The Cryosphere Discuss., 7, 3497–3541, 2013 www.the-cryosphere-discuss.net/7/3497/2013/ doi:10.5194/tcd-7-3497-2013 © Author(s) 2013. CC Attribution 3.0 License.



This discussion paper is/has been under review for the journal The Cryosphere (TC). Please refer to the corresponding final paper in TC if available.

Empirical estimation of present-day Antarctic glacial isostatic adjustment and ice mass change

B. C. Gunter¹, O. Didova¹, R. E. M. Riva¹, S. R. M. Ligtenberg², J. T. M. Lenaerts², M. A. King^{3,4}, M. R. van den Broeke², and T. Urban⁵

¹Delft University of Technology, P.O. Box 5048, 2600 GA, Delft, the Netherlands

²Institute for Marine and Atmospheric Research Utrecht, Utrecht University, P.O. Box 80.005, 3508 TA, Utrecht, the Netherlands

³School of Civil Engineering and Geosciences, Newcastle University, Newcastle upon Tyne NE1 7RU, UK

⁴School of Geography and Environmental Studies, University of Tasmania, Hobart 7001, Australia

⁵The University of Texas at Austin, Center for Space Research, 3925 West Braker Ln, Ste 200, Austin, TX 78759, USA

Received: 16 June 2013 – Accepted: 28 June 2013 – Published: 16 July 2013

Correspondence to: B. C. Gunter (b.c.gunter@tudelft.nl)

Published by Copernicus Publications on behalf of the European Geosciences Union.





Abstract

This study explores an approach that simultaneously estimates Antarctic mass balance and glacial isostatic adjustment (GIA) through the combination of satellite gravity and altimetry data sets. The results improve upon previous efforts by incorporating ⁵ reprocessed data sets over a longer period of time, and now include a firn densification model to account for firn compaction and surface processes. A range of different GRACE gravity models were evaluated, as well as a new ICESat surface height trend map computed using an overlapping footprint approach. When the GIA models created from the combination approach were compared to in-situ GPS ground station displace-¹⁰ ments, the vertical rates estimated showed consistently better agreement than existing GIA models. In addition, the new empirically derived GIA rates suggest the presence of

- GIA models. In addition, the new empirically derived GIA rates suggest the presence of strong uplift in the Amundsen Sea and Philippi/Denman sectors, as well as subsidence in large parts of East Antarctica. The total GIA mass change estimates for the entire Antarctic ice sheet ranged from 53 to 100 Gt yr⁻¹, depending on the GRACE solution
- ¹⁵ used, and with an estimated uncertainty of ±40 Gt yr⁻¹. Over the time frame February 2003–October 2009, the corresponding ice mass change showed an average value of -100±44 Gt yr⁻¹ (EA: 5±38, WA: -105±22), consistent with other recent estimates in the literature, with the mass loss mostly concentrated in West Antarctica. The refined approach presented in this study shows the contribution that such data combinations
 ²⁰ can make towards improving estimates of present day GIA and ice mass change, par-
- ticularly with respect to determining more reliable uncertainties.

1 Introduction

25

Over the past decade, there has been general consensus within the glaciological and geodesy communities that the ice sheet of Antarctica is currently experiencing a significant loss in ice mass, on the order of tens to hundreds of gigatons ($1 \text{ Gt} = 10^{12} \text{ kg}$) per year (Chen et al., 2006; Rignot et al., 2008; Horwath and Dietrich, 2009; Jacob





et al., 2012; Shepherd et al., 2012). The impact of this ice loss is substantial, as the resulting change in sea level and ocean currents have global environmental and societal consequences. For Antarctica, the mass change estimates from the Gravity Recovery and Climate Experiment (GRACE) have large uncertainties associated with them due

- to a number of inaccurately known input models (Thompson et al., 2004; Seo et al., 2008; Gunter et al., 2010). Of these, the dominant error comes from the inaccurate knowledge of glacial-isostatic adjustment (GIA) (Gunter et al., 2009), which is the deformation of the solid earth due to the slow return of mantle material that was displaced by the changing ice load during the last ice age (which peaked ~ 21 kyr ago). The fact
- that the GRACE mission can only observe the total gravitational accelerations acting on the mission's twin satellites means that GRACE measurements are not able to distinguish between accelerations due to mass changes caused by the loss/gain of ice from those accelerations caused by the GIA-induced surface uplift. As a result, the effects of GIA are typically removed in the data processing with modeled values; how-
- ever, the uncertainty in current GIA models is > 50 % of the mass change estimates derived from GRACE (Velicogna and Wahr, 2006), i.e., it is also at the scale of tens of gigatons of equivalent ice mass change. This is due to the very sparse (in both space and historical time) geophysical and climatological data available for Antarctica, which are required to constrain historical changes in ice history and hence GIA models. This
- ²⁰ uncertainty in the GRACE estimates makes the monitoring and prediction of current mass loss trends much less reliable, and highlights the need to make improvements in the determination of the GIA and ice mass change signals.

An alternative to forward modelling present-day GIA is to estimate present-day uplift (Wahr et al., 2000). One approach to accomplish this involves incorporating estimates of ice elevation (satellite altimetry) and bedrock uplift (GPS). The addition of the altimetry products is particularly important because they track absolute *volume* changes,

25

as opposed to the absolute *mass* change measured by GRACE. While these are two completely different observables, they are complementary and permit the separation of the GIA and ice mass changes, given knowledge of ice/rock densities. This is possible





because the large density contrast between rock and ice make the altimetry products much more sensitive to the volume changes associated with ice mass changes, while the gravity products are much more sensitive to the mass changes associated with GIA. For example, a 1 cm uplift due to GIA would be barely detectable by satellite

- altimetry, but the corresponding (large) mass change from this small uplift would be clearly observable from GRACE. Previous studies have demonstrated the feasibility of this approach (Wahr et al., 2000; Velicogna and Wahr, 2002), with the first real-data combination produced by Riva et al. (2009). As a joint estimation problem, GIA and ice mass change trends are simultaneously computed, creating a self-consistent set of es-
- timates. In addition, as a data-driven approach, the errors of the input data sets can be used to generate realistic and spatially varying uncertainties of the resulting GIA and mass change estimates through standard error propagation techniques. In the time since the first real-data combination was achieved, several major improvements to the methodology and data sets have taken place, resulting in new estimates of Antarctic GIA and ice sheet mass balance that this paper seeks to highlight.
 - New contributions of this study include the use of updated data from GRACE and the Ice Cloud and land Elevation Satellite (ICESat) mission, which have both recently undergone a complete reprocessing that has noticeably improved the data quality compared to previous releases. For the GRACE data, a range of both unconstrained and
- regularized solutions are evaluated to better categorize the impact that different processing strategies can have on the results. The ICESat data was processed using a recently developed technique involving the use of overlapping footprints (OFPs). The approach was first developed by Slobbe et al. (2008) for a study of the Greenland ice sheet, but has not been applied previously to Antarctica. The OFP approach was ex-
- panded and improved for this study, and made use of the latest release of ICESat data (R633). The OFP method has many benefits over standard repeat-track and crossover techniques, and is particularly well-suited for Antarctica due to the high density of laser shots available. The technique also allows for the independent determination of





the systematic campaign biases which are present in the ICESat surface height rates, a critical item when considering long-term ice sheet volume changes.

Another important contribution of this study is the use of a firn densification model (FDM) that estimates spatial and temporal variations in firn layer temperature, depth

- and mass, and which also accounts for penetration, retention and refreezing of meltwater. Most similar studies to date have relied on a simplified density assumption to convert altimetric heights to mass, often just a constant value. For many regions which experience highly variable accumulation rates, as well as glacial thinning and high GIA rates, a constant surface density assumption significantly misrepresents the true sur-
- face characteristics. Finally, the GIA component of the data combination was compared against vertical height displacement measurements collected from a network of dozens of permanent GPS ground stations. Such comparisons allowed the various data combinations to be evaluated, both with each other, as well as against state-of-the-art GIA models.
- ¹⁵ This paper will assess the impact of these new developments on the GIA and ice mass change estimates of Antarctica, as well as provide an outlook for future investigations. As will be shown later, the resulting GIA models compare favorably with other modelled estimates, but do suggest some areas, such as the Amundsen Sea Sector, may be experiencing much higher uplift rates than previously thought. These regions
- ²⁰ of higher GIA uplift in turn affect the total ice mass change, resulting in mass balance estimates that are lower than most found in the literature for Antarctica.

2 Methodology

25

The underlying methodology used to combine the altimetric and gravimetric data sets is adapted from earlier work by Riva et al. (2009), summarized here for convenience. In short, the technique relies on the fact that satellite altimetry measurements primarily observe surface processes, such as accumulation and ablation, whereas the mass change measurements from satellite gravimetry are sensitive to the mass change





of both GIA and surface processes. By exploiting the difference in density between ice/snow, ρ_{surf} , and the solid earth, ρ_{rock} , the following relationship can be established which relates the vertical height rates of GIA, \dot{h}_{GIA} , to the mass, height, and density values for a given location.

$$5 \quad \dot{h}_{\text{GIA}} = \frac{\dot{m}_{\text{GRACE}} - \rho_{\text{surf}} \cdot \dot{h}_{\text{ICESat}}}{\rho_{\text{rock}} - \rho_{\text{surf}}}$$

A 400 km Gaussian smoothing is applied to ensure the various components in Eq. (1) have the same spatial resolution, but this is only done after elements with equivalent resolution are first combined. For example, the multiplication of the surface density and ICESat height rates is done before applying the smoothing, since these two grids have approximately the same spatial resolution. How the surface and rock densities are treated will be covered in the next section, as well as the consideration of elastic effects.

3 Data sets

Several data sets are used to perform the combination, as well as validate the results. ¹⁵ For this study, the total mass change estimates were derived from GRACE and the surface height trends derived from ICESat. The properties of the surface, i.e., surface mass balance (SMB) and firn layer changes, were taken from Antarctic climate and firn densification models. The solid earth densities were assumed to be 4000 kg m⁻³ for land, transitioning to 3400 kg m⁻³ under the ice-shelves, consistent with Riva et al.

(2009). Only the surface heights and surface processes over the grounded ice sheet were used, since these changes do not contribute to mass change over the ice shelves, while the GRACE and solid earth densities were used over both land and ocean regions. The time period under investigation covers the entire ICESat mission period, from February 2003 to October 2009.



(1)



3.1 Gravimetry

The GRACE mission has collected data on the time-variable nature of Earth's gravity field since its launch in March 2002. A number of research centers produce monthly gravity field models, using different processing methodologies. A range of gravity mod-

- ⁵ els are examined in this study, including those generated by the University of Texas at Austin Center for Space Research (CSR), the GeoForschungsZentrum (GFZ), and Delft University of Technology (TUD). Both RL04 and RL05 solutions were evaluated when available, as well as regularized solutions using various techniques. Degree one coefficients were added to all solutions using values generated from the approach of
- ¹⁰ Swenson et al. (2008) (using RL05 GRACE data), and the $C_{2,0}$ harmonics were replaced with those derived from satellite laser ranging (Cheng and Tapley, 2004). For the RL04 models, the secular trends that are removed from select zonal coefficients were restored, as these rates are believed to mostly represent the effects of GIA.

For all solutions except the Delft Mass Transport (DMT-1b) models produced at TUD

- (Liu et al., 2010), which use a specialized method for the trend estimation (Siemes et al., 2013), a linear trend was estimated for each harmonic coefficient across the entire time series of monthly models (again, covering only the time period from February 2003 to October 2009). The trend was co-estimated with a bias, annual periodic, and tidal S2 (161 day) periodic terms. Earlier studies (Seo et al., 2008) indicated that
- additional aliasing may occur at other tidal frequencies, e.g., K2 (1362.7 days); however, an investigation into these showed that only S2 showed a noticeable influence on the long-term trends over Antarctica, particularly for the newer RL05 solutions. Evidence for this is provided in Fig. 1, which shows the amplitude of the estimated K2 periodic signal in units of equivalent water height (EWH) computed from both a repre-
- sentative GRACE solution (CSR RL04 DDK3 in this case) and the 330 km Gaussian smoothed surface mass balance (SMB) estimates from the RACMO2 climate model (see Sect. 3.3). The fact that the majority of the areas with larger amplitudes in the GRACE solution (Fig. 1a) are spatially correlated with those seen in the SMB esti-





mates (Fig. 1b), suggests that the signal seen in the GRACE data is genuine mass variability at this frequency.

For the unconstrained CSR and GFZ solutions, the estimated long-term trend was then de-striped using an approach similar to that outlined by Swenson and Wahr
⁵ (2006), but with the filtering parameters described by Chambers and Bonin (2012). Even though these parameters were created with ocean applications in mind, the choice of polynomial degree (5th order for RL04, 4th order for RL05) and starting degree and order (12 for RL04, 15 for RL05) were found to perform better than other alternative parameters tested, and were therefore used for this study. No de-striping
¹⁰ was applied to any of the regularized solutions.

Several sets of regularized solutions were included in the analysis, to examine the potential impact that different spatial filtering techniques may have on the final results. This included the Wiener-type filter described by Kusche (2007), which was applied to the RL04 (DDK3) and RL05 (DDK5) solutions for both the CSR and GFZ. A recently developed set of filtered solutions developed by Save et al. (2012), utilizing an L-curve

- ¹⁵ developed set of filtered solutions developed by Save et al. (2012), utilizing an L-curve method with Tikhonov regularization, were also evaluated (named here "CSR Reg"). Finally, for the DMT1-b solutions, the anisotropic filtering method developed by Klees et al. (2008) is applied after the long-term coefficient trend is estimated (along with bias, annual, and S2 terms).
- Not all solutions are generated to the same spherical harmonic degree and order, and truncating them to the lowest common resolution, e.g., 60 × 60, can noticeably degrade their quality. Therefore, most solutions were left in their native resolution when possible, as indicated in Table 2. For the GFZ unconstrained solutions, however, leaving them at their original resolution resulted in the presence of a significant amount of
- noise in the trends, requiring a small degree of additional Gaussian smoothing after the trend fitting and de-striping process. The amount of Gaussian smoothing for these unconstrained GFZ solutions was kept at a minimum in an effort to maximize the signal content in the solutions, and was approximately 200 km.





In total, 10 different GRACE solutions were evaluated, with the geographical plots for a representative selection of these cases shown in Fig. 2. The plots for all 10 solutions can be found in Fig. S1 of the Supplement. As can be seen, the trends for nearly all solutions are quite similar; however, some variations can be seen in terms of magnitude and resolution of finer features. As will be seen later, these variations will have an important influence of the outcome of the estimated GIA and ice mass change values from the data combinations.

3.2 Altimetry

The ICESat mission was the first Earth-orbiting laser altimeter and, while no longer operational, it was able to collect valuable information on the long-term surface height change of Antarctica over a period which directly coincides with when the gravity data from GRACE was collected. The surface height change trends used for this study were computed using the latest release of ICESat data (R633), and were computed using an approach involving overlapping footprints (OFPs), similar to that described by Slobbe et al. (2008) for Greenland. This is the first time the OFP approach has been applied to

- ¹⁵ et al. (2008) for Greenland. This is the first time the OFP approach has been applied to Antarctica. The technique is well suited for observing long term trends at a high spatial resolution, since the co-location of the laser shots used in the height change estimates do not rely on interpolation and/or surface approximations inherent in other techniques, such as cross-over and repeat-track analysis. The technique is particularly useful for
- height change studies in Antarctica due to the high density of laser shots from the near-polar orbit of ICESat. The data processing uses a set of editing criteria to remove outliers, and estimates a custom set of inter-campaign biases, the details of which are outlined below.

3.2.1 Overlapping footprint approach

²⁵ The basic principle of the OFP approach is illustrated in Fig. 3a, where an overlapping footprint pair is defined as any two individual ICESat laser shots whose ground foot-





print have at least some overlapping area. The technique described by Hughes and Chraibi (2011) was used to determine if the ellipses characterizing any two laser shots physically overlapped, as well as the percentage of overlap if they did. The two laser shots can come from any two ICESat campaigns and are not restricted to ascending or descending tracks; however, OFPs within the same campaign are excluded in this study due to the higher uncertainties they introduce. The height change (d*h*) from an OFP pair can be divided by the time difference (d*t*) of the two shots to compute a height change rate (d*h*/d*t*). To reduce the impact of slope effects, the degree of overlap can be used as an editing criterium so that the centers of the two footprint ellipses are as close as possible. This option will be used in the next section when estimating the inter-campaign biases.

To find potential OFPs, the maximum radius for each campaign (the footprint radius is not constant over time) is first determined based on all available shots. Any two shots whose centers are closer than the total distance of their respective campaign radii were campaidered OFP campaidered.

- radii were considered OFP candidates. Depending on the shape and orientation of the two laser footprints, it is possible that two footprints can be close enough that their circumscribing circles overlap without the actual footprint ellipses overlapping, as illustrated in Fig. 3b. These neighboring shots in Fig. 3b offer the same information content as those in Fig. 3a, since the shot centers are still within twice the (maximum)
- ²⁰ semi-major axis distance from each other in both scenarios. As such, the ICESatderived surface height trends used later include these neighboring shots, termed here "near-neighboring" (NNs) shots, to distinguish them from the physically overlapping OFPs. Approximately 151 million total OFP/NN shots were used, of which 76 million were NNs. Unless otherwise noted, future references to OFPs will imply that both OFP and NN pairs are included.

The original full set of R633 laser shots used in the analysis was edited in accordance with that outlined in Urban et al. (2013), and included the use of standard quality flags, as well as other criteria such as the use of only single peak shots, a maximum gain value of 150, and a maximum co-elevation angle of 0.45° . In addition, any dh/dt values





computed from individual OFP/NN pairs greater than 12 m yr^{-1} were excluded, as this is assumed larger than most known glacial thinning or ablation processes. A linear trend in time (without annual terms) was fit across all (d*h*, d*t*) pairs satisfying the editing criteria within 20 km × 20 km area blocks, with the uncertainties determined by scaling

- the formal error from the least squares regression by the estimated variance of unit weight (EVUW) computed from the post-fit residuals (Urban et al., 2013). This EVUW scaling also helps to account for errors due to any seasonal variations that might be present. The estimated d*h*/d*t* values from this process are shown in Fig. 4a, with the corresponding uncertainties in Fig. 4b. When integrated only over the grounded ice
 sheet, using the boundaries defined by Zwally et al. (2012), the total volume change
- is approximately $-109 \pm 68 \text{ km}^3 \text{ yr}^{-1}$. Most of the uncertainty is located in the Antarctic Peninsula and Transantarctic Mountains, and is caused by both a combination of poor sampling and steep topography.

3.2.2 Estimation of campaign biases

- ¹⁵ The ICESat laser shots are known to have a systematic bias in them that can introduce cm-level errors if neglected (Gunter et al., 2009). To minimize the effect of these campaign-specific biases, an approach to estimate their magnitude was adopted using a low-precipitation zone (LPZ) in East Antarctica, in the same line as Gunter et al. (2010) and Riva et al. (2009). The rationale is that East Antarctica is one of the driest
- ²⁰ places on Earth, and has relatively flat topography, so very little surface height change is expected to take place in this region. The exact region used to estimate the campaign biases is shown in Fig. 5, and was derived using output from the regional climate model to be discussed in Sect. 3.3. In particular, the region corresponds to an area that is estimated to have less than 21.9 mm EWH yr⁻¹ of average yearly solid precipitative
- ²⁵ flux, a value chosen by trial-and-error to create a continuous low-precipitation zone that is sufficiently isolated from areas of steep topography.





Using this LPZ, a select set of ICESat measurements were used to precisely estimate the biases. One of the advantages of the OFP approach is that the degree of overlap between two laser shots can be tuned to a specific range. With a higher overlap criterion, the shots have more coincident ground coverage, but at the expense of reducing the number of OFPs used, since fewer shots will satisfy the criterion. For the determination of the campaign bias, it was important that the shots involved in the OFPs have high levels of overlap, to reduce any potential errors caused by changes in the surface topography within the footprint area. As such, the OFPs used for the bias estimates were required to have at least an 80 % overlap with one or both of the laser shots. In addition, any d*h* values greater than 1 m were considered outliers and excluded (this overlap and d*h* editing criteria were only used for the determination of the campaign biases). A time series of the least median of squares of the remaining d*h* values were created, using each campaign as a reference, for a total of 18 different bias profiles (gray lines in Fig. 6). For example, the bias profile using campaign 3b

- ¹⁵ would consist of dh values from the OFP combinations 1a-3b, 2a-3b, 2b-3b, etc. The mean of each profile was removed before taking the median value at each time step (dark blue). To investigate the influence of possible accumulation or compaction in the LPZ, the firn densification model (Sect. 3.3) was used to predict any surface change of the firn. The model did suggest a small surface lowering over the LPZ, on the order of
- $_{20}$ -0.15 cm yr⁻¹ (magenta), and this value was removed from the median values to arrive at the final campaign bias estimates (cyan) shown in Table 1. Standard deviations for each campaign bias are also provided in the table. There is a small amount of variation in the biases from campaign to campaign, but the overall trend of the bias estimates is 1.58 ± 0.08 cm yr⁻¹. This is generally consistent with the earlier 2.0 cm yr⁻¹ bias es-
- timates computed using mean sea surface comparisons (Gunter et al., 2009), as well as other estimates in the literature (Urban and Schutz, 2005; Siegfried et al., 2011; Ewert et al., 2012); however, these previous bias estimates used earlier ICESat data releases, so are not directly comparable to the estimates of this study. Also note that because the mean was removed from the individual profiles, the values represent the



bias offset with respect to the mid-point of the ICESat mission lifetime. The estimated biases were removed from the individual laser shots involved in the height change calculation for each OFP, i.e., before the trend-fitting by blocked area discussed in the previous section.

5 3.3 Climate data

In order to separate the deformation caused by surface processes (ice, firn) from those of the solid-earth (GIA), both the volume and mass change of the ice sheet needs to be known. There are many complex processes at work that complicate the determination of these quantities, including regional variations in temperature, accumulation, and firn compaction. To account for these, a firn densification model (FDM) developed by Ligtenberg et al. (2011) is used that is forced at the surface with realistic 6 hourly climate output from the regional atmospheric climate model RACMO2 produced by Lenaerts et al. (2012). This FDM model accounts for compaction of the firn over time, and is used in conjunction with the time-varying estimates of the total SMB from RACMO2 to estimate the mass change of the firn layer. Figure 7 shows the total automate beight rate, and accounts for the firn layer.

- surface height rate, and associated uncertainties, as derived from the FDM model over the study period. It is important to note that the FDM of Fig. 7b only represents the surface height changes of the firn, and does not reflect changes due to either the solid earth or ice dynamics. Furthermore, the mass change of the firn over time, \dot{m}_{firn} , is derived from the SMB, which is a separate product generated from RACMO2, although
 - both the FDM and SMB estimates are inherently linked.

Two basic assumptions were made to account for height differences that were found to exist between the altimetry measurements and the FDM. First, the uncertainties of the height estimates derived from the ICESat and FDM data sets were defined over each grid cell as

$$\sigma_h = \sqrt{\sigma_{\rm ICESat}^2 + \sigma_{\rm FDM}^2},$$

25



(2)

using the standard deviations shown in Figs. 4 and 7. In order to convert the volume changes derived from the ICES at data into mass, the density of the volume change needs to be known. Because RACMO2 only models firn processes, any negative differences between the ICES at and FDM surfaces that was greater than $2\sigma_b$ for any given grid cell were assumed to be the result of ice dynamics (glacier thinning), and 5 the density assigned to this volume loss was that of ice (917 kg m⁻³). Similarly, any positive height differences beyond the $2\sigma_h$ level were attributed to an underestimation of SMB by RACMO2, and given a density closer to that of snow using a static density profile similar to that of Kaspers et al. (2004). The justification for the densities assigned to positive height differences is shown in Fig. 8. This plot shows the derived 10 density (Fig. 8c) computed from those regions where the (GRACE - SMB) differences were greater than $20 \text{ kgm}^{-2} \text{ yr}^{-1}$, and the (ICESat – FDM) were greater than $6 \text{ cm} \text{ yr}^{-1}$. The resulting densities in those areas are predominantly in the 350-600 kg m⁻³ range, with a mean value of 396 kgm^{-3} , suggesting that the use of snow densities for these positive height anomalies is reasonable. The only exception to the rules of positive or 15 negative height differences was for the region of the Kamb Ice Stream in West Antarc-

- tica, where no ice discharge takes place, and the positive height change is assumed to be a build-up of ice (glacier thickening) with a density of 917 kg m⁻³. If the height differences between ICESat and the FDM fell within the $2\sigma_h$, the height measurements were
- ²⁰ considered to be within the uncertainty of the data sets, and the volume/mass of the difference was neglected. It is important to note that these assumptions only deal with potential residual signal observed between ICESat and the FDM. The majority of the surface mass changes come directly from the SMB estimates (i.e., \dot{m}_{firn}) derived from RACMO2. As such, the utilization of the SMB and FDM in the combination approach ²⁵ required a modification of Eq. (1),

$$\dot{h}_{\rm rock} = \frac{\dot{m}_{\rm GRACE} - [(h_{\rm ICESat} - h_{\rm firn}) \cdot \rho_{\alpha} + \dot{m}_{\rm firn}]}{\rho_{\rm rock} - \rho_{\alpha}}$$



(3)

where

$$\rho_{\alpha} = \begin{cases} 917 \, \text{kg m}^{-3}, & \text{if } \dot{h}_{\text{ICESat}} - \dot{h}_{\text{firn}} < 0 \\ & \text{and } |\dot{h}_{\text{ICESat}} - \dot{h}_{\text{firn}}| > 2\sigma_h \\ \rho_{\text{surf}}, & \text{if } \dot{h}_{\text{ICESat}} - \dot{h}_{\text{firn}} > 0 \\ & \text{and } |\dot{h}_{\text{ICESat}} - \dot{h}_{\text{firn}}| > 2\sigma_h \\ 0, & \text{otherwise.} \end{cases}$$

(4)

It should be noted that, in the combination, an elastic correction is made for the load represented by the right-hand-side of the numerator in Eq. (3), i.e., from the surface mass variations computed from ICESat and the SMB data. This correction ultimately has a negligibly small influence on the final results, but is done in the interest of completeness.

3.4 Vertical site-displacements

The output from the combination represented by Eq. (3) is a vertical rate associated with GIA over Antarctica, hereafter referred to as the empirical rates. While a map of these values can be compared against other GIA models in an attempt to assess its accuracy, an alternative approach is to compare the empirical rates with those observed by ground-based GPS stations. For this study, vertical displacement rates for up to 35 GPS stations were used for the comparisons to be shown later. The processing of the GPS displacements followed that of Thomas et al. (2011), and includes both seasonal

- and permanent stations. Likewise, the trends in vertical displacement for these stations have been corrected for elastic deformation effects using ice mass flux estimates for 2006 (where available, otherwise values for 2000 were used) taken from Rignot et al. (2008). Furthermore, the time frames for the GPS trend analysis do not necessarily
- overlap with the GRACE and ICESat data sets; however, this is not as critical for the GPS rates, since GIA is assumed to evolve at near-constant rates over relatively short geologic time frames (e.g., decades), and elastic effects are removed.





Following a similar approach as Whitehouse et al. (2012), comparisons with the GPS data were done by computing the weighted root-mean-square of the residuals (WRMS) between the vertical empirical or modelled rates and those observed from the GPS stations (i),

5 WRMS =
$$\sqrt{\frac{\sum w_i \cdot (\dot{h}_{GIA} - \dot{h}_{GPS})^2}{\sum w_i}}$$

where the weight,

$$w_i = \frac{1}{\sigma_{\rm GPS}^2 + \sigma_{\rm GIA}^2}$$

10

incorporates both the uncertainty of the individual GPS stations (σ_{GPS}) as well as the uncertainty of the GIA estimate (σ_{GIA} , described later in Sect. 5.1) at the station location. The uncertainties for the GPS stations ranged from < 0.3 mm (indicated by large

symbols in Fig. 11), 0.3 to 1.5 mm (medium symbols), and > 1.5 mm (small symbols). Additional details of the comparisons to the GPS displacements will be discussed later in Sect. 5.

4 GIA bias correction

- One of the early observations from the combination results was the presence of a mmlevel bias in the empirically derived GIA rates. Earlier investigations into this suggested that the cause of this bias could come from several sources (Gunter et al., 2010). For example, if there exists a secular trend in the geocenter motion (degree 1 coefficients), then any Z-component rate would be unaccounted for in this analysis. The uncertainty
- $_{\rm 20}$ in the determination of C_{2,0} (related to Earth's oblateness) from GRACE has been recognized for some time, and is why values from satellite laser ranging are still recommended to be used in place of those found in the official data products. Any trend



(5)

(6)



or other inconsistency in the coefficient values used for $C_{2,0}$ would translate into a rate bias for Antarctica. Errors in the ICESat campaign bias could also contribute to the differences seen, as would any inconsistency in reference frames used by the various data sets. It is important to note that every 1 mm of offset in the GIA rates would translate into approximately 50 Gt yr⁻¹ of solid-earth mass change, so while the magnitude of the offset is small, its impact on the solution can be significant.

To address the issue of potential offsets in the solutions, use was made again of the LPZ shown in Fig. 5. The rate of GIA in this region is expected to be very small, i.e., significantly less than the unknown bias offset caused by the various sources discussed above. As such the LPZ is used as a calibration area, where both the mean surface

- ¹⁰ above. As such, the LPZ is used as a calibration area, where both the mean surface height change (Sect. 3.2.2) and subsequent GIA is assumed to be zero in that region. In terms of a practical implementation, this is accomplished by computing the mean value over the LPZ of the smoothed \dot{h}_{GIA} values generated from Eq. (3). This mean value, termed the "LPZ GIA bias", is then removed from all GIA values uniformly. The magnitude of the LPZ GIA bias for each case investigated is shown in Table 2.
- Calibrating the solutions to the LPZ provides a simple but effective way to deal with the range of bias contributors (i.e., geocenter, reference frame, campaign bias, etc.) that are currently not known at the mm-level or less. The LPZ bias correction also allows each solution to be compared more equivalently, since the bias contributors which are removed may be different for each case. The primary consequence for using the LPZ in this way is that the GIA solutions created become regional to Antarctica, and therefore cannot be used to estimate global GIA effects, such as the contributions from the Northern Hemisphere.

5 Combination results

5

The geographical plots of a select set of the resulting GIA models created from the LPZ calibration approach are shown in Fig. 9 (the full set of plots can be found in Fig. S2 of the Supplement). The corresponding mass change values are provided in Table 2, ex-



pressed in total gigatons per year $(Gtyr^{-1})$ and divided into regions representing East Antarctica (EA), West Antarctica (WA), and the total Antarctic Ice Sheet (AIS), following the grounding lines defined by Zwally et al. (2012) (extended outwards by 400 km to account for the smoothing). Once the GIA mass change rates were obtained, they were

⁵ subtracted from the total mass change estimated from GRACE to derive a corresponding ice mass change value, also shown in Table 2. Since the earlier LPZ GIA bias was estimated using all components in Eq. (3) (i.e., including SMB, surface heights, and GRACE), in order to compute the ice mass change estimates in a consistent manner, a separate LPZ bias was estimated for only GRACE, i.e., the "LPZ GRACE bias", the values of which are shown in Table 2 in units of EWH. Again, this is done to ensure that the mean value of mass change over the LPZ is set to zero.

5.1 Uncertainty analysis

Errors in the GIA and ice mass change estimates from the combination approach were computed using formal error propagation techniques, resulting in what are believed to

¹⁵ be realistic error uncertainties. Where possible, uncertainties provided for the individual input sources were used, while for other sources certain assumptions were made, the details of which are outlined below.

For the GRACE data, the uncertainties were derived using formal error propagation techniques and the publicly available calibrated errors provided by the CSR for each ²⁰ monthly solution, along with the uncertainties provided with the degree 1 and $C_{2,0}$ coefficients. The calibrated errors were first propagated into equivalent water height (EWH) using the functional model described by Wahr et al. (1998). These errors were in turn propagated onto the trend component, using the same parameterization described earlier in Sect. 3.1. Though not shown, the GRACE errors do have a latitudinal element to them.

²⁵ dependency to them, but for Antarctica they are relatively uniform at approximately 1– 1.5 mm yr⁻¹ EWH. It is important to note that the errors for GRACE are assumed to be the same for all solutions evaluated, and which is a source of future refinement for the





combination approach. The errors for the ICESat trends made use of the EVUW-scaled uncertainties discussed in Sect. 3.2, which are shown in Fig. 4b. The FDM provided has associated uncertainties, as shown in Fig. 7b; however, the SMB information used to determine $\dot{m}_{\rm firn}$ in Eq. (3) do not have estimated uncertainties, so a standard de-

- viation of 10% of the value for each grid point was used as a conservative estimate, similar to that employed by Rignot et al. (2008). For the rock densities, a standard deviation representing 100 kg m⁻³ of the value for each grid point was assumed. Likewise, for the surface density value used when treating the differences between ICESat and the FDM, a 10% standard deviation was also used per grid point.
- ¹⁰ The aforementioned input data uncertainties were then formally propagated using Eq. (3) to generate total uncertainties for the three major mass change quantities (total mass change, GIA-related mass change, and ice mass change) for EA, WA, and the AIS. The uncertainties for the AIS were computed by taking the square-root of the sum-of-squares of the EA and WA uncertainties. This is consistent with the analysis
- ¹⁵ done as part of the recent Ice Sheet Mass Balance Intercomparison Exercise (IMBIE) (Shepherd et al., 2012), and is justified by the fact that the primary signals in EA and WA are sufficiently separated that their errors can be treated as independent of each other. These results are summarized on the last row of Table 2, with the geographical variation of the uncertainties shown in Fig. 10. The GIA uncertainties $(1-\sigma)$ over the
- ²⁰ AIS are 40 Gtyr⁻¹, with the regions of higher uncertainties located in the areas most expected, such as the Amundsen Sea Sector (ASE) and Wilkes/Adelie Land (WA). The ice mass change estimates are relatively well defined for WA at 22 Gtyr⁻¹, with more uncertainty over EA, due primarily to the much larger surface area involved. In general, the ice mass change uncertainties match those of the IMBIE study, as well as
- ²⁵ other recent studies (King et al., 2012; Jacob et al., 2012). The GIA uncertainty rates are inherently difficult to quantify with current modeling techniques, and is therefore one of the strengths of the data-driven approach. A more detailed discussion on the implications of these uncertainties on the results will be provided in the next sections.





5.2 Comparisons with GPS ground stations

To gain more insight into the performance of the estimated GIA rates, as well as to ensure an equal comparison with existing GIA models, the GPS rates were compared to several variants of the estimated GIA uplift rates. The first approach uses the same

- ⁵ WRMS calculation described by Eqs. (5) and (6), using the empirical rates corrected with the LPZ GIA bias described earlier, along with the estimated GPS and GIA uncertainties. Both the full 35-station set of GPS stations were used, as well as a smaller subset of 29 stations. The 29-station subset was chosen to remove the influence that stations on Graham Land (GRA) might have on the WRMS calculations, as well as two
- other stations which showed vertical rates with large differences (> 5 mm yr⁻¹), or were opposite in sign, to neighbouring GPS sites. GRA is a particularly dynamic region, and there are many factors that could impact the comparison of the GPS and derived GIA rates (Scambos et al., 2004; Thomas et al., 2011). Examples include potentially strong elastic effects on the GPS stations, the fact that ICESat is relatively data poor in this
- ¹⁵ region, and the ability of GRACE to resolve the mass change of narrow north-south oriented features. The WRMS comparisons for both sets of GPS stations are shown in Table 3, with the stations excluded in the 29-station subset designated by square symbols in Fig. 11.

The results shown in Table 3 are useful for evaluating the performance of the various individual cases computed from the combination approach, primarily because the uncertainty of the resulting GIA rates can be used in the WRMS calculation. For comparisons of the empirically derived GIA rates to those from existing GIA models, the uncertainties of these models are not always available. Therefore, the comparisons with the GIA models were handled slightly differently, with the intention of making the comparisons more equivalent. The individual assumptions and choice of Earth model parameters for each of the models is different, and again may result in a bias offset with the observed GPS rates. To account for these, a bias term was estimated and removed





between the GPS and modelled-GIA rates before the WRMS was computed. A similar

systematic bias term was also estimated for the empirical rates from the combination approach, and was removed in addition to the LPZ-bias term discussed previously. As shown in Table 4, the average systematic bias magnitude is approximately 1 mm yr^{-1} , and has an estimated uncertainty of ~ 0.3 mm yr⁻¹, demonstrating the bias to be sta-

- tistically significant. The removal of the GPS bias serves to reduce all solutions to the same frame as the GPS network, and ideally allows the WRMS values computed to reflect the spatial correlation with the station displacements and not additional systematic differences such as global reference frame differences or far-field model assumptions. In addition to the systematic bias correction, because model uncertainties are not provided for all models, only the uncertainties of the GPS stations were used in the WRMS
 - calculations. This is equivalent to setting σ_{GIA} to zero in Eq. (6).

The empirical rates were compared to the rates predicted from three recent GIA models: the ICE-5G model (Peltier, 2004)¹, the IJ05 model (Ivins and James, 2005), and the W12a model (Whitehouse et al., 2012)². The Simon et al. (2010) revision

- of the IJ05 model was used (full sea-level equation and global ocean loading) with no Antarctic continent load change since 800 yr BP. Also included in the comparisons were the results from the earlier study by Riva et al. (2009) (Riva09). As before, comparisons were made using both the full 35 and 29-station data sets. The results are listed in Table 4, and show both the original WRMS and bias-corrected WRMS values. Note
- that the WRMS values shown for Riva09, IJ05 and ICE-5G are corrections to the values shown in Thomas et al. (2011), and partially repeated in Whitehouse et al. (2012), due to an error in their WRMS calculations (the updated values do not affect the ranking of these models in these earlier works). To visually examine the differences, a selection of three empirical solutions representing the various GRACE processing variations (CSR
- ²⁵ RL04 DDK5, CSR RL05 Reg, and DMT1-b) are plotted in Fig. 11 alongside the three GIA models, with all figures representing the 35-station case after adjustment for the





¹www.psmsl.org/train_and_info/geo_signals/gia/peltier/

²www.dur.ac.uk/pippa.whitehouse/

3518

systematic bias. For reference, plots of the original unadjusted GIA models can be found in Fig. S3 of the Supplement.

5.3 Discussion

Several observations can be made when examining the results of the combinations
 and the comparisons with the GPS vertical displacements. First, the spatial pattern of
 the empirically derived rates is mostly similar across all solutions, with most of the vari ations involving differences in magnitude. For example all solutions indicate sizeable
 uplift in WA, and a slight degree of subsidence for most of the EA interior. Similar patterns of subsidence are also observed in the W12a and IJ05 models. The same can
 be said for the uplift beneath the Filchner Ronne Ice Shelf (FRIS) and Ross Ice Shelf (RIS). The magnitude of this uplift does vary depending on the solution considered (including models), but in general the geographical location of the signal is common to all cases.

In the ASE, the empirical models indicate a level of uplift not typically predicted in this area by the other models based on ice history reconstruction. There are several plausible reasons that might explain this signal. The first is that genuine GIA uplift is taking place in this area, as suggested by a recent study by Groh et al. (2012); however, the in-situ data used to validate the results of this study were derived from only two seasonal GPS campaigns, so these data have large uncertainty and need to be confirmed

- ²⁰ by additional long-term GPS measurements. The error analysis for the combination approach, shown in Fig. 10c produced a $1-\sigma$ uncertainty level of approximately 2 mm yr⁻¹ for the ASE, making the 6+ mm yr⁻¹ uplift rates shown by all of the empirical rates in Fig. 9 statistically significant (i.e., greater than the 95% confidence interval), providing additional evidence that the uplift observed is real. The earlier results obtained by
- Riva et al. (2009) do not show the same degree of uplift in the ASE (see Fig. S3), even though a similar technique was employed. The difference can be explained by the fact that the new approach presented here considers firn compaction and surface processes via the FDM and SMB estimates from RACMO2. The Riva et al. (2009)





study did not account for any surface height or density change caused by the sizeable amount of accumulation (> 10 cm yr^{-1} , see Fig. 7) that takes place in the ASE, and assigned all volume loss a density of ice. Doing so generates a lower mass loss rate for the region; however, now that these surface processes are taken into account, the altimetry-derived mass loss is greater for the ASE, resulting in a positive mass offset when compared to GRACE that is interpreted as GIA uplift in the inversion.

- Naturally, there are other plausible explanations for the observed uplift in the ASE. It is possible that the gridded ICESat height change maps may overestimate the total volume loss in the ASE, or that GRACE is underestimating the mass loss. In either
- of these cases, the unaccounted for positive mass differential would be interpreted as GIA uplift in the combination. Alternatively, the SMB estimates could be overestimating the amount of accumulation in the region, again causing the positive mass differential with what GRACE observes to be treated as GIA uplift. While no long-term GPS vertical rates are currently available in the ASE, there have been a handful of permanent
- stations recently installed which will help validate these claims³. These future GPS measurements should also help to clarify the magnitude and spatial extent of the uplift, as some of the GIA solutions predict more widespread uplift than others. In particular, the RL04-based solutions tend to produce a larger extent of GIA uplift over the ASE than the RL05-based solutions, while the RL05 solutions indicate more uplift over the FRIS.

In the Philippi/Denman (PD) sectors, the empirical GIA rates shows a stronger uplift pattern than those found in the GIA models (Figs. 9 and S3). It is not believed that the estimated uplift is the result of any unmodelled accumulation, as the ICESat and FDM results agree well in this region, and the positive mass anomaly in the area is consistently observed in the GRACE solutions (Fig. 2), in particular in the regularized

²⁵ consistently observed in the GRACE solutions (Fig. 2), in particular in the regularized solutions, which tend to have higher spatial resolution. Also, the uncertainty analysis does not suggest any unusual circumstances in the area. Unfortunately, the comparisons with the GPS rates are inconclusive, since the few stations in the area are located



CCC I

5

³www.polenet.com

on the perimeter of the region in question. As such, the presence of genuine GIA uplift in the region will require more investigation before this can be confirmed.

Looking at the WRMS values in Table 3, most solutions compare well with each other, with differences only at the $0.1-0.2 \text{ mm yr}^{-1}$ level. Again, these were computed using only the LPZ CIA bias calibration and considering the uncertainty of both the

- ⁵ using only the LPZ GIA bias calibration and considering the uncertainty of both the GIA and GPS stations. When examining Table 4, which only considered GPS station uncertainties and removed an additional systematic bias term, more variation in the results can be seen. The CSR RL04 DDK3 solution showed the lowest WRMS after the systematic bias is removed at 1.1 mm yr⁻¹, but the results of the other regularized
- solutions for both RL04 and RL05 are comparable, particularly for the 35-station set. It is interesting to note that the RL04 solutions have a larger systematic bias correction than the RL05 solutions, which is likely due to the difference in reference frames used in the GPS and RL04 GRACE data processing. In nearly all cases, the 29-station results are lower than the 35-station set. When comparing the empirical results to the model nearly all cases are used in the solutions.
- results, either with or without the systematic bias removed, the empirical rates show consistently lower values, with the IJ05 model having the closest similarity in terms of WRMS and spatial distribution of GIA uplift.

Regarding the ice mass change estimates, the values for all cases were relatively consistent. This is primarily a consequence of the fact that the surface height change

- ²⁰ information was fixed to that determined by the altimetry and FDM. In the combination, this essentially determines the variability of the firn and ice layers, forcing any variation in mass change seen by GRACE to go into the GIA estimates. The empirically derived ice mass change rate of $-100 \pm 44 \,\text{Gtyr}^{-1}$ for the AIS from this study agrees to within a single standard deviation to the recent IMBIE study (Shepherd et al., 2012)
- for a similar time frame (-57 ± 50 , October 2003–December 2008, using W12a and IJ05_R2; $-137 \pm 49 \,\text{Gt yr}^{-1}$ using ICE-5G), and shows slightly more ice mass loss than the recent results by King et al. (2012) (-68.7 ± 17.5 , using W12a), but less ice loss than estimates by Jacob et al. (2012) (-165 ± 36 , $1-\sigma$, using ICE-5G). In particular, the increased GIA predicted for the ASE in this study produces significantly more ice



loss in the WAIS (-105 ± 22) that those of the IMBIE study (-68 ± 23), but are similar to those from King et al. (2012) ($-117 \text{ Gtyr}^{-1} \pm 9.2$).

6 Conclusions

- This study revisited the approach developed by Riva et al. (2009) for estimating present-day GIA and ice mass change using a combination of satellite altimetry and gravimetry. An updated and extended ICESat surface height change map was combined with a range of different GRACE solutions, along with an advanced regional atmospheric climate model and associated firn densification model (FDM). New ICESat surface trends were computed, for the first time over Antarctica, using an overlapping
- footprint approach, complete with a custom set of campaign biases. The FDM and surface mass balance (SMB) estimates derived from RACMO2 addressed a key limitation in the earlier study, and enabled the combination approach to treat variations in surface height and density due to firn compaction and other surface processes. Another key element of the analysis was the calibration of the results to a low-precipitation zone
- in East Antarctica, which helped reduce the impact of the mm-level (unknown) biases inherent to the satellite input data sets. Lastly, knowledge of the uncertainties for the various input data sources provided the opportunity to generate realistic error assessments of both the GIA and ice mass change estimates through formal error propagation methods. The total GIA mass change estimates for the AIS ranged from 53 to
- ²⁰ 100 Gt yr⁻¹ (EA: 31 to 53 Gt yr⁻¹; WA: 19 to 48 Gt yr⁻¹), depending on the GRACE solution used, with an estimated uncertainty of \pm 40 Gt yr⁻¹ (EA: \pm 34; WA: \pm 21). Over the time frame February 2003 to October 2009, the corresponding ice mass change averaged -100 ± 44 Gt yr⁻¹ (EA: 5 ± 38 Gt yr⁻¹, WA: -105 ± 22 Gt yr⁻¹) across all solutions. The empirically derived GIA rates show some noticeable differences to other recent
- ²⁵ GIA models derived using the more traditional ice history reconstruction approach, such as in the Amundsen Sea (ASE) sector and the Philippi/Denman (PD) sectors, but also show many similarities, such as the general subsidence in East Antarctica and





uplift beneath the Ross and Filchner Ronne Ice Shelf. The empirical GIA rates generated from this approach showed good overall agreement to an independent set of GPS-derived vertical rates, although there are no long-term GPS records in some of the suspected uplift zones, such as the ASE and PD sectors, so the estimated verti-

5 cal rates in these areas cannot currently be verified. Nonetheless, the results from the combination approach demonstrate that the technique has the potential to reduce the uncertainty surrounding both Antarctic GIA and ice mass change estimates.

Supplementary material related to this article is available online at: http://www.the-cryosphere-discuss.net/7/3497/2013/tcd-7-3497-2013-supplement. pdf.

Acknowledgements. The authors would like to thank Pavel Ditmar and Hassan Hashemi Farahani for providing the DMT solutions, Himansu Save for the CSR Regularized solutions, and Jürgen Kusche for the DDK solutions. Additional thanks go to Pippa Whitehouse for the W12a GIA model, Erik Ivins for the IJ05 model, and Richard Peltier for the ICE-5G model. All figures were produced using the Generic Mapping Tools (GMT). This study was partially funded by the Division for Earth and Life Sciences (ALW) with financial aid from the New Netherlands Polar Programme (NNPP) of the Netherlands Organization for Scientific Research (NWO). MAK is a recipient of an Australian Research Council Future Fellowship (project number FT110100207).

References

10

15

- Chambers, D. P. and Bonin, J. A.: Evaluation of Release-05 GRACE time-variable gravity coef-20 ficients over the ocean, Ocean Sci., 8, 859-868, doi:10.5194/os-8-859-2012, 2012. 3504 Chen, J. L., Wilson, C. R., Blankenship, D. D., and Tapley, B. D.: Antarctic mass rates from GRACE, Geophys. Res. Lett., 33, L11502, doi:10.1029/2006GL026369, 2006. 3498 Cheng, M. and Tapley, B. D.: Variations in the Earth's oblateness during the past 28 years. J. 25
 - Geophys. Res., 109, B09402, doi:10.1029/2004JB003028, 2004. 3503



- Ewert, H., Popov, S. V., Richter, A., Schwabe, J., Scheinert, M., and Dietrich, R.: Precise analysis of ICESat altimetry data and assessment of the hydrostatic equilibrium for sub-glacial Lake Vostok, East Antarctica, Geophys. J. Int., 191, 557–568, doi:10.1111/j.1365-246X.2012.05649.x, 2012. 3508
- ⁵ Groh, A., Ewert, H., Scheinert, M., Fritsche, M., Rülke, A., Richter, A., Rosenau, R., and Dietrich, R.: An investigation of Glacial Isostatic Adjustment over the Amundsen Sea sector, West Antarctica, Global Planet. Change, 98–99, 45–53, doi:10.1016/j.gloplacha.2012.08.001, 2012. 3518

Gunter, B., Urban, T., Riva, R., Helsen, M., Harpold, R., Poole, S., Nagel, P., Schutz, B., and

- ¹⁰ Tapley, B.: A comparison of coincident GRACE and ICESat data over Antarctica, J. Geodesy, 83, 1051–1060, doi:10.1007/s00190-009-0323-4, 2009. 3499, 3507, 3508
 - Gunter, B., Riva, R., Urban, T., Harpold, R., Schutz, B., Nagel, P., and Helsen, M.: Evaluation of GRACE and ICESat mass change estimates over Antarctica, in: Gravity, Geoid and Earth Observation, edited by: Mertikas, S. P., vol. 135 of International Association of
- Geodesy Symposia, Springer, Berlin, Heidelberg, doi:10.1007/978-3-642-10634-7_75, 563-569, 2010. 3499, 3507, 3512
 - Horwath, M. and Dietrich, R.: Signal and error in mass change inferences from GRACE: the case of Antarctica, Geophys. J. Int., 177, 849–864, doi:10.1111/j.1365-246X.2009.04139.x, 2009. 3498
- Hughes, G. B. and Chraibi, M.: Calculating Ellipse Overlap Areas, available at: http://arxiv.org/ abs/1106.3787, 2011. 3506
 - Ivins, E. R. and James, T. S.: Antarctic glacial isostatic adjustment: a new assessment, Antarct. Sci., 17, 541–553, 2005. 3517
- Jacob, T., Wahr, J., Pfeffer, W., and Swenson, S.: Recent contributions of glaciers and ice caps to sea level rise, Nature, 482, 514–518, 2012. 3498, 3515, 3520
 - Kaspers, K. A., van de Wal, R. S. W., van den Broeke, M. R., Schwander, J., van Lipzig, N. P. M., and Brenninkmeijer, C. A. M.: Model calculations of the age of firn air across the Antarctic continent, Atmos. Chem. Phys., 4, 1365–1380, doi:10.5194/acp-4-1365-2004, 2004. 3510
 King, M. A., Bingham, R. J., Moore, P., Whitehouse, P. L., Bentley, M. J., and Milne, G. A.:
- Lower satellite-gravimetry estimates of Antarctic sea-level contribution, Nature, 491, 586– 589, doi:10.1038/nature11621, 2012. 3515, 3520, 3521





- Klees, R., Revtova, E. A., Gunter, B. C., Ditmar, P., Oudman, E., Winsemius, H. C., and Savenije, H. H. G.: The design of an optimal filter for monthly GRACE gravity models, Geophys. J. Int., 175, 417–432, doi:10.1111/j.1365-246X.2008.03922.x, 2008. 3504
- Kusche, J.: Approximate decorrelation and non-isotropic smoothing of time-variable GRACE-
- type gravity field models, J. Geodesy, 81, 733–749, doi:10.1007/s00190-007-0143-3, 2007. 3504
 - Lenaerts, J. T. M., van den Broeke, M. R., van de Berg, W. J., van Meijgaard, E., and Kuipers Munneke, P.: A new, high-resolution surface mass balance map of Antarctica (1979– 2010) based on regional atmospheric climate modeling, Geophys. Res. Lett., 39, L04501, doi:10.1029/2011GL050713, 2012, 3509
- doi:10.1029/2011GL050/13, 2012. 3509
 Ligtenberg, S. R. M., Helsen, M. M., and van den Broeke, M. R.: An improved semi-empirical model for the densification of Antarctic firn, The Cryosphere, 5, 809–819, doi:10.5194/tc-5-809-2011, 2011. 3509
 - Liu, X., Ditmar, P., Siemes, C., Slobbe, D., Revtova, E., Klees, R., Riva, R., and Zhao, Q.: DEOS
- ¹⁵ Mass Transport model (DMT-1) based on GRACE satellite data: methodology, validation, and application to estimating the ice mass balance of Greenland, Geophys. J. Int., 181, 769–788, 2010. 3503
 - Peltier, W.: Global glacial isostasy and the surface of the ice-age Earth: the ICE-5G (VM2) Model and GRACE, Annu. Rev. Earth Planet. Sci., 32, 111–149, 2004. 3517
- Rignot, E., Bamber, J., van den Broeke, M., Davis, C., Li, Y., van de Berg, W., and van Meijgaard, E.: Recent Antarctic ice mass loss from radar interferometry and regional climate modelling, Nat. Geosci., 1, 106–110, doi:10.1038/ngeo102, 2008. 3498, 3511, 3515
 - Riva, R. E. M., Gunter, B. C., Urban, T. J., Vermeersen, B. L. A., Lindenbergh, R. C., Helsen, M. M., Bamber, J. L., van den Broeke, M. R., and Schutz, B. E.: Glacial isostatic
- adjustment over Antarctica from combined ICESat and GRACE satellite data, Earth Planet.
 Sc. Lett., 288, 516–523, doi:10.1016/j.epsl.2009.10.013, 2009. 3500, 3501, 3502, 3507, 3517, 3518, 3521
 - Save, H., Bettadpur, S., and Tapley, B.: Reducing errors in the GRACE gravity solutions using regularization, J. Geodesy, 86, 695–711, doi:10.1007/s00190-012-0548-5, 2012. 3504
- Scambos, T. A., Bohlander, J. A., Shuman, C. A., and Skvarca, P.: Glacier acceleration and thinning after ice shelf collapse in the Larsen B embayment, Antarctica, Geophys. Res. Lett., 31, L18402, doi:10.1029/2004GL020670, 2004. 3516





- Seo, K.-W., Wilson, C. R., Chen, J., and Waliser, D. E.: GRACE's spatial aliasing error, Geophys. J. Int., 172, 41–48, doi:10.1111/j.1365-246X.2007.03611.x, 2008. 3499, 3503
- Shepherd, A., Ivins, E. R., A, G., Barletta, V. R., Bentley, M. J., Bettadpur, S., Briggs, K. H., Bromwich, D. H., Forsberg, R., Galin, N., Horwath, M., Jacobs, S., Joughin, I., King, M. A.,
- Lenaerts, J. T. M., Li, J., Ligtenberg, S. R. M., Luckman, A., Luthcke, S. B., McMillan, M., Meister, R., Milne, G., Mouginot, J., Muir, A., Nicolas, J. P., Paden, J., Payne, A. J., Pritchard, H., Rignot, E., Rott, H., Sørensen, L. S., Scambos, T. A., Scheuchl, B., Schrama, E. J. O., Smith, B., Sundal, A. V., van Angelen, J. H., van de Berg, W. J., van den Broeke, M. R., Vaughan, D. G., Velicogna, I., Wahr, J., Whitehouse, P. L., Wingham, D. J.,
- ¹⁰ Yi, D., Young, D., and Zwally, H. J.: A reconciled estimate of ice-sheet mass balance, Science, 338, 1183–1189, doi:10.1126/science.1228102, 2012. 3499, 3515, 3520
 - Siegfried, M., Hawley, R., and Burkhart, J.: High-resolution ground-based GPS measurements show intercampaign bias in ICESat elevation data near summit, Greenland, IEEE T. Geosci. Remote, 49, 3393–3400, doi:10.1109/TGRS.2011.2127483, 2011. 3508
- ¹⁵ Siemes, C., Ditmar, P., Riva, R., Slobbe, D., Liu, X., and Farahani, H.: Estimation of mass change trends in the Earth's system on the basis of GRACE satellite data, with application to Greenland, J. Geodesy, 87, 69–87, doi:10.1007/s00190-012-0580-5, 2013. 3503
 - Simon, K., James, T., and Ivins, E.: Ocean loading effects on the prediction of Antarctic glacial isostatic uplift and gravity rates, J. Geodesy, 84, 305–317, doi:10.1007/s00190-010-0368-4, 2010. 3517

20

- Slobbe, D., Lindenbergh, R., and Ditmar, P.: Estimation of volume change rates of Greenland's ice sheet from ICESat data using overlapping footprints, Remote Sens. Environ., 112, 4204–4213, doi:10.1016/j.rse.2008.07.004, 2008. 3500, 3505
- Swenson, S. and Wahr, J.: Post-processing removal of correlated errors in GRACE data, J. Geophys. Res., 33, L08402, doi:10.1029/2005GL025285, 2006. 3504
 - Swenson, S., Chambers, D., and Wahr, J.: Estimating geocenter variations from a combination of GRACE and Ocean Model output, J. Geophys. Res., B08410, doi:10.1029/2007JB005338, 2008. 3503

Thomas, I. D., King, M. A., Bentley, M. J., Whitehouse, P. L., Penna, N. T., Williams, S. D. P.,

³⁰ Riva, R. E. M., Lavallee, D. A., Clarke, P. J., King, E. C., Hindmarsh, R. C. A., and Koivula, H.: Widespread low rates of Antarctic glacial isostatic adjustment revealed by GPS observations, Geophys. Res. Lett., 38, L22302, doi:10.1029/2011GL049277, 2011. 3511, 3516, 3517







Thompson, P. F., Bettadpur, S. V., and Tapley, B. D.: Impact of short period, non-tidal, temporal mass variability on GRACE gravity estimates, Geophys. Res. Lett., 31, L23S10, doi:10.1029/2003GL019285, 2004. 3499

Urban, T. and Schutz, B. E.: ICESat sea level comparisons, Geophys. Res. Lett., 32, L06619, doi:10.1029/2005GL024306, 2005. 3508

Urban, T., Pie, N., Felikson, D., Gunter, B., Harpold, R., and Schutz, B.: Comparison of repeat track, crossover, and overlapping footprint elevation change methods from satellite altimetry: ICESat-1 elevation changes of Greenland and Antarctica, Geophys. J. Int., in preparation, 2013. 3506, 3507

5

15

25

¹⁰ Velicogna, I. and Wahr, J.: A method for separating Antarctic postglacial rebound and ice mass balance using future ICESat Geoscience Laser Altimeter System, Gravity Recovery and Climate Experiment, and GPS satellite data, J. Geophys. Res.-Sol. Ea., 107, 2263, doi:10.1029/2001JB000708, 2002. 3500

Velicogna, I. and Wahr, J.: Measurements of time-variable gravity show mass loss in Antarctica, Science, 311, 1754–1756. doi:10.1126/science.1123785, 2006, 3499

Wahr, J., Molenaar, M., and Bryan, F.: Time variability of the Earth's gravity field: hydrological and oceanic effects and their possible detection using GRACE, J. Geophys. Res., 103, 30205–30230, doi:10.1029/98JB02844, 1998. 3514

Wahr, J., Wingham, D., and Bentley, C.: A method of combining ICESat and GRACE satellite

- ²⁰ data to constrain Antarctic mass balance, J. Geophys. Res., 105, 16279–16294, 2000. 3499, 3500
 - Whitehouse, P. L., Bentley, M. J., Milne, G. A., King, M. A., and Thomas, I. D.: A new glacial isostatic adjustment model for Antarctica: calibrated and tested using observations of relative sea-level change and present-day uplift rates, Geophys. J. Int., 190, 1464–1482, doi:10.1111/j.1365-246X.2012.05557.x, 2012. 3512, 3517
 - Zwally, J. H., Giovinetto, M. B., Beckley, M. A., and Saba, J. L.: Antarctic and Greenland Drainage Systems, GSFC Cryospheric Sciences Laboratory, available at: http://icesat4. gsfc.nasa.gov/cryo_data/ant_grn_drainage_systems.php, (last access: 12 July 2013), 2012. 3507, 3514

| Campaign | Start Date | End Date | #Days | Bias (m) | σ (m) |
|----------|-------------|-------------|-------|----------|-------|
| 1a/b | 20 Feb 2003 | 29 Mar 2003 | 38 | -0.046 | 0.017 |
| 2a | 25 Sep 2003 | 19 Nov 2003 | 55 | -0.057 | 0.015 |
| 2b | 17 Feb 2004 | 21 Mar 2004 | 34 | -0.038 | 0.017 |
| 2c | 18 May 2004 | 21 Jun 2004 | 35 | -0.004 | 0.047 |
| 3a | 3 Oct 2004 | 8 Nov 2004 | 37 | -0.053 | 0.034 |
| 3b | 17 Feb 2005 | 24 Mar 2005 | 36 | -0.035 | 0.023 |
| 3c | 20 May 2005 | 23 Jun 2005 | 35 | -0.019 | 0.024 |
| 3d | 21 Oct 2005 | 24 Nov 2005 | 35 | 0.008 | 0.020 |
| 3e | 22 Feb 2006 | 28 Mar 2006 | 34 | 0.009 | 0.013 |
| 3f | 24 May 2006 | 26 Jun 2006 | 33 | -0.002 | 0.026 |
| 3g | 25 Oct 2006 | 27 Nov 2006 | 34 | 0.020 | 0.014 |
| 3h | 12 Mar 2007 | 14 Apr 2007 | 34 | 0.015 | 0.010 |
| 3i | 2 Oct 2007 | 5 Nov 2007 | 37 | 0.012 | 0.014 |
| Зј | 17 Feb 2008 | 21 Mar 2008 | 34 | 0.031 | 0.013 |
| 3k | 4 Oct 2008 | 19 Oct 2008 | 16 | 0.043 | 0.029 |
| 2d | 25 Nov 2008 | 17 Dec 2008 | 23 | 0.029 | 0.025 |
| 2e | 9 Mar 2009 | 11 Apr 2009 | 34 | 0.045 | 0.056 |
| 2f | 30 Sep 2009 | 11 Oct 2009 | 12 | 0.037 | 0.055 |

 Table 1. Estimated ICESat campaign biases and uncertainties by campaign.



| Table 2. Estimates | of the mas | s change | components | derived | from th | e data-driven | approach. |
|----------------------|--------------|----------|------------|---------|---------|---------------|-----------|
| Uncertainties are 1- | - <i>σ</i> . | | | | | | |

| | Max | L | PZ bias | Total Est. Mass Change | | | | | lc | Ice mass change, | | |
|--------------------|------------|-------------------------------|--------------------|------------------------|--------|-------------------------|------|---------|---------------------------|------------------|----------|-------------------------|
| | Sph. Harm. | GIA | GRACE | fror | n GRAC | E (Gtyr ⁻¹) | Esti | nated (| GIA (Gtyr ⁻¹) | GRA | ACE – GI | A (Gtyr ⁻¹) |
| Solution | deg × ord | $\mathrm{mm}\mathrm{yr}^{-1}$ | $\rm mmyr^{-1}EWH$ | EA | WA | AIS | EA | WA | AIS | EA | WA | AIS |
| CSR RL04 | 60 | 1.8 | 1.4 | 57 | -71 | -13 | 52 | 34 | 87 | 5 | -105 | -100 |
| CSR RL04 DDK3 | 60 | 1.7 | 1.0 | 53 | -66 | -13 | 48 | 40 | 87 | 5 | -105 | -100 |
| CSR RL05 | 60 | 1.9 | 1.7 | 42 | -77 | -35 | 37 | 28 | 65 | 5 | -105 | -100 |
| CSR RL05 DDK5 | 60 | 1.9 | 1.7 | 42 | -78 | -36 | 37 | 27 | 64 | 5 | -105 | -100 |
| CSR RL05 Reg | 120 | 1.9 | 1.7 | 42 | -78 | -36 | 36 | 27 | 63 | 5 | -105 | -100 |
| GFZ RL04 | 120 | 1.8 | 1.2 | 58 | -61 | -3 | 53 | 45 | 98 | 4 | -106 | -101 |
| GFZ RL04 DDK3 | 120 | 1.6 | 0.6 | 58 | -59 | -1 | 53 | 46 | 100 | 5 | -105 | -101 |
| GFZ RL05 | 90 | 2.2 | 2.5 | 36 | -81 | -45 | 31 | 23 | 54 | 5 | -105 | -99 |
| GFZ RL05 DDK5 | 90 | 2.1 | 2.4 | 39 | -85 | -47 | 33 | 19 | 53 | 5 | -104 | -99 |
| DMT-1b | 60 | 1.6 | 0.6 | 40 | -58 | -18 | 34 | 47 | 81 | 6 | -106 | -99 |
| IJ05 | | | | | | | 33 | 41 | 74 | | | |
| ICE-5G | | | | | | | 53 | 48 | 101 | | | |
| W12a | | | | | | | 5 | 46 | 51 | | | |
| Riva09 | | | | | | | 59 | 33 | 92 | | | |
| Est. Uncertainties | | | | 18 | 6 | 19 | 34 | 21 | 40 | 38 | 22 | 44 |





Table 3. Comparison of estimated GIA rates with GPS vertical rates, using the uncertainties for the both the GPS and GIA uplift rates in the WRMS calculations.

| | WR | IMS |
|---------------|--|--|
| Solution | 29 GPS station set mm yr ⁻¹ | 35 GPS station set mm yr ⁻¹ |
| CSR RL04 | 1.9 | 2.1 |
| CSR RL04 DDK3 | 1.9 | 2.1 |
| CSR RL05 | 1.8 | 2.1 |
| CSR RL05 DDK5 | 1.8 | 2.1 |
| CSR RL05 Reg | 1.8 | 2.1 |
| GFZ RL04 | 2.0 | 2.1 |
| GFZ RL04 DDK3 | 2.0 | 2.1 |
| GFZ RL05 | 1.8 | 2.0 |
| GFZ RL05 DDK5 | 1.8 | 2.1 |
| DMT-1b | 1.8 | 2.0 |

| Dientreeinn Dat | T(7, 3497–3 | CD 9541, 2013 | | | | | |
|-----------------|------------------------|--------------------------|--|--|--|--|--|
| | Estima | ation of | | | | | |
| _ | present-da | y Antarctic | | | | | |
| | GIA and | ice mass | | | | | |
| | cha | change | | | | | |
| | B. C. Gu | B. C. Gunter et al. | | | | | |
| U | | | | | | | |
| | Title | Title Page | | | | | |
| _ | Abstract | Introduction | | | | | |
| | Conclusions | References | | | | | |
| 00 | Tables | Figures | | | | | |
| 5 | | | | | | | |
| | I | ►I | | | | | |
| | • | | | | | | |
| _ | Back | Close | | | | | |
| 010 | Full Scre | een / Esc | | | | | |
| 2. 2 5 | Printer-frier | Printer-friendly Version | | | | | |
| 0000 | Interactive | Discussion | | | | | |
| 2 | | | | | | | |



| | 2 | 9 GPS station | set | 3 | 35 GPS station set | | | |
|---------------|------|-----------------|--------------------|------|--------------------|-------------------|--|--|
| Solution | WRMS | Systematic bias | bias-corr. WRMS | WRMS | Systematic bias | bias-corr WRMS | | |
| CSR RL04 | 1.7 | 1.3 | 1.2 | 1.9 | 1.2 | 1.5 | | |
| CSR RL04 DDK3 | 1.7 | 1.3 | 1.1 | 1.9 | 1.2 | 1.4 | | |
| CSR RL05 | 1.6 | 0.9 | 1.4 | 1.8 | 0.8 | 1.6 | | |
| CSR RL05 DDK5 | 1.6 | 0.9 | 1.3 | 1.8 | 0.8 | 1.6 | | |
| CSR RL05 Reg | 1.6 | 0.9 | 1.3 | 1.8 | 0.8 | 1.6 | | |
| GFZ RL04 | 1.9 | 1.1 | 1.5 | 1.9 | 1.1 | 1.6 | | |
| GFZ RL04 DDK3 | 1.8 | 1.4 | 1.2 | 1.9 | 1.3 | 1.4 | | |
| GFZ RL05 | 1.6 | 0.6 | 1.4 | 1.7 | 0.5 | 1.6 | | |
| GFZ RL05 DDK5 | 1.6 | 0.8 | 1.4 | 1.8 | 0.6 | 1.7 | | |
| DMT-1b | 1.6 | 0.9 | 1.4 | 1.8 | 0.8 | 1.6 | | |
| IJ05 | 3.4 | 2.5 | 2.2 | 3.4 | 2.4 | 2.4 | | |
| ICE-5G | 3.0 | 1.2 | 2.8 | 3.0 | 1.1 | 2.8 | | |
| W12a | 2.2 | 1.4 | 1.7 | 2.3 | 1.3 | 1.8 | | |
| Riva09 | 2.1 | 1.6 | 1.4 | 2.2 | 1.5 | 1.7 | | |

Table 4. Comparison of estimated GIA rates with GPS vertical rates, using only uncertainties for the GPS uplift rates in the WRMS calculations.

| TCD 7, 3497–3541, 2013 | | | | | |
|--|------------------------|--|--|--|--|
| Estimation of present-day Antarctic GIA and ice mass change | | | | | |
| | | | | | |
| Title | Page | | | | |
| Abstract | Introduction | | | | |
| Conclusions | References | | | | |
| Tables | Figures | | | | |
| I | ۶I | | | | |
| • | | | | | |
| Back | Close | | | | |
| Full Screen / Esc | | | | | |
| Printer-friendly Version | | | | | |
| Interactive | Interactive Discussion | | | | |
| | | | | | |

Discussion Paper

Discussion Paper

Discussion Paper

Discussion Paper





Fig. 2. Long-term mass change trends in units of EWH computed from the following GRACE solutions: **(a)** CSR RL04, **(b)** CSR RL04 DDK3, **(c)** CSR RL05, **(d)** CSR RL05 Regularized, **(e)** GFZ RL05 DDK5, **(f)** DMT-1b.





Fig. 3. Illustration of (a) an ICESat overlapping footprint (OFP) pair, and (b) near-neighboring shots.

| Discussion Pap | T(7, 3497–3 | CD 9541, 2013 | | | | |
|----------------|------------------------|-------------------------|--|--|--|--|
| ber | Estima | ation of | | | | |
| | present-da | y Antarctic | | | | |
| | GIA and | ice mass | | | | |
| iscu | cna | nge | | | | |
| ISSI | B. C. Gu | nter et al. | | | | |
| on F | | | | | | |
| ap | | Desit | | | | |
| Ū, | The | Page | | | | |
| — | Abstract | Introduction | | | | |
| Discu | Conclusions | References | | | | |
| ISSI | Tables | Figures | | | | |
| on F | | | | | | |
| ap | [◄ | ►I | | | | |
| er | | | | | | |
| _ | Back | Close | | | | |
| Discuse | Full Scre | Full Screen / Esc | | | | |
| sion F | Printer-frier | ndly Version | | | | |
| aper | Interactive | Interactive Discussion | | | | |





Fig. 4. (a) ICESat dh/dt estimates from the OFP approach and, (b) corresponding uncertainties.





Fig. 5. Outline of the low-precipitation zone (LPZ) overlaid onto estimates of average yearly solid precipitative flux in units mm EWH yr⁻¹, together with the following location indicators: Amundsen Sea (ASE), Graham Land (GRA), Filchner Ronne Ice Shelf (FRIS), Enderby Land (END), Philippi/Denman (PD), Wilkes/Adelie Land (WA), Ross Ice Shelf (RIS).



Discussion Paper

Discussion Paper





Fig. 6. Illustration of the ICESat campaign biases determined over the LPZ for each individual campaign (grey), the mean value (blue), and the mean minus the surface deformation (cyan) predicted from the FDM (magenta).





Fig. 7. (a) FDM surface height velocities and (b) corresponding uncertainties.





Fig. 8. (a) $(GRACE - SMB) > 20 \text{ kgm}^{-2} \text{ yr}^{-1}$, (b) $(ICESat - FDM) > 6 \text{ cm} \text{ yr}^{-1}$, (c) Derived density $(mean 396 \text{ kg m}^{-3}).$





Discussion Paper















Fig. 11. Comparison of the 35-station WRMS, after an additional bias is removed with respect to the GPS rates, for (a) CSR RL04 DDK3, (b) CSR RL05 Reg, (c) DMT-1b, (d) ICE-5G, (e) IJ05, and (f) W12a.

