalent to the upper-mantle viscosity of 5×10^{20} Pa s. These values lie in the range of previously applied viscosity values in Antarctica (Nield et al. 2012; Whitehouse et al., 2012; Ivins et al., 2013; van der Wal et al., 2015). For the radial layering of the elastic properties, we adopt the Preliminary Reference Earth Model (PREM; Dziewonski & Anderson 1981).

Table 5. Earth model parameters associated with the disc load ensemble simulations. The viscoelastic parameterization of the Earth model is discretized in six radial layers; upper and lower crust, mantle lithosphere, asthenosphere, upper and lower mantle. The lower mantle extends down to the core mantle boundary (CMB; at the depth of 2763 km). Elastic layers are represented by a quasi-infinite viscosity of 10^{30} Pa s.

Layer	Depth (km)	Viscosity (Pa s)	Unique param. val.
West Antarctica			
Upper crust	20	10^{30}	1
Lower crust DL [yes/no]	30	$[10^{30}/10^{18}]$	2
Mantle lithosphere	[30, 90, steps of 10]	10^{30}	7
Asthenosphere	200	$[1 \times 10^{18}, 3 \times 10^{18}, 1 \times 10^{19}, 3 \times 10^{19}]$	4
Upper mantle	670	5×10^{20}	1
Lower mantle	CMB	2×10^{22}	1
Number of simulations West Antarctica			56
East Antarctica			
Crust	30	10^{30}	1
Mantle lithosphere	[150, 200]	10^{30}	2
Upper mantle	670	5×10^{20}	1
Lower mantle to CMB	CMB	2×10^{22}	1
Number of simulations East Antarctica			2
Elastic earth			
Crust and mantle to CMB	CMB	10^{30}	1
Total number of simulations			59

Later, in the joint inversion, the distribution of viscoelastic response functions is based on the Earth structure model of Priestley & McKenzie (2013). Priestley & McKenzie (2013) provide a global distribution of viscosity values up to a depth of 400 km, which is sampled at the location of the geodesic grid. We then define a threshold value for the viscosity (here, 10^{22} Pas) above which the Earth response is considered purely elastic and infer the associated thickness of the elastic lithosphere. [The impact on the final joint inversion estimate of changing](#)

the threshold value of 10^{22} Pas is presented in REGINA Paper II (Fig. S4 in Sasgen et al. 2017)

[R1_712].-Note that the Earth response in the equilibrium state only depends on the lithosphere thickness (independent of viscosity), which is therefore considered as the main Earth model parameters in the joint inversion. Further details are presented in REGINA paper II, Sasgen et al. (2017).

Gravity and displacement rate response functions

5.3 The calculated response functions for surface deformation (radial displacement) and gravity (geoid height change) are discretized along 1507 latitudinal points within the range $0 \leq \vartheta \leq 90$. Simulations are typically run over 2 kyr with a temporal resolution of $\Delta t = 10$ yr (plus two time steps with constant load thickness). For East Antarctic parameterizations, the simulation period was extended to 20 kyr due to the higher upper-mantle viscosities and associated slower relaxation. However, note that the ratio of geoid-height change versus radial displacement falls off to a factor of $1/e^2$ relative to the initial value [R1_700][R2_700]-after ca. 2 kyr of simulation (Appendix A.6, Fig. A.5). The forcing expected in central East Antarctica is an increase in accumulation towards present-day conditions after ca. 7 ka BP (van Ommen et al. 2004), justifying also the use of equilibrium kernels for East Antarctica. The time derivatives of the radial displacement y^u and of the geoid height change y^g are calculated with a central difference scheme.

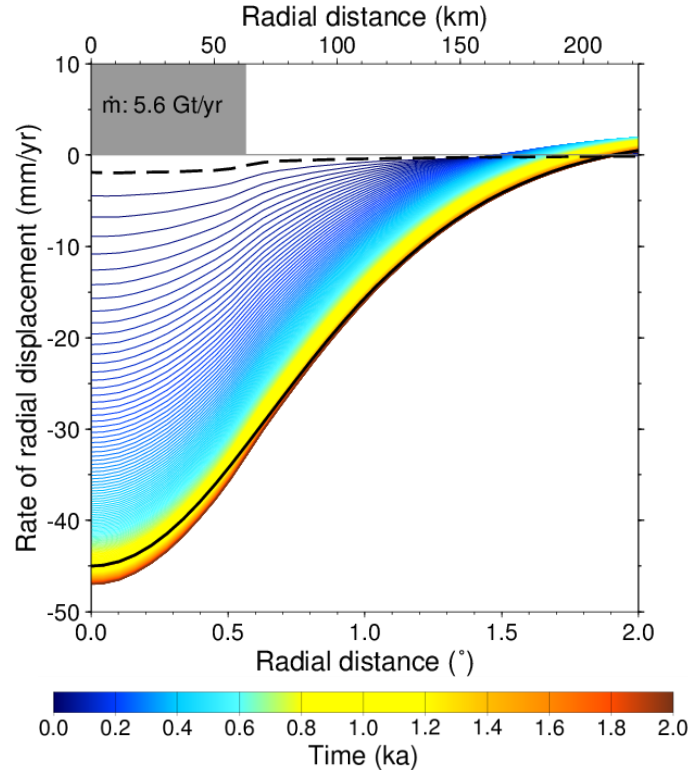


Figure 7. Displacement rates over the simulation period of 2 kyr, for an exemplary set of Earth model parameters ($h_L = 30$ km ; $\eta_{AS} = 1 \times 10^{18}$ Pa s). Shown is the load dimension (grey shading), as well as the instantaneous elastic response (dashed black line) and purely viscoelastic relaxation response only after 2 kyr and without a load change (solid black line). The other curves show the rates for the time epoch indicated by the color scale.

Examples of response functions to the loading detailed in Section 5.1 for the rate of radial displacement, y^u , and rate of geoid-height change, $-y^g \dot{e}$, are shown in Figs 7 and 8, respectively. Instantaneously, the increasing load, $\dot{\sigma}(t) = \text{const.}$, induces an elastic response that is characterized by subsidence and an increase in the direct gravitational potential (dashed lines in Fig. 7 and Fig. 8, respectively). This is the elastic response function adopted in the joint inversion. Note that the elastic response function will not differ between East and West Antarctica, as it is entirely based on the distribution of densities and elastic parameters provided by the PREM. As the load build-up continues, the instantaneous response is followed by the

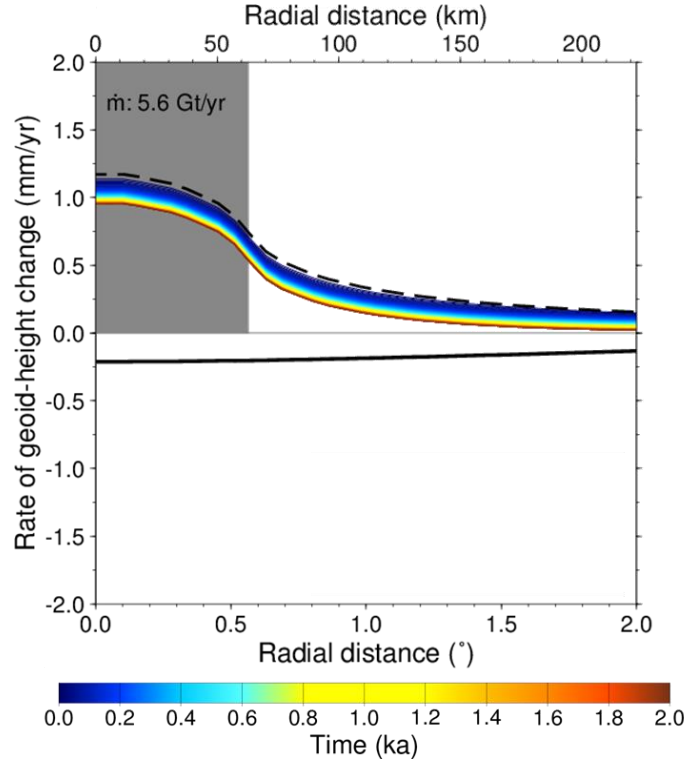


Figure 8. Same as Fig. 7, but for the rate of geoid-height change and Earth model parameters

$h_L = 90 \text{ km}$; $\eta_{AS} = 1 \times 10^{18} \text{ Pa s}$. Note the change in sign in the rate when the increase in direct gravitational attraction through load increase ceases after 2 kyr (black solid line vs. colored lines).

viscoelastic response, which depends in timing and magnitude on the underlying lithosphere and viscosity structure, further increasing the displacement rates, $y^u \dot{u}$ (blue to red lines in Fig. 7). The compensation by solid Earth deformation is reflected in the decreasing geoid rate, $y^g \dot{e}$ (Fig. 8). After a certain time, which depends on the value of the asthenosphere viscosity, a new dynamic equilibrium state is reached at which $\dot{u}$ and $\dot{e}$ do not change in time any more. In the last two time steps, the load is kept constant ($\dot{\sigma}(t) = 0$), and the responses in $\dot{u}$ and $\dot{e}$ are only caused by the relaxation of the Earth's viscoelastic deformation (solid black line in Figs. 7 and 8), which is the viscoelastic response function adopted in the joint inversion.

Discussion of effects of selected earth model parameterizations on GIA response

771 Fig. 9 shows the response of $\dot{u}$ for four end-member sets of Earth model parameters with
thick lithosphere, weak asthenosphere (*TkWk*: $h_L = 90 \text{ km}$; $\eta_{AS} = 1 \times 10^{18} \text{ Pa s}$), thick
5,4 lithosphere, strong asthenosphere (*TkSg*: $h_L = 90 \text{ km}$; $\eta_{AS} = 3 \times 10^{19} \text{ Pa s}$), thin
774 lithosphere, weak asthenosphere (*TnWk*: $h_L = 30 \text{ km}$; $\eta_{AS} = 1 \times 10^{18} \text{ Pa s}$) and thick
lithosphere, strong asthenosphere (*TnSg*: $h_L = 30 \text{ km}$; $\eta_{AS} = 3 \times 10^{19} \text{ Pa s}$), without a
ductile layer, DL. In this context, thick / thin and strong / weak refer to values in comparison to
777 the ‘average’ value of the ensemble for West Antarctica; an elastic lithosphere of thickness 90
km (here, ‘*Tk*’) is in the range of global average continental lithosphere usually applied in GIA
studies (e.g. Peltier, 2004), or that of East Antarctica (150 to 200 km). Fig. 10 shows the
780 response in $\dot{u}$ for the same end-member set of Earth model parameters with a DL included. It
should be stated that the Earth structure with $h_L = 30 \text{ km}$ and a DL is considered very extreme,
because in this case the ductile layer extends down to the asthenosphere and an elastic mantle
783 lithosphere is missing.

Fig. 9 and 10 show that for the weak asthenosphere ($\eta_{AS} = 1 \times 10^{18} \text{ Pa s}$), viscoelastic
deformation is visible already after one decade of loading (or unloading), leading to
786 considerably larger subsidence rates compared to the purely elastic case even on very short time
scales. For these Earth model parameters, a new dynamic equilibrium state is achieved within
a few centuries. The rates of subsidence in this equilibrium then primarily depend on the support
789 provided by the flexure of the elastic lithosphere.

For the extreme *TnWk* case, equilibrium rates of -45 mm/yr are achieved at the load
centre, and considerable subsidence of -20 mm/yr already occurs after ten years of loading
792 (Fig. 9). Increase in asthenosphere viscosity (*TnSg* case) reduces the viscous material transport

and leads to a slower adjustment towards the dynamic equilibrium state, which takes more than 1 kyr. It should be stated that in our definition of the ensemble parameters, reducing the

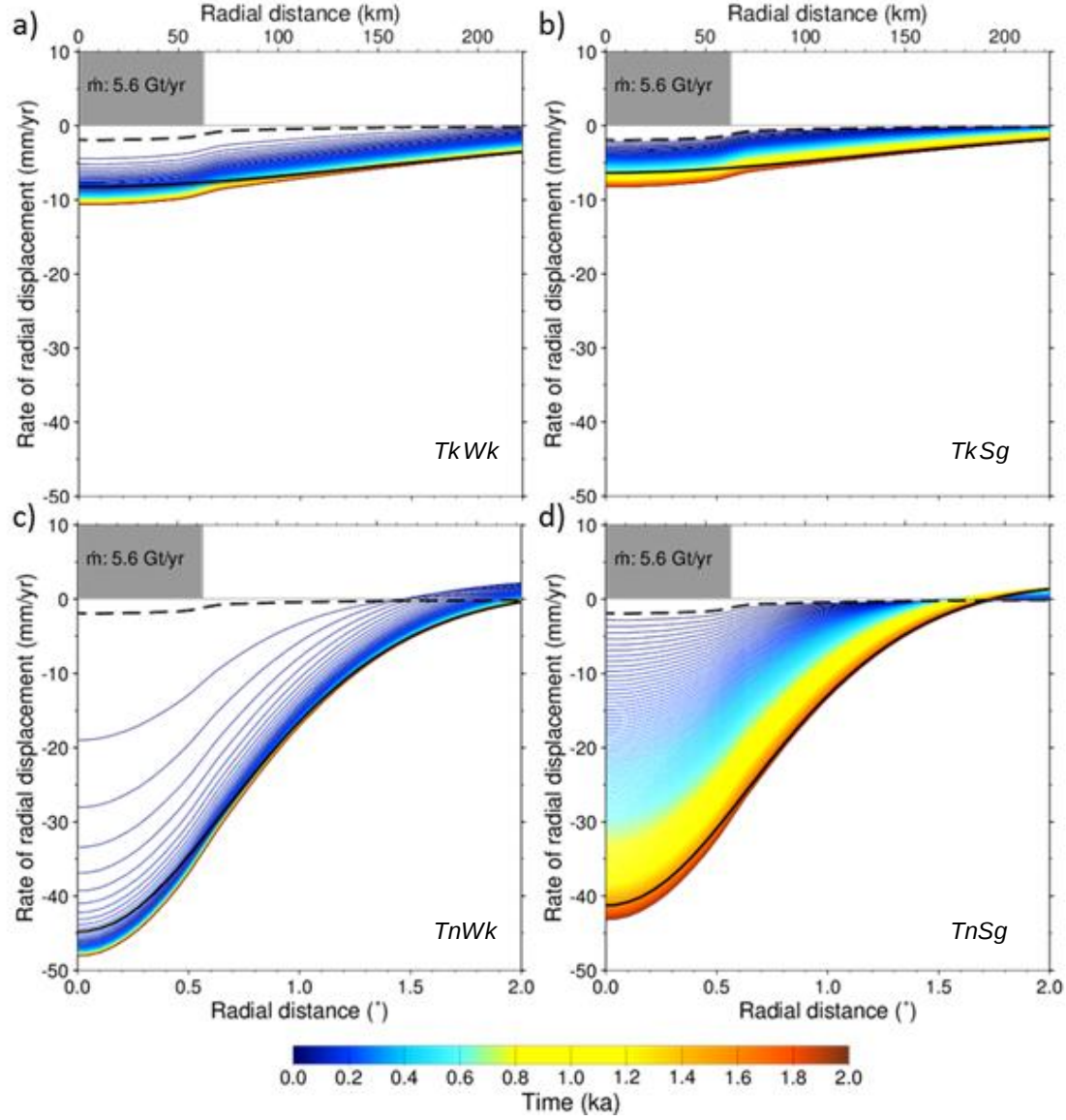


Figure 9. Same as Figure 7, but for four end-member sets of Earth model parameters, without a DL and lithosphere thickness / asthenosphere viscosity of a) $h_L = 90 \text{ km} / \eta_{AS} = 1 \times 10^{18} \text{ Pa s}$ ($TkWk$), b) $h_L = 90 \text{ km} / \eta_{AS} = 3 \times 10^{19} \text{ Pa s}$ ($TkSg$), c) $h_L = 30 \text{ km} / \eta_{AS} = 1 \times 10^{18} \text{ Pa s}$ ($TnWk$) and d) $h_L = 30 \text{ km} / \eta_{AS} = 3 \times 10^{19} \text{ Pa s}$ ($TnSg$).

795 lithosphere thickness in turn increases the thickness of the asthenosphere (bottom depth of asthenosphere is fixed), which facilitates lateral material transport inside the asthenosphere.

The consideration of the DL in the Earth structure causes a thinning of the effective elastic

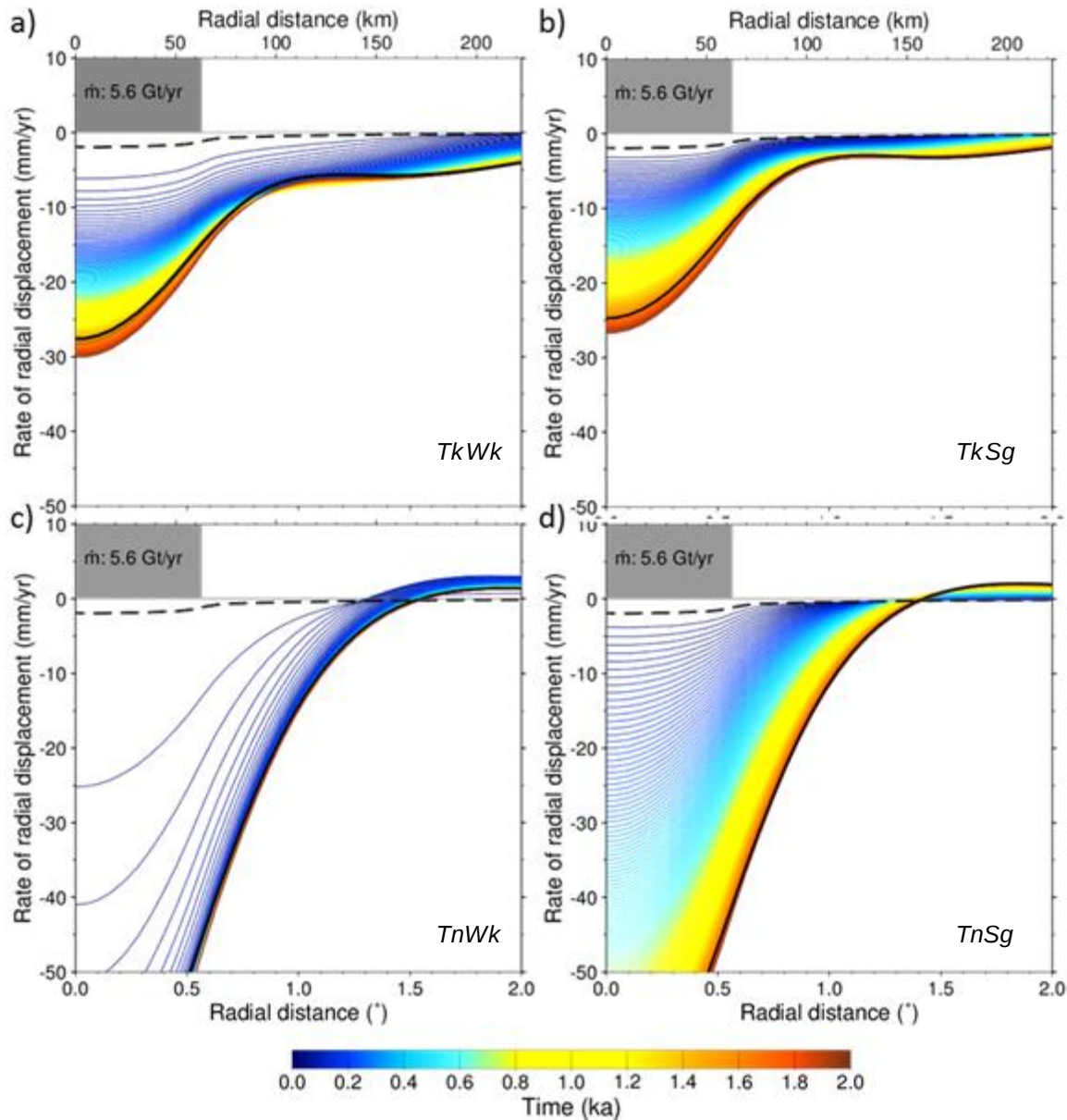


Figure 10. Same as Figure 9, a) TkWk, b) TkSg (b), TnWk (c) and TnSg (d), but with the Earth structure including a DL.

798 lithosphere. As a consequence, greater and more localized subsidence rates are produced for all sets of parameters (Fig. 10). Interestingly, in case of a thick elastic lithosphere (90 km), the

radial displacement exhibits a local minimum at around 120 and 160 km distance from the load
centre (Fig. 10), which is a consequence of the viscous material transport inside the ductile
layer. The maximum equilibrium rate of -76 mm/yr is achieved for the *ThWk* case with DL,
where the viscous deformation leads to rates of -25 mm/yr already after 10 yrs of loading.

Assumptions and limitations

Although the approach of modelling response functions to axisymmetric disc loads and
subsequently superposing them is very efficient in terms of the computational cost, this
simplification introduces some limitations. First, the superposition of response functions
representing different Earth structures neglects the transmission of stresses between these
regions — a problem that can only be resolved with fully three-dimensional solid Earth
modelling (e.g. van der Wal et al., 2015). The largest impact for the displacement rates is
expected in regions with lateral contrasts in lithosphere thickness and mantle viscosity such as
the Transantarctic Mountains. Second, the constant disc radius of about 63 km implies that
finer-scale deformation cannot be resolved. Although this resolution is adequate for interpreting
GRACE data (spatial half-wavelength of *ca.* 200 km) smaller-scale loading excitement may be
necessary for interpreting local GPS measurements near to the loading, particularly for the
elastic response to present-day glacial changes. Furthermore, the viscoelastic response
functions describe the Earth response in an equilibrium state for a constant rate of load change;
if the load exhibits more complex temporal variations, this assumption is violated. Finally, it is
assumed that the lithosphere thickness, upper- and lower-mantle viscosities are approximately
known. [R1_795].

The viscoelastic response functions and related ancillary data are directly accessible in the

Pangaea repository:

5.6

825 http://hs.pangaea.de/model/Sasgen-etal_2017/Viscoelastic_response_functions.zip

Viscoelastic kernels

Output files contain 1507 latitudinal points ($0 \leq \vartheta \leq 90$) covering a region greater than the

5.6.1

828 size of the Antarctic domain, as well 203 time steps of West Antarctica (213 time steps for East

Antarctica, because of extending the simulation period to 15 kyrs). The time derivative of the

radial displacement, u , is calculated with a central difference scheme, $y^u := [u(t+\Delta t/2) - u(t-$

831 $\Delta t/2)]/\Delta t$. The difference between two time steps is $\Delta t = 10$ yr. The same applies to the rate of

geoid-height change, y^g . Note that the load is constant during the last two time steps (no rate

of change); therefore, the kernels represent the viscoelastic relaxations only, without the

834 instantaneous elastic deformation or the direct gravitational attraction of the load.

The results are stored independently for the rheology of East and West Antarctica, the latter with and without a ductile layer in the elastic part of the lithosphere. The data are stored in a

837 Matlab® file format, which is also readable with GNU Octave

<https://www.gnu.org/software/octave/> .

- ‘Viscoel_response_WA_with_DL.mat’ – Response functions for West Antarctica

840 *with* ductile layer

- ‘Viscoel_response_WA_no_DL.mat’ – Response functions for West Antarctica

without ductile layer

- ‘Viscoel_response_EA_no_DL.mat’ – Response functions for East Antarctica

without ductile layer

- ‘Time_EA_rheo.mat / Time_WA_rheo.mat’ – Time [kyr] file related to response
846 file for East and West Antarctica
- ‘Coord_Co-Latitude.mat’ – Co-latitude [°] of the response functions

The response matrix summary the data as follows:

849 *West Antarctica:*

VE_WA_no_DL has the following entries, [HL, AV, LAT, TIME, VAR]

HL: Lithosphere thickness; 30 km, 40 km, ..., 90 km (7 entries)

852 AV : Asthenosphere Viscosity; $1 \times 10^{18} \text{ Pa s}$, $3 \times 10^{18} \text{ Pa s}$, $10 \times 10^{18} \text{ Pa s}$, $30 \times 10^{18} \text{ Pa s}$ (4 entries)

LAT: Latitude grid node, corresponding to file ‘Coord_Co-Latitude.mat’ (1537 entries)

855 TIME: Time, corresponding to file ‘Time_WA_rheo.mat’ (202 entries)

VAR: Variable type; 1: rate of radial displacement in mm/yr, 2: rate of geoid-height change in mm/yr.

858 The response kernels for *East Antarctica* are organized analogue, [HL, LAT, TIME, VAR]

HL: Lithosphere thickness; 150 km, 200 km (2 entries)

Note that the asthenosphere and upper mantle viscosity is constant at $5 \times 10^{20} \text{ Pa s}$ and
861 therefore has no entry.

The spectral resolution underlying these fields is spherical-harmonic cut-off degree 2048.

572 The user should apply an adequate smoothing filter when using for inverting GRACE gravity

864 fields. Filtered kernels are available upon request by the author.

Geodesic grid

The computation of the geodesic grid is not an original contribution of the authors, but

867 based on the grid generator of the ICON GCM project, <http://icon-downloads.zmaw.de/>. For

completeness, we provide the data set with disc locations based. An alternative resource for downloading geodesic grids at different resolutions in netCDF format can be found here:

870 <http://kiwi.atmos.colostate.edu/BUGS/geodesic/> .

The files format is:

vert-7.mask.cont_and_shelf.re.dat: Longitude [°], Latitude [°]

873 vert-7.mask.cont_and_shelf.re.proj.dat: X [km], Y [km], (projected coordinates, WGS-84,
Polar Stereographic, 71°S true latitude, 0°E central longitude)

876

Lithosphere thickness

5.6.3

The thickness of the elastic lithosphere at the locations of the geodesic grid for different
879 values of the viscosity threshold applied to the data set of Priestley & McKenzie, 2013.

lith_thresh_21.disc.txt (threshold 10^{21} Pa s, thicker lithosphere)

lith_thresh_22.disc.txt (threshold 10^{22} Pa s, lithosphere adopted in the GIA estimate)

882 lith_thresh_23.disc.txt (threshold 10^{23} Pa s, thinner lithosphere)

5.6.4 The 1175 entries correspond to the locations of the geodesic grid (Section 5.6.2).

Open source code for viscoelastic modelling

885 The open source software package SELEN allows the computation of the Maxwell-viscoelastic Earth response to user-defined ice sheet evolutions, in particular also a simplified disc-load forcing as presented in this paper. The program is downloadable at:

888 <https://geodynamics.org/cig/software/selen/>

6. CONCLUSIONS

In this paper, we have presented refined temporal linear trends of surface elevation, gravity

field change and bedrock displacement based on Envisat/ICESat (2003-2009), GRACE (2003-2009) and GPS (1995-2013.7), respectively. In addition, we have performed forward modelling of the viscoelastic response of the solid Earth to a disc-load forcing. These response functions are particularly suited to represent the distinct geological regimes of East and West Antarctica in the joint inversion of multiple satellite data. Similarly, the functions can be applied to the other geographical regions as well. The data and code necessary to reproduce our results, or apply our approach to a different problem, is provide at www.pangea.de, <https://doi.pangaea.de/10.1594/PANGAEA.875745> (follow link “View dataset as HTML”).

We have refined surface-elevation rates for the Antarctic ice sheet for the time interval 2003-2009 by combining Envisat and ICESat altimetry data. The straightforward ~~compositing~~ approach performs a grid-based comparison of the noise in the elevation rates obtained from Envisat and ICESat. For large parts of the ice sheet, the elevation rate is based on ICESat data, particularly, along for the rough terrain along coast, as well as close to the Pole (polar gap of Envisat). Envisat contributes in some low-relief areas in East Antarctica and along the Antarctic Peninsula, as well as along single spurious ICESat tracks. Thus, the composite elevation rates are maximized in terms of spatial coverage and minimized in terms of ~~the~~ uncertainties.

The GPS processing carried out as part of the RATES and REGINA projects has produced a comprehensive data set of Antarctic 118 GPS records, which, for continuous sites, spans a longer time interval (1995-2013) than those of previous studies (Thomas et al. (2011), 1995-2011; Argus et al. (2014), 1994-2012; Martín-Español et al. (2016b), 2009-2014). The ensemble processing done for the REGINA project has allowed us to assess the contribution of systematic error sources. In addition, for sites where there is potential doubt ~~over~~ about the quality of the metadata or the behaviour of the site, we have adopted a

‘conservative but realistic’ approach to assigning new confidence limits. The screening of GPS
915 data for outliers involved careful manual assessment, encompassing the review of measurement
logs and notes on problems in the field. The data quality is reflected in the uncertainty estimates
for the GPS rates, which therefore represents more reliable input data than GPS rates based on
918 processing without manual intervention. Note, however, SMB variations might also contribute
to the GPS uplift rates given that the time spans of these data vary [R2_867].

We have optimized the post-processing sequence for estimating the temporal linear trend
921 and its uncertainty in the GRACE gravity field solutions for the region of Antarctica. In
particular, we have derived optimal parameters for de-striping the monthly gravity fields over
Antarctica according to Swenson & Wahr (2006). In addition, we have removed de-trended
924 ~~interannual~~ SMB fluctuations [R1_476] from the GRACE time series, to obtain a more
representative uncertainty estimate based on the post-fit RMS residual. We have included
month-dependent weighting in the least-squares estimate of the gravity field rates to account
927 for the varying quality of the monthly GRACE solutions. The optimization of the de-correlation
filter of Swenson & Wahr (2006) to the signals expected in Antarctica reduced the residual
uncertainty and improved the reliability of inferred mass anomalies.

930 With the aim of joining the multiple satellite data using the knowledge of the geophysical
processes involved, we have calculated elastic and viscoelastic response functions of the solid
Earth. The viscoelastic response functions represent the gravity field change and surface
933 displacement to a disc-load forcing for a variety of Earth model parameters; particularly,
~~however,~~ values of mantle viscosity and lithosphere thickness strongly varying- between the
distinct geological regimes of West and East Antarctica.

936 In particular, we have investigated the effect of a ductile layer in the crustal lithosphere on

the viscoelastic rebound signature. We show that for moderate load changes of 0.45 m/yr water-equivalent (here, applied as disc load with a radius of ca. 63 km), uplift rates reach the cm/yr level within decades assuming asthenosphere viscosities $< 10^{19}$ Pa s and lithosphere thickness < 50 km; both plausible values for parts of West Antarctica. Including a ductile layer in the crustal lithosphere further attenuates the uplift rates and localizes the deformational response.

This suggests that GIA in West Antarctica may locally be a result of more recent, centennial load changes, most notably in the Amundsen Sea Embayment and in part of the Antarctic Peninsula (Nield et al. 2012). Similar conclusion were reached by Ivins & James (2005) and Nield et al. (2014), even though it is not possible to constrain the exact timing of the load from our approach [R1_921].

The advantage of the viscoelastic response kernels is that a meaningful ratio of rate of the gravity disturbance versus rate of the surface displacement [R1_923] is calculated for each choice of the Earth model parameters, avoiding the approximation with an average rock density (e.g. Riva et al. 2009; Gunter et al. 2014). Using the response functions allows us to reconcile GIA signatures with measurements of large bedrock uplift and small gravity field increase in the Amundsen Sea Embayment, associated with weak Earth structures. Clearly, the response functions adopted here represent only the viscoelastic equilibrium state and, thus, are considered only an intermediate step to full dynamic modelling of the GIA response. Nevertheless, this approximation represents a significant improvement of other joint inversion methods, as it bases the joint inversion on physically meaningful response kernels. With extra data on the past ice evolution, such as Paleo thickness rates, our approach can be expanded to address the temporal evolution as well.

In the succeeding paper REGINA part II (Sasgen et al. -2017submitted), we perform the

joint inversion for present-day ice-mass changes and GIA in Antarctica, based on the input data sets and viscoelastic response functions presented here. We validate our results using forward-modelling results and other empirical models, and show the impact on CryoSat-2 volume and GRACE mass balances, respectively. Note, however, that the post-processing methods and viscoelastic functions presented here are applicable also to other geographical regions with superimposed present-day mass change and GIA signatures.

7. DATA AVAILABILITY

The altimetry, gravimetry, GPS and viscoelastic modelling data used in this project are available at <https://doi.pangaea.de/10.1594/PANGAEA.875745> in the www.pangaea.de archive.

The data description and user documentation are given for each data type within the respective subsection of this paper (Sections 2 to 5).

AUTHOR CONTRIBUTION

Ingo Sasgen conceived, managed and summarized this study with support of Mark R. Drinkwater. Alba Martín-Español, Bert Wouters and Jonathan L. Bamber performed the altimetry analysis. Alexander Horvath, Martin Horwath and Roland Pail undertook the gravity field analysis, with contributions from Ingo Sasgen. Elizabeth J. Petrie and Peter J. Clarke ~~analysis-analyzed~~ and clustered the GPS data with critical input from Terry Wilson. Volker Klemann and Hannes Konrad performed the viscoelastic modelling, with contributions from Ingo Sasgen. All authors were involved in writing and reviewing this manuscript.

COMPETING INTEREST

The authors declare that they have no conflict of interest.

981 **ACKNOWLEDGEMENTS**

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987 Frederique Rémy for the Envisat data and Veit Helm for providing the AWI L2 CryoSat-2 re-tracked and corrected elevation measurements. The GPS data used was mainly downloaded from publically available archives. We acknowledge work done by the International GNSS

990 Service (Dow et al., 2009), UNAVCO and the Scientific Committee on Antarctic Research in maintaining such archives, together with the efforts of all the GPS site operators in collecting and making available the data, a particularly challenging task in Antarctica (see Table S2 for

993 more information on individual sources).

*A.1 ICESat campaigns and operation periods**Table A.1. ICESat 633 Level 2 data for the time span February 2003 until October 2009 used in this study.*

Start Date	End Date	Days in Operation	Laser Identifier
20/02/2003	29/03/2003	38	1AB
25/09/2003	19/11/2003	55	2A
17/02/2004	21/03/2004	34	2B
18/05/2004	21/06/2004	35	2C
03/10/2004	08/11/2004	37	3A
17/02/2005	24/03/2005	36	3B
20/05/2005	23/06/2005	35	3C
21/10/2005	24/11/2005	35	3D
22/02/2006	28/03/2006	34	3E
24/05/2006	26/06/2006	33	3F
25/10/2006	27/11/2006	34	3G
12/03/2007	14/04/2007	34	3H
02/10/2007	05/11/2007	37	3I
17/02/2008	21/03/2008	34	3J
04/10/2008	19/10/2008	16	3K
25/11/2008	17/12/2008	23	2D
09/03/2009	11/04/2009	34	2E
30/09/2009	11/10/2009	12	2F

We apply rates of firn compaction, h_{comp} , using output of the firn compaction model provided by Ligtenberg (2011), which is driven by RACMO2/ANT (Lenaerts 2010). However,

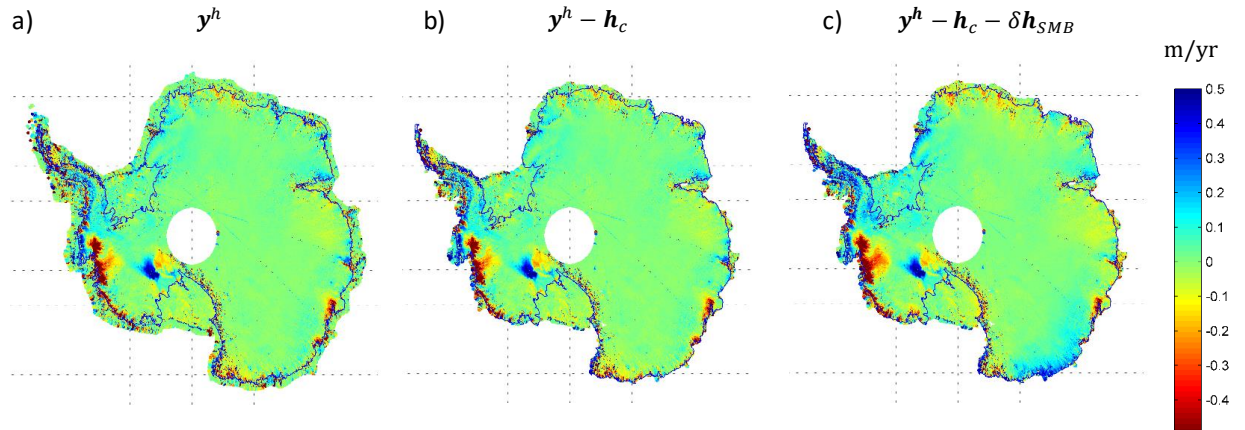


Figure A.1 Rate of elevation change y^h (m/yr), derived from a) ICESat/Envisat initial data, b) ICESat/Envisat minus firn ~~compaction~~ h_{comp} , and c) ICESat/Envisat minus firn compaction h_{comp} and modelled SMB anomalies δh_{SMB} .

1002 we do not apply a correction for anomalies in the surface-mass balance (SMB), δh_{SMB} , as e.g. undertaken by Gunter et al. (2014), due to the problem of defining an adequate reference period for the ice sheet. The impact of each correction is shown in Fig. A.1. Note that annual anomalies of the firn densification for the years 2003-2013 are available in the data archive
1005 http://hs.pangaea.de/model/Sasgen-et al_2017/Ice_sheet_topographic_change.zip

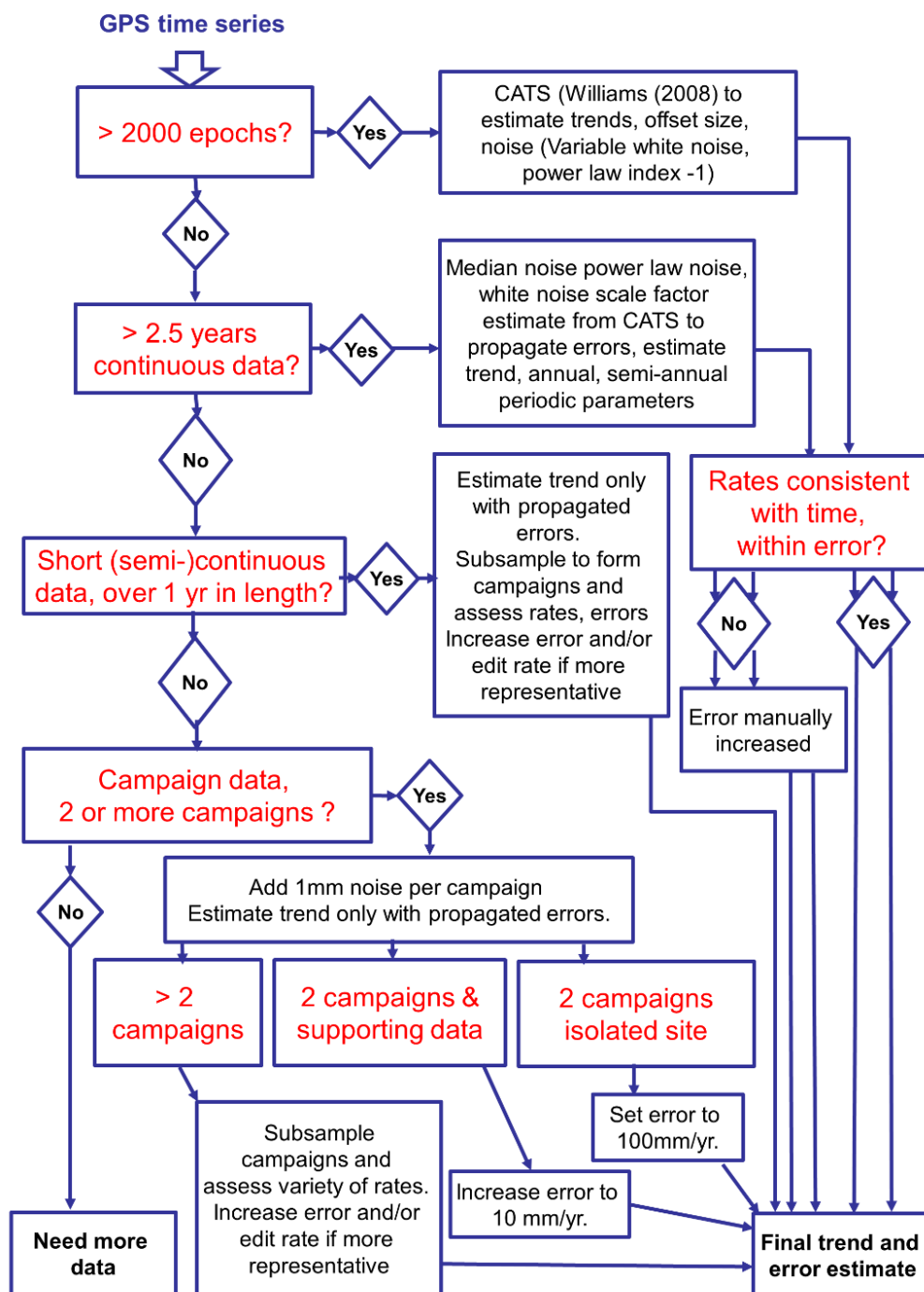


Figure A.2 Flowchart showing the estimation process for the temporal linear trends of the bedrock for Antarctic GPS site timeseries. After Petrie et al. (2016) (SCAR poster)

A.4 Uplift rates at all GPS site used in this study

Table A.2 GPS uplift rates for this study. The columns are: site name, estimated uplift rate y^u (mm/yr), estimated uncertainty σ^u (mm/yr), rate method, uncertainty method, approx latitude (dec. degrees), ~~approximately~~ longitude (dec. degrees). Methods are: cats: estimated by the CATS noise analysis software ('cats'), median uncertainty from CATS sites propagated ('prop'), manual intervention in rate due to potential systematic uncertainties ('rman') and manual intervention in uncertainties due to potential systematic errors

Site name	y^u	σ^u	Method y^u	Method σ^u	Lat. (°)	Lon. (°)	Doi/ data source or description
aboa	0.6	0.5	cats	cats	-73.04	-13.41	Finnish Geodetic Institute
brip	1.4	0.7	cats	cats	-75.80	158.47	doi:10.7283/T5W09473
buri	2.3	0.7	cats	cats	-79.15	155.89	doi:10.7283/T5RB72W7
cas1	1.5	0.2	cats	cats	-66.28	110.52	IGS: Dow et al. (2009)
cote	1.4	0.7	cats	cats	-77.81	162.00	doi:10.7283/T5GT5KGN
crar	0.7	0.4	cats	cats	-77.85	166.67	UNAVCO*
dav1	-1.6	0.6	rman	eman	-68.58	77.97	IGS: Dow et al. (2009)
dum1	-0.3	0.3	cats	cats	-66.67	140.00	IGS: Dow et al. (2009)
flm5	2.0	0.6	cats	cats	-77.53	160.27	doi:10.7283/T5V40SH6
ftp4	1.9	0.6	cats	cats	-78.93	162.56	doi:10.7283/T5B27SKD
maw1	-0.4	0.2	cats	cats	-67.60	62.87	IGS: Dow et al. (2009)
mcm4	0.8	0.2	cats	cats	-77.84	166.67	IGS: Dow et al. (2009)
min0	2.0	0.8	cats	cats	-78.65	167.16	doi:10.7283/T5TM78BX
ohi2	3.4	2.0	cats	eman	-63.32	-57.90	IGS: Dow et al. (2009)
palm	4.8	3.0	cats	eman	-64.78	-64.05	IGS: Dow et al. (2009)
ramg	2.4	0.8	cats	cats	-84.34	178.05	doi:10.7283/T51N7ZFR
rob4	1.1	0.5	cats	cats	-77.03	163.19	doi:10.7283/T5NC5ZG8
sctb	0.9	0.5	cats	cats	-77.85	166.76	doi:10.7283/T5CF9N6P
syog	1.1	0.2	cats	cats	-69.01	39.58	IGS: Dow et al. (2009)
tnb1	0.1	0.5	cats	cats	-74.70	164.10	Dubbini et al. (2010)
vesl	0.4	0.3	cats	cats	-71.67	-2.84	IGS: Dow et al. (2009)
a351	-0.9	1.8	prop	eman	-72.91	74.91	Geoscience Australia**
a368	-0.2	1.2	prop	eman	-74.29	66.79	Geoscience Australia**
arct	-0.1	4.4	prop	eman	-80.04	-80.56	SCARP***
art1	-3.1	10.0	prop	eman	-62.18	-58.90	Dietrich et al. (2004)
back	16.8	5.0	prop	eman	-74.43	-102.48	doi:10.7283/T5D21VWM
bean	2.1	4.3	rman	eman	-75.96	-69.30	doi:10.7283/T55Q4T6R

belg	-1.4	0.7	prop	prop	-77.87	-34.63	Dietrich et al. (2004)
benn	9.3	1.9	prop	prop	-84.79	-116.46	doi:10.7283/T5891447
berp	25.2	0.7	prop	prop	-74.55	-111.88	doi:10.7283/T54J0CC2
bhil	2.9	4.4	rman	eman	-66.25	100.60	Geoscience Australia**
bren	3.1	1.1	rman	eman	-72.67	-63.03	doi:10.7283/T52V2D7X
capf	4.0	1.4	rman	eman	-66.01	-60.56	doi:10.7283/T5XP736P
cjam	-2.3	100.0	prop	eman	-63.10	-62.72	SCARP***
clrk	3.6	1.4	prop	prop	-77.34	-141.87	doi:10.7283/T5MK6B6C
coat	-0.1	7.3	prop	eman	-77.81	162.00	Raymond et al. (2004)
crdi	2.1	0.6	prop	prop	-82.86	-53.20	doi:10.7283/T5C24TQS
cwal	0.4	100.0	prop	eman	-63.25	-62.18	SCARP***
dal1	4.9	34.4	prop	eman	-62.24	-58.68	Dietrich et al. (2004)
dall	-17.0	100.0	prop	eman	-62.24	-58.66	Dietrich et al. (2004)
devi	1.9	1.0	prop	prop	-81.48	161.98	doi:10.7283/T57942Z0
dupt	11.5	1.1	prop	prop	-64.81	-62.82	doi:10.7283/T5KD1W62
eacf	-4.8	15.0	rman	eman	-62.08	-58.39	Brazil
elph	6.3	100.0	prop	eman	-61.22	-55.14	SCARP***
esp1	5.6	100.0	prop	eman	-63.40	-57.00	Dietrich et al. (2004)
fall	4.8	1.3	prop	prop	-85.31	-143.63	doi:10.7283/T53J3B84
ferr	-5.5	31.0	rman	eman	-62.09	-58.39	Dietrich et al. (2004)
fie0	-0.9	1.9	prop	prop	-76.14	168.42	doi:10.7283/T5KK993F
flm2	3.8	11.7	rman	eman	-77.53	160.27	doi:10.7283/T53T9FHJ
fonp	13.5	1.8	prop	prop	-65.25	-61.65	doi:10.7283/T5668BG6
for1	-0.2	2.9	prop	eman	-70.78	11.83	Dietrich et al. (2004)
for2	-0.3	2.7	prop	eman	-70.77	11.84	Dietrich et al. (2004)
fos1	3.1	1.3	prop	eman	-71.31	-68.32	doi:10.7283/T54T6GF7
frei	-4.4	0.7	prop	prop	-62.19	-58.98	Bevis et al. (2009)
ftp1	-2.2	3.4	prop	eman	-78.93	162.56	doi:10.7283/T53T9FHJ
gmez	1.5	4.8	rman	eman	-73.89	-68.54	doi:10.7283/T58G8HT4
grw1	-7.0	8.6	prop	eman	-62.22	-58.96	Dietrich et al. (2004)
haa1	3.9	100.0	prop	eman	-77.04	-78.29	British Antarctic Survey
haag	6.1	1.1	rman	eman	-77.04	-78.29	doi:10.7283/T5FT8JB8
howe	0.6	1.1	rman	eman	-87.42	-149.43	doi:10.7283/T5ZW1J65
hown	3.9	0.8	prop	prop	-77.53	-86.77	doi:10.7283/T56971WH
hton	4.8	3.7	prop	eman	-74.08	-61.73	doi:10.7283/T5222RV6
hugo	0.9	1.3	prop	prop	-64.96	-65.67	doi:10.7283/T5FQ9TW3
iggy	2.3	1.1	prop	eman	-83.31	156.25	doi:10.7283/T5QC01T9
jnsn	4.0	1.7	prop	prop	-73.08	-66.10	doi:10.7283/T5SJ1HP1
Intk	4.6	3.1	rman	eman	-74.84	-73.90	doi:10.7283/T5J1017P
lply	2.0	8.1	rman	eman	-73.11	-90.30	doi:10.7283/T5DV1H50
lwn0	2.1	1.0	prop	prop	-81.35	152.73	doi:10.7283/T5T43RD8
mait	0.4	1.1	rman	eman	-70.77	11.74	Dietrich et al. (2004)
mar1	7.1	10.0	prop	eman	-64.24	-56.66	Dietrich et al. (2004)
mbl1	2.5	3.0	prop	eman	-78.03	-155.02	Donnellan & Luyendyk (2004) + doi:10.7283/T5CJ8BS7

mbl2	2.3	10.0	prop	eman	-76.32	-144.31	Donnellan & Luyendyk (2004)
mbl3	1.3	17.9	rman	eman	-77.34	-141.87	Donnellan & Luyendyk (2004)
mcar	3.7	1.4	prop	prop	-76.32	-144.30	doi:10.7283/T55D8Q41
mirn	24.4	100.0	prop	eman	-66.55	93.01	SCAR
mkib	4.7	2.6	rman	eman	-75.28	-65.60	doi:10.7283/T5D798HD
mtcx	-3.8	10.0	prop	eman	-78.52	162.53	Raymond et al. (2004)
ohg1	4.5	10.0	prop	eman	-63.32	-57.90	Dietrich et al. (2004)
ohig	4.0	0.7	prop	prop	-63.32	-57.90	Former IGS: Dow et al. (2009)
pal1	8.1	10.0	prop	eman	-64.77	-64.05	Dietrich et al. (2004)
patn	4.8	0.7	prop	prop	-78.03	-155.02	doi:10.7283/T5PC30PX
pece	0.7	4.2	prop	eman	-85.61	-68.56	doi:10.7283/T5930RG1
pra1	4.2	10.0	prop	eman	-62.48	-59.65	Dietrich et al. (2004)
prat	-9.6	100.0	prop	eman	-62.48	-59.65	doi:10.7283/T5M32T21 ,
prtt	-5.0	100.0	prop	eman	-62.48	-59.67	doi:10.7283/T5K35RZP
reyl	151.3	300.0	prop	eman	-62.20	-58.98	SCARP***
rob1	5.4	5.1	prop	eman	-77.03	163.19	doi:10.7283/T5M32T21 ,
robi	8.7	1.5	prop	prop	-65.25	-59.44	doi:10.7283/T5K35RZP ,
rot1	6.5	10.0	prop	eman	-67.57	-68.13	doi:10.7283/T5057D6V ,
rotb	5.0	0.4	prop	prop	-67.57	-68.13	doi:10.7283/T53T9FHJ
roth	5.5	1.4	prop	prop	-67.57	-68.13	Nield et al. (2014)
sdly	-0.3	1.4	prop	prop	-77.14	-125.97	SCAR
sig1	23.0	100.0	prop	eman	-60.71	-45.59	doi:10.7283/T56M34Z7
smr1	0.5	10.0	prop	eman	-68.13	-67.10	doi:10.7283/T56M34Z7
smrt	1.2	0.9	prop	prop	-68.13	-67.10	IGS: Dow et al. (2009)
sppt	12.9	100.0	prop	eman	-64.29	-61.05	doi:10.7283/T5S46Q7F
sugg	4.7	1.3	rman	eman	-75.28	-72.18	Dietrich et al. (2004)
svea	1.3	1.1	prop	prop	-74.58	-11.23	Dietrich et al. (2004)
thur	-1.2	2.5	rman	eman	-72.53	-97.56	Dietrich et al. (2004)
tomo	47.7	20.3	rman	eman	-75.80	-114.66	Alfred Wegener Institute / Instituto Antartico Argentina
trve	2.5	5.6	rman	eman	-69.99	-67.55	Bevis et al. (2009)
ver1	0.3	100.0	prop	eman	-65.25	-64.26	doi:10.7283/T5CV4G1M
ver3	-6.2	100.0	prop	eman	-65.25	-64.26	Sjoberg et al. (2011)
vnad	4.4	1.1	prop	prop	-65.25	-64.25	doi:10.7283/T5862DRZ
w01b	1.4	10.0	prop	eman	-87.42	-149.44	doi:10.7283/T5BZ64B0
w02b	2.3	10.0	prop	eman	-85.61	-68.56	doi:10.7283/T5NS0RZ9
w03a	-1.4	10.0	prop	eman	-81.58	-28.40	SCAR
w03b	1.7	10.0	prop	eman	-81.58	-28.40	SCAR
w05a	2.3	10.0	prop	eman	-80.04	-80.56	doi:10.7283/T52F7KQ1
w05b	7.4	10.0	prop	eman	-80.04	-80.56	doi:10.7283/T5445JTQ

w06a	-2.2	100.0	prop	eman	-79.63	-91.28	
w07a	3.3	100.0	prop	eman	-80.32	-81.43	
w08a	-1.5	100.0	prop	eman	-75.28	-72.18	
w09a	2.2	100.0	prop	eman	-82.68	-104.40	
wasa	0.6	3.2	prop	eman	-73.04	-13.41	Sweden
whn0	2.2	0.9	prop	prop	-79.85	154.22	doi:10.7283/T5R49P2M
whtm	7.7	0.8	prop	prop	-82.68	-104.39	doi:10.7283/T5ZP44DZ
wiln	4.9	0.9	prop	prop	-80.04	-80.56	doi:10.7283/T53F4MX9
* https://www.unavco.org/projects/project-support/polar/geodetic/benchmarks/sites/crar.html (accessed 1 June 2017)							
**Geoscience Australia GNSS archive at ftp://ftp.ga.gov.au/geodesy-outgoing/gnss/ as of 1 June 2017. See also Brown, N. and Woods, A., 2008. Antarctic Geodesy 2006 – 2007 Field Report. Geoscience Australia, Record 2009/32. 77pp.							
*** SCARP Campaign datasets, doi:10.7283/T5T151QB , doi:10.7283/T59P2ZZD , doi:10.7283/T5K35RZP . Also see https://gcmd.nasa.gov/records/GCMD_JCADM_USA_SCARP.html							

Table A.3. Comparison of ‘prop,eman’ GPS uplift rates for this study with rates from other studies.

	REGINA		Thomas et al. 2011		Argus et al. (2014)		Wolstencroft et al. (2015)	
	y^u	σ^u	y^u	σ^u	y^u	σ^u	y^u	σ^u
a351	-0.9	1.8	0.8	1.3	1.1	3.5		
a368	-0.2	1.2	0.4	1.0				
for1	-0.2	2.9	-1.4	0.8				
for2	-0.3	2.7	2.1	0.9				
fos1	3.1	1.3	2.1	0.4	2.9	1.2	3.9	1.1
ftp1	-2.2	3.4	2.1	2.8				
hton	4.8	3.7						
mb11	2.5	3.0	0.6	1.5				
mb12	2.3	10	0.2	4.1			6.4	0.9
rob1	5.4	5.1	7.5	2.6				
w01a(-howe)	-0.3	10	-2.5	1.7	0.9	1.2		
w01b	1.4	10	-3.1	1.7				
w02a(-pece)	0.3	10	2.8	1.2	-1.2	1.9		
w02b	2.3	10	0.5	1.9				
w03a	-1.4	10	-3.2	1.8	-1.1	2.4		
w03b	1.7	10	-1.7	1.8				
w04a	3.7	100	3.0	1.1				
w05a	2.3	10	3.5	2.0				
w05b	7.4	10	5.3	1.2				
w06a	-2.2	100	-2.2	2.4	-4.7	4.4		
w07a	3.3	100	3.3	2.1	4.6	3.1		
w08a(b/sugg)	-1.5	100	1.3	1.3				
w09a	2.2	100	4.5	2.6				

A.5 Choice of GRACE cut-off degree and biasing

In this study, we identify GRACE coefficients of CSR RL05 up to-degree and order 50
 1014 appropriate to yield the most robust gravity field rates over Antarctica. Figure A.3 provides
 another indication based on the degree-power spectrum of the geoid rates. It is visible that GFZ
 RL05 and CSR RL05 are very similar up to degree and order 50, where the power spectra show
 1017 minima. For higher degrees, however, the power of the gravity field recovered with GRACE

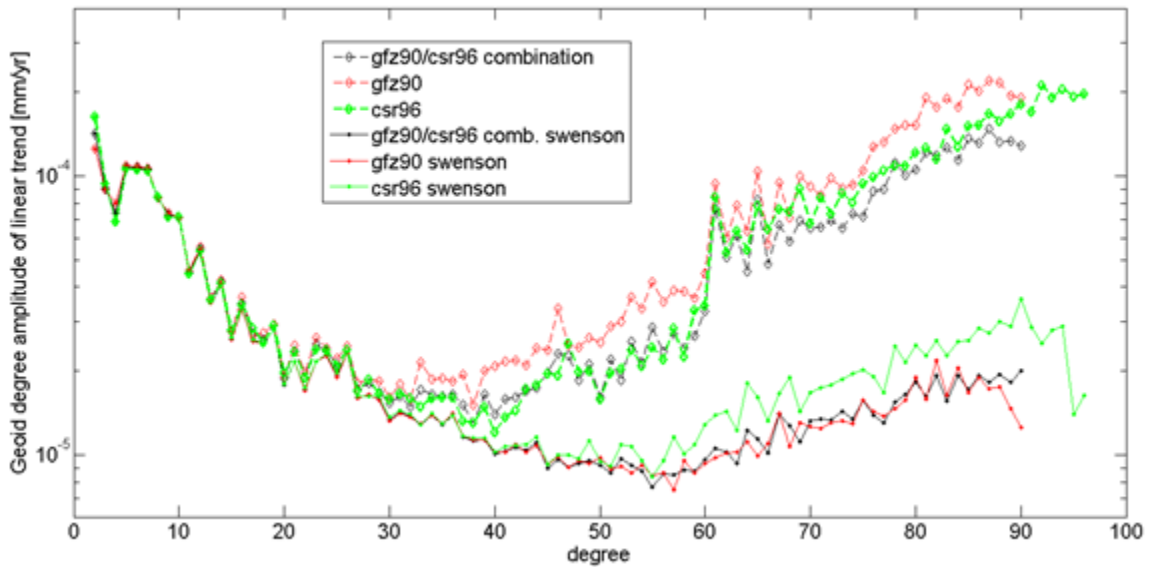


Figure A.3: Degree-amplitude spectrum of the rate of geoid-height change (mm/yr) for unfiltered (diamond-dashed lines) and for Swenson-filtered (solid lines) solutions. Red: GFZ; green: CSR; black combination of GFZ and CSR with equal weights.

increases due to increasing noise, for the unfiltered coefficients particularly faster for GFZ
 RL05 than for CSR RL05.

1020 The filtering of the GRACE gravity fields was optimized for reducing noise over
 Antarctica. The effect on the RMS uncertainties is shown in Fig. 3. Additionally, Fig. A.4

presents the difference of between the GRACE rates filtered only with a Gaussian smoothing
 1023 filter of 200 km, and additionally with the optimized Swenson filter. It is visible that the
 differences in the rate of geoid-height change and the associated rate of equivalent water-height
 change, respectively, show a stripe-like noise pattern. This suggests that the de-stripping is
 1026 superior over conventional Gaussian smoothing, even at high latitudes, where GRACE ground-
 track spacing is very dense. It is also important to note that the filter does not introduce any
 magnitude bias, or changes the spectral content of the gravity field rates, which is important
 1029 when applying only Gaussian smoothing of 200 km (without Swenson filtering) to the altimetry
 data set and response kernels.

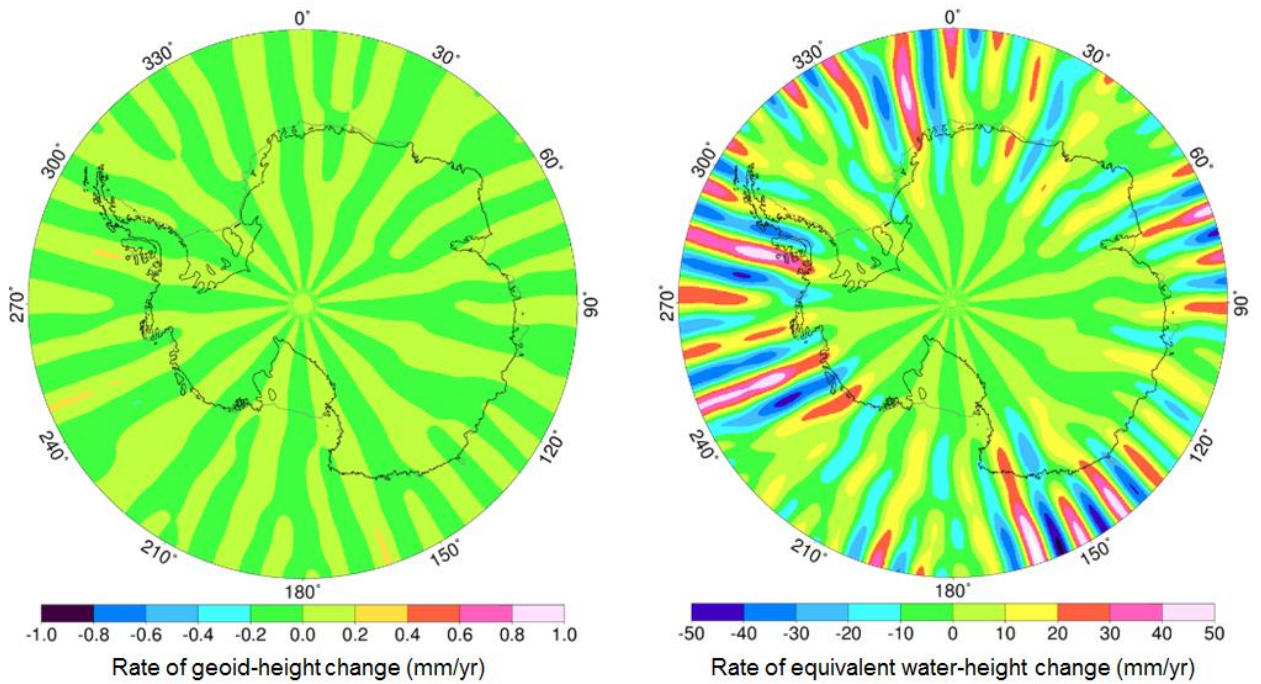


Figure A.4: Spatial rate of geoid-height change (left) and rate of equivalent water-height change (right) (mm/yr) for the difference between the GRACE trends processed by Gaussian smoothing of 200 km and the optimal Swenson filter & Gaussian smoothing. The solutions are CSR RL05, the spherical-harmonic cut-off degree is 50.

The viscoelastic response kernels employed (Section 5) describe the viscoelastic equilibrium state for the forcing with a disc load of constant radius and constant rate of mass increase (likewise mass loss). We neglect transitional changes of the solid Earth for load changes that have not reached the equilibrium state in terms of geoid-height change and surface displacement. Although, the deformation and gravity signature in equilibrium eventually only depends on the lithosphere thickness, the time to reach the equilibrium is controlled by the viscosity parameters chosen. Fig. A.5 shows the evolution of the standardized ratio of the geoid-

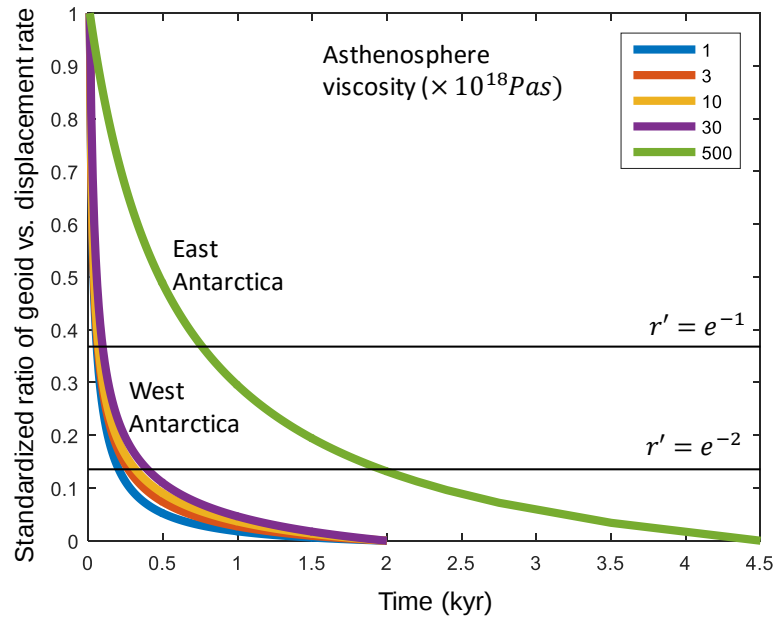


Figure A.5: Standardized ratio of the rate of geoid-height change versus the rate of radial displacement for different values of the asthenosphere viscosity. Note that the ratio is calculated at the load center.

height change vs. surface displacement over time, calculated as
 1041 $r' = [r(t) - r(t = t_{tmax})] / \max[r(t) - r(t = t_{tmax})]$, where $r = y^g(t)/y^u(t)$ is

evaluated at the load centre. It is visible that for the weaker West Antarctic rheology (asthenosphere viscosity between 1×10^{18} Pa s and 3×10^{19} Pa s) r' falls to $1/e^2$ within the 500 yr. For East Antarctica (1×10^{20} Pa s), $r' = e^{-2}$ is reached within 2 kyrs. With this quasi-stationary solution approach, the inference on the timing of the past ice mass change is limited to an upper limit in terms of magnitude, and a lower limit in terms of load duration; a similar ratio is achieved by a thinner lithosphere thickness, which has not reached viscoelastic equilibrium state, and earlier load changes are fully relaxed, respectively.

A.7 Assessment of SMB fluctuations on GPS uplift rates [R2_867]

We assess the impact of SMB fluctuations on the uplift rate at the GPS station locations

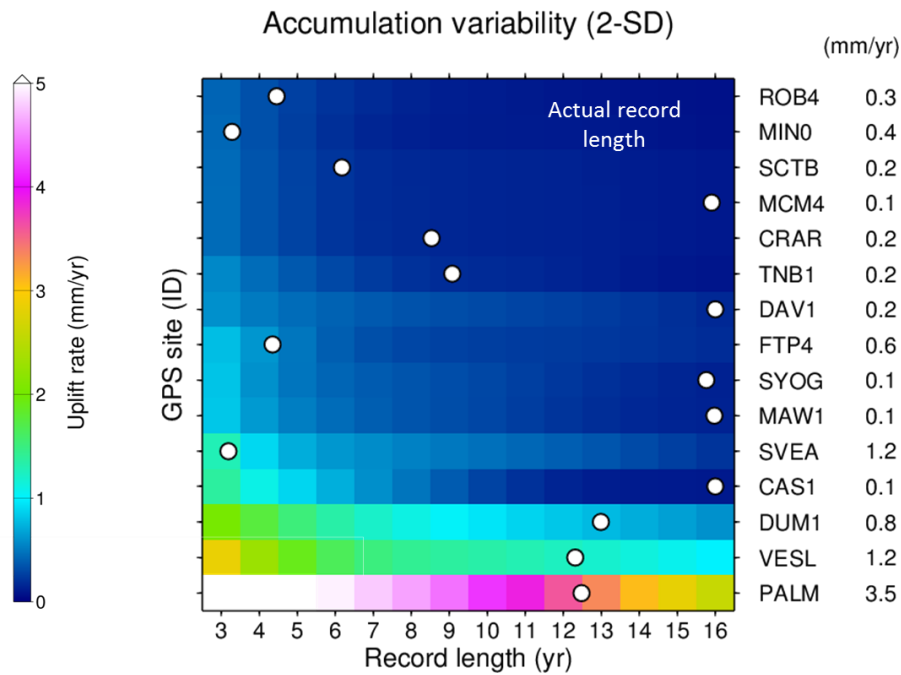


Figure A.6: Standard deviation (2-sigma) of the uplift rates caused by accumulation variability for different GPS stations and time periods. It is visible that the uncertainty decreases with record length; for most regions, trend uncertainties are below 0.4 mm/yr for the actual GPS record length.

using the modelled SMB of RACMO2 for the years 1979-2010. We compute the elastic

deformation related to cumulative monthly SMB, de-trended for the entire simulation period

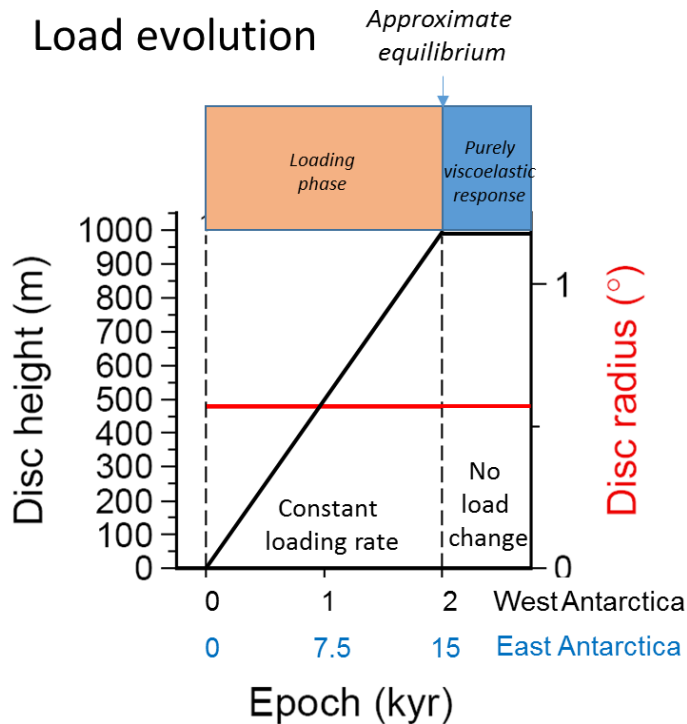
1053 1979-2010. We then estimate the temporal linear trends at the GPS station locations for a moving window of varying width from 3 to 16 years. Then, for each window width, we estimate the standard deviation of the apparent trend induced by SMB for selected stations (Fig. A.7).

1056 Typically, the uncertainty of uplift rate due to SMB variability is below 0.4 mm/yr for the actual GPS record length. An exception is PALM, which is located on the Antarctica Peninsula - a region with annual accumulation of up to 4 m/yr equivalent water height. Here, even after 12 years of measurements, GPS uplift rates are likely to contain accumulation signals of 4 mm/yr.

A similar effect of the SMB fluctuations is expected at VESL.

A.7A.8 Load evolution for the viscoelastic response functions [R2_867]

1062 The load increases, with a fixed radius, at a constant rate of *ca.* 5.6 Gt/yr until an approximate equilibrium state is reached; 2 kyr for West Antarctica and 15 kyr for East Antarctica (Fig. A.7). Then the load is applied without a change to obtain the purely viscoelastic response of the Earth model, i.e. without direct gravitational attraction of the load and the instantaneous elastic response. The associated Earth response constitutes the viscoelastic response functions adopted in the joint inversion.



[R2_1014] Figure A.7: Load function applied to obtain the viscoelastic response functions.

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