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Glacial Isostatic Adjustment over Antarctica from combined ICESat and GRACE satellite data

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ABSTRACT

The glacial history of Antarctica during the most recent Milankovitch cycles is poorly constrained relative to the Northern Hemisphere. As a consequence, the contribution of mass changes in the Antarctic ice sheet to global sea-level change and the prediction of its future evolution remain uncertain. The process of Glacial Isostatic Adjustment (GIA) represents the ongoing response of the solid Earth to the Late-Pleistocene deglaciation and, therefore, provides information about Antarctic glacial history. Moreover, insufficient knowledge of GIA hampers the determination of present-day changes in the Antarctic mass balance through satellite gravity measurements. Previous studies have laid the theoretical foundation for distinguishing between signals of ongoing GIA and contemporary ice mass change through the combination of satellite gravimetry and satellite altimetry. This distinction is made possible by the fact the GIA-induced changes (involving relatively dense rock) will produce a different combination of topography and gravity change than those produced by variations in ice or firn thickness (due to the lower density of these materials); however, no conclusive results have been produced to date. Here we show that, by combining laser altimetry and gravity data from the ICESat and GRACE satellite missions over the period March 2003–March 2008, the GIA contribution can indeed be isolated. The inferred GIA signal over the Antarctic continent, which represents the first result derived from direct observations by satellite techniques, strongly supports Late-Pleistocene ice models derived from glacio-geologic studies. The GIA impact on GRACE-derived estimates of mass balance is found to be 100 ± 67 Gt/yr.

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1. Introduction

Estimates of present-day GIA are crucial for determining current ice mass balance estimates over Antarctica, especially when using data from the Gravity Recovery and Climate Experiment (GRACE) (Tapley et al., 2004). This is due to the fact that, while the GRACE mission is able to accurately detect large scale changes in mass over the Antarctic continent, the technique itself is not able to distinguish between mass change due to GIA and that due to ongoing ice loss or gain. Traditionally, separating current ice mass change from GIA relies upon a modelled estimate of GIA (Velicogna and Wahr, 2006). Those estimates are particularly important over the West Antarctic Ice Sheet (WAIS), where the total GIA signal is about 2–3 times the total mass change measured by GRACE. Current GIA models rely on a reconstruction

of the ice load since the Last Glacial Maximum (LGM), which is poorly constrained and remains uncertain (Bentley, 1999; Denton and Hughes, 2002). The reconstructions are usually obtained from glacio-hydro-isostatic models (Lambeck et al., 2002; Peltier, 2004), glacial geology (Ivins and James, 2005), or glaciological models (Le Meur and Huybrechts, 1996; Huybrechts, 2002). Estimates of solid Earth deformation induced by variations in the surface ice load also require an additional parameterization of Earth's interior (elastic properties, density and viscosity). Therefore, the process of modelling present-day GIA involves a number of assumptions that, combined with the sparse availability of independent geodetic data over Antarctica, lead to large uncertainties in the final result of mass change.

The possibility to measure GIA from satellite observations originates from the combination of GRACE data, which detects total mass changes, with surface elevation changes derived from the Ice Cloud and land Elevation Satellite (ICESat) laser altimetry mission, which has been measuring Antarctic surface elevations since early 2003 (Zwally et al., 2002). The principle of separating GIA from current ice

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mass change through the combination of GRACE and ICESat measurements relies on the fact that rock and ice have very different densities. This means that variations in bedrock elevation or ice thickness have a different impact on topography than on the Earth's gravity field. In addition, both elevation and mass change signals are affected by changes in firn (partially compacted snow) thickness, driven by time-varying snow accumulation and compaction. By making use of the principle of mass conservation, it is possible to combine the effect of temporal changes in bedrock topography, ice thickness and firn thickness in order to solve for a single unknown, i.e., GIA.

In practice, we have to face the problem that we need to combine three different signals (GIA, ice and firn), but we only have two satellite datasets (GRACE mass and ICESat elevation) at our disposal. One possible solution to this involves constraining one of the three signals by means of an additional dataset. Previous studies have offered suggestions on how this might be done. [Velicogna and Wahr \(2002\)](#) investigated the use of GPS measurements to constrain GIA, [Rignot et al. \(2008\)](#) derived estimates of ice mass changes using a mass budget approach (the difference between outgoing and incoming fluxes), and [Helsen et al. \(2008\)](#) constrained firn variations. Unfortunately, each of these approaches has severe limitations: 1) the few reliable GPS time-series currently available are mostly localized in coastal areas, and therefore cannot constrain features in the interior; 2) the mass budget approach is hampered by large uncertainties in, particularly, the input (accumulation) over the WAIS; and 3) there is currently no firn variation model available for our observational period.

In this study, we show how a 5-year-long observation window and the use of a hybrid ice-firn surface density model to derive mass changes from altimetry measurements are enough to separate GIA from surface processes. Our results represent the first measurement of present-day GIA over the whole of Antarctica and, being derived from satellite observations, provide new constraints on GIA that are completely independent from any previous reconstruction of the Antarctic glacial history.

2. Datasets

We make use of 5 years of observations (March 2003–March 2008) to determine a linear trend of surface elevation change from ICESat and of surface mass change from GRACE.

The ICESat elevation changes, shown in [Fig. 1a](#), are computed from more than 5 million cross-over measurements using thirteen fully calibrated (RL428) campaign data sets. A campaign bias of 2.6 cm/yr was removed based on the value obtained by minimizing the elevation changes measured in the most arid region of East Antarctica (i.e., areas with less than 2 cm water equivalent height per year of average solid precipitation between 2003 and 2006, according to data from the European Centre for Medium-Range Weather Forecasts, ECMWF). This bias value is consistent with the calibration trend (2.0 cm/yr) that is obtained over the oceans from a comparison to the GSFC00 ([Wang, 2001](#)) mean sea-surface model (reference model) ([Urban and Schutz, 2005](#)). The resulting dataset was first spatially homogenized over a 20×20 km grid and then linearly interpolated in order to fill data gaps. Spatial averaging was based on a L1-norm (i.e., median values), instead of the more common L2-norm (i.e., mean values), because the former is known to be more robust with respect to the presence of outliers ([Claerbout and Muir, 1973](#)). Data-points outside the ice grounding line ([Vaughan et al., 1999](#)) have been removed. Due to its orbit inclination (94°) and pointing angle (nearly nadir), ICESat data do not cover latitudes higher than about 86° . We have chosen to deal with the resulting data gap by assuming elevation changes to be null inside it: this approximation is justified by the negligible (about 1 Gt/yr) mass change occurring within the polar cap, both as observed by GRACE and predicted by forward GIA models.

The GRACE mass trend, shown in [Fig. 1b](#), has been obtained from the publicly available RL04 global monthly gravity solutions produced by the Center for Space Research (CSR, Austin) ([Bettadpur, 2007](#)). The degree 2 spherical harmonic coefficient (describing variations in the Earth's oblateness) for these solutions was replaced with that obtained

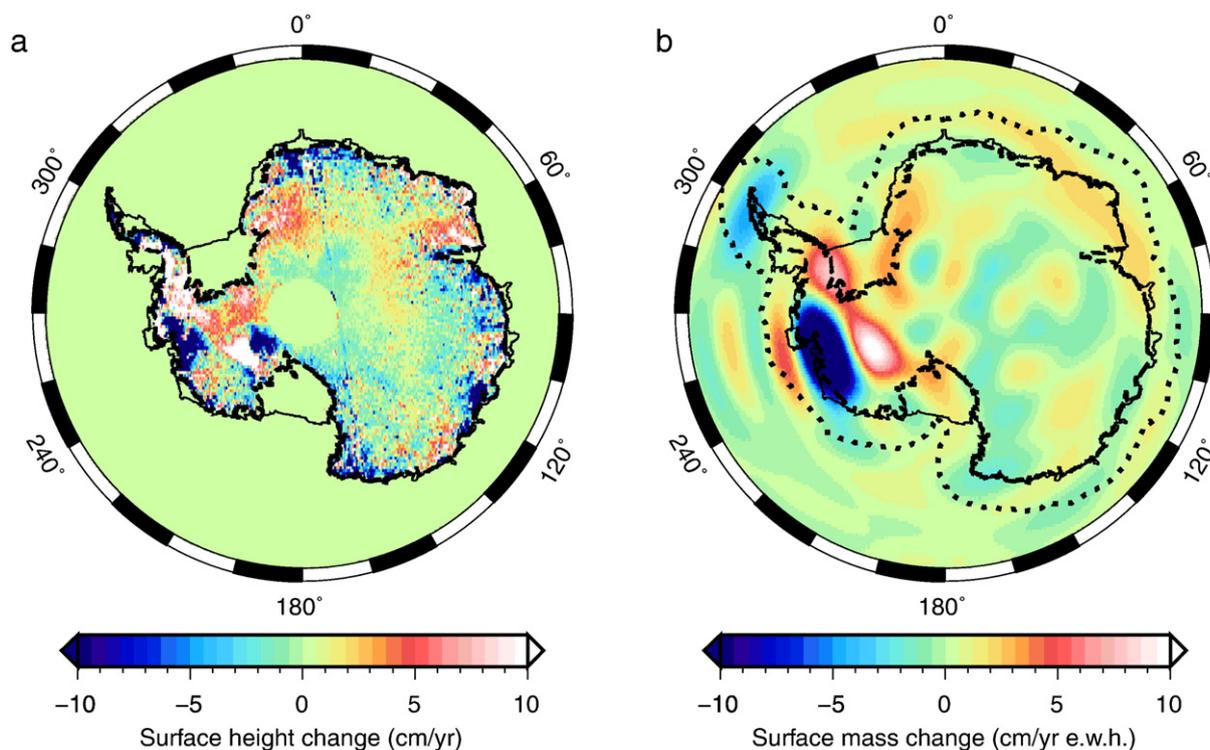


Fig. 1. Map views of the satellite data used in this study, representing trends of (a) elevation changes from ICESat, and (b) mass changes from GRACE. In panel b, a dotted line over the oceans indicates the 400-km boundary from the ice grounding line.

from Satellite Laser Ranging (SLR) (Cheng and Tapley, 2004). Degree 1 coefficients were included using values produced by (Swenson et al., 2008), and the secular rates of certain low degree harmonics (C21, S21, C30, etc.) that get removed in the standard RL04 processing were restored. For each month, the north-south oriented noise artefacts present in the (unregularized) spherical harmonic solutions have been removed through the application of a “destriping” filter similar to that implemented by Swenson and Wahr (2006). The solutions are then converted into equivalent water height on a regular grid ($0.2 \times 0.2^\circ$), with a linear trend evaluated at each grid point, where we have accounted for the effect of periodic signals due to annual variations and to the S2- and K2-tide (with cycles of 161 and 1362.7 days respectively) (Ray and Luthcke, 2006). More details on the satellite datasets and on the (post-)processing strategies can be found in Gunter et al. (2009).

3. Method

3.1. Combination strategy

The principle of mass conservation, assuming a bottom (rock) and a top (ice/firn) layer with different thickness change rates and different densities, requires the following equation to be satisfied:

$$\dot{h}_{\text{GIA}} = \frac{\dot{m}_{\text{GRACE}} - \rho_{\text{surf}} \cdot \dot{h}_{\text{ICESat}}}{\rho_{\text{rock}} - \rho_{\text{surf}}} \quad (1)$$

where superscript dots indicate time derivatives (i.e., rates), \dot{h}_{ICESat} represents elevation change rates as observed by ICESat, \dot{m}_{GRACE} mass variation rates as observed by GRACE (in terms of equivalent water height), \dot{h}_{GIA} variation rates in bedrock topography, ρ_{rock} and ρ_{surf} the average density per unit area of the rock and surface layers respectively (assumed to be constant in time). Eq. (1) can be easily derived from combining two equations for the conservation of mass ($\dot{m}_{\text{GRACE}} = \dot{m}_{\text{GIA}} + \dot{m}_{\text{surf}}$) and of volume ($\dot{h}_{\text{ICESat}} = \dot{h}_{\text{GIA}} + \dot{h}_{\text{surf}}$), and taking into account the relations between mass changes per unit area and elevation changes ($\dot{m}_{\text{GIA}} = \rho_{\text{rock}} \cdot \dot{h}_{\text{GIA}}$, and $\dot{m}_{\text{surf}} = \rho_{\text{surf}} \cdot \dot{h}_{\text{surf}}$). The combination, therefore, is based on four datasets: two trends from satellite measurements and two density maps.

In order to homogenize the spatial resolution of all datasets, we apply a Gaussian smoothing filter with a half-width of 400 km: this operation is necessary because of the considerably higher resolution of the ICESat measurements (about 30–50 km over Antarctic coastal areas) with respect to the limited GRACE resolution (where 400 km is necessary to average most of the spatially correlated noise not eliminated by the destriping filter). Note that the operation of smoothing redistributes spectral power over lower frequencies with the result of spreading the high-frequency signal over a wider area, but it does not cause a significant net signal loss. Smoothing also contributes to reduce the impact of the polar gap in the ICESat dataset and of the destriping filter on the GRACE signal amplitudes. Before smoothing the GRACE dataset, we also mask regions further than 400 km from the ice grounding line (indicated by a dotted line in Fig. 1b), because we consider those regions as being dominated by noise and not related to processes occurring over continental areas. Our methodology is considerably different from that discussed by (Wahr et al., 2000), who proposed to combine the two datasets in the spectral domain, therefore making use of the different wavelength dependence of the signal between GRACE and ICESat. We believe that a combination of smoothed datasets in the spatial domain is closer to the way GRACE measures mass changes, since we expect GRACE to capture the totality of the signal, though with a limited spatial resolution (i.e., GRACE itself acts as a smoothing device).

A crucial step to use Eq. (1) is represented by the choice of an appropriate density map to account for changes in the ice sheet (ρ_{surf}). In the case of Antarctica, where accumulation is represented by snow

and surface melt can be neglected over the grounded ice sheet, a perfectly balanced ice sheet (i.e., with neither mass nor thickness changes) would imply that precipitation, firn compaction and ice flow take place at a constant rate. Therefore, the presence of an elevation or mass change signal means that any of the three processes is deviating from the secular rate. Since, in Eq.(1), we locally make use of a single value of ρ_{surf} , we have to make an *a priori* choice of the dominating process. In order to reduce the complexity of the problem, we assume that the secular accumulation rate is in balance with the vertical ice velocity and neglect the influence of firn compaction on the elevation change signal. Furthermore, due to the limited time span of our satellite measurements, we assume that the observed elevation changes are directly related to the yearly accumulation variability (Helsen et al., 2008). Consequently, we make use of surface snow density (ranging 320–450 kg/m³) to represent ρ_{surf} , which becomes untenable over a longer time-span (exactly how many years depends on the local annual accumulation rate). The only exceptions to this are areas where rapid changes in ice velocity have been documented, in which case we consider ice dynamics to be the dominating process, and the density of pure ice to be the appropriate choice for ρ_{surf} . Therefore, our density model for the top layer (Fig. 2a) combines results from a surface density model (Kaspers et al., 2004) with areas of pure ice for those ice streams with balance velocities larger than 25 m/yr (data from Bamber et al., 2000) and discharging into the Amundsen Sea Embayment (ASE), where large mass imbalances have been observed (Rignot et al., 2008), and Graham Land (tip of the Peninsula, up to a latitude of -70°). A separate case is represented by the now stagnant Kamb Ice Stream (Joughin and Tulaczyk, 2002), where neither the density of surface firn nor of pure ice seem to reconcile ICESat and GRACE measurements over the area. For this region, we have chosen an intermediate density of 600 kg/m³.

The use of the Kaspers et al., (2004) distribution of surface snow density, which is based on a set of density observations from snowpits and firn cores that mostly represent the upper metre, is justified by the fact that the average yearly accumulation rate over the entire grounded Antarctic ice sheet is equal to a snow layer of about 0.5 m with an average density of 350 kg/m³ (van de Berg et al., 2005). As far as the regions dominated by ice changes are concerned, we have chosen a reference balance velocity of 25 m/yr from the analysis of ICESat data over Thwaites Glacier: those regions are generally characterised by elevation change rates larger than 10 cm/yr, therefore supporting our hypothesis that the dominating process is of ice dynamical origin.

We derive the density of the rock layer from the ratio between mass changes and topography changes induced by GIA. For a visco-elastic and self-gravitating Earth, this ratio has been found to be equal to two thirds of the average earth density (i.e., about 3700 kg/m³) (Wahr et al., 2000), as it can be derived from studies on the gravitational signature of GIA (Wahr et al., 1995; James and Ivins, 1998; Fang and Hager, 2001). However, when we take into account the gravitational coupling between ocean mass (re)distribution and solid earth deformation by solving the full sea-level equation (Farrell and Clark, 1976), we find the same ratio to be generally higher over the continents and lower under the oceans. Consequently, we have refined the effective rock density model (Fig. 2b) by allowing a smooth transition from 4000 kg/m³ for land to 3400 kg/m³ under the ice-shelves. Those values have been chosen after comparing forward model results obtained from different combinations of parameters, and are meant to approximate the areas with the largest GIA signal (i.e., the WAIS and coastal areas in East Antarctica). The effect of this refinement is an increase in the amplitudes of the GIA solution by about 2.5%.

Last, in order to account for the elastic response of the solid Earth to current changes in surface load, we have applied a scaling factor of 1.015 (i.e., an increase of 1.5%) to the ICESat trend. We have empirically determined this value by modelling the elastic deformation of a

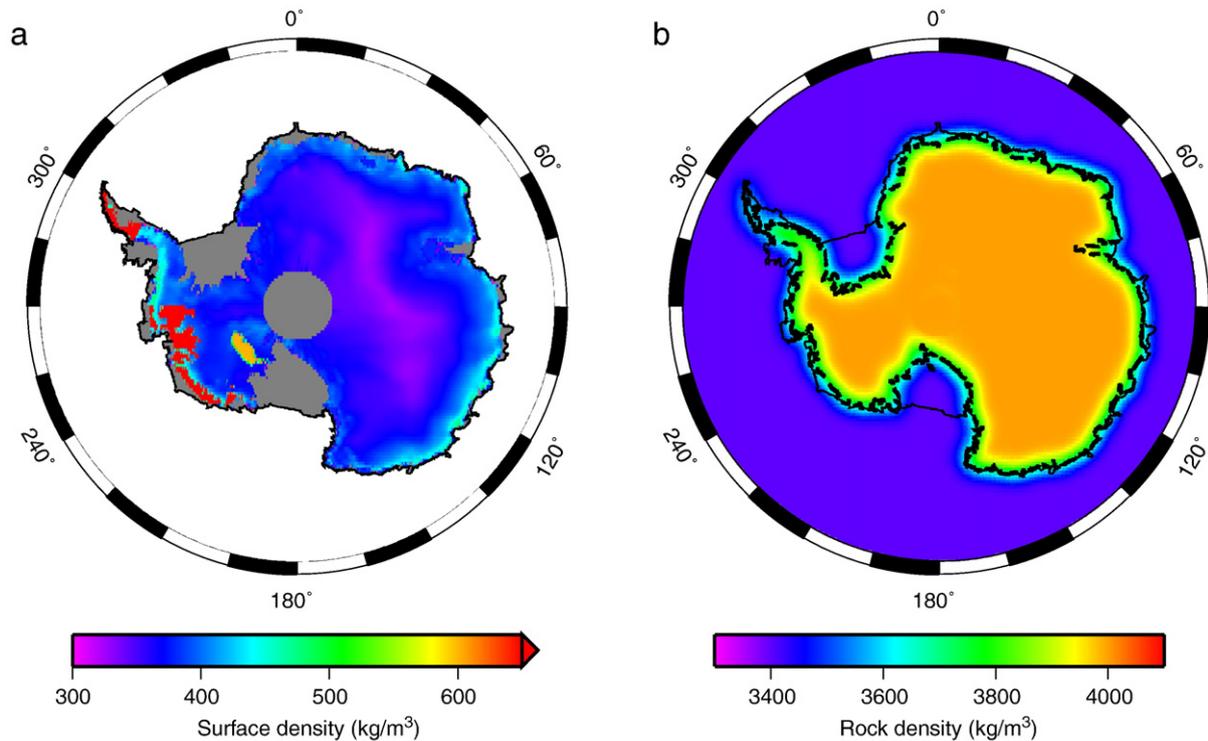


Fig. 2. Density models used in this study, representing (a) surface firn density, and (b) effective rock density. In panel (a), saturated values represent pure ice (917 kg/m^3), and data do not extend further than 86°S , to be consistent with the spatial coverage of ICESat. In panel (b), the effective rock density represents a refinement with respect to the average value of 3700 kg/m^3 , and it is meant to account for the effect of gravitational coupling between the solid earth and the oceans.

compressible, self-gravitating Earth to a load concentrated over the fast-flowing ice streams. Note that the elastic effect in the GRACE trend is implicitly taken into account by the conversion from geoid elevation changes to equivalent water height changes via the loading Love numbers (Wahr et al., 1998).

3.2. Error assessment

The GIA solution presented in this study is derived from four independent datasets: the two trends obtained from satellite data (GRACE mass and ICESat elevation) and the two density models (rock and surface layer). The representative errors for each dataset have been defined as follows:

- i- for the GRACE data, we have used the calibrated errors for the spherical harmonic coefficients provided with the CSR RL04 solution for the month of August 2006. The month of August 2006 was chosen randomly, and compared to the total spectrum of errors (i.e., cumulative error at degree 60) for all GRACE monthly solutions, ranks within one standard deviation from the mean;
- ii- for the ICESat data, we have used the standard deviation of the mean trend at each $20 \times 20 \text{ km}$ cell to model the measurement noise. In addition, we have introduced an uncertainty of 0.3 cm/yr on the correction for campaign biases (equal to half of the difference between the bias determined from the comparison to the GSFC00 mean sea surface model and the bias obtained by minimizing the elevation changes in the internal region of East Antarctica);
- iii- for the surface firn density, we have allowed a spatially-varying uncertainty equal to one third of the difference between the adopted density and the model lower boundary of 320 kg/m^3 , while we do not allow any uncertainty for those regions with a density of pure ice;
- iv- for the effective rock density, we have allowed a spatially-varying uncertainty equal to one third of the difference between the

adopted density and the model average value of 3700 kg/m^3 ; we have verified that this uncertainty is larger than the effect of lithospheric thickness variations.

Subsequently, by generating a set of normally distributed random values, we have simulated data errors following two strategies:

- a- for the satellite measurements, we have produced a white noise error realization from the given error estimates;
- b- for the density models and for the ICESat campaign bias, we have generated the error in the form of a bias, meaning that the same uncertainty was assumed to characterise the whole model.

Finally, after producing 1000 different realizations of the (normally distributed) data errors, we have computed the standard deviation of the mean.

In the case of the satellite measurement errors, we have limited our study to the effect of white noise. In reality, the GRACE and ICESat trends will also be influenced by systematic errors (e.g., due to the specific processing strategies and instrument calibrations) and coloured noise (e.g., due to the correlation between measurements taken along the same track and to the presence of 'stripes' in the GRACE solutions), which we have not attempted to assess. However, for the ICESat case, we are being rather conservative, since we use the signal r.m.s. as our measurement error. Our GRACE error, on the other hand, should be considered a lower bound, because we do not assess the impact of coloured noise (Horwath and Dietrich, 2009). The reason for representing the density errors as biases, instead of noise, comes from the fact that they are not derived from measurement errors, rather from specific assumptions. In the case of the effective rock density error, we consider the allowed variability as a conservative estimate, because it allows the density to vary up to 600 kg/m^3 from the value obtained by accounting only for solid Earth deformation (3700 kg/m^3). The proposed variability for surface firn density means that values in areas characterized by large precipitation can be as high as 600 kg/m^3 , therefore accounting for most of the

uncertainty in the choice of the representative thickness of the top layer. Over Kamb Ice Stream, firn density can vary between the density of fresh snow and that of pure ice.

4. Results

By using Eq. (1) and the datasets discussed above, we obtain the map of GIA elevation changes shown in Fig. 3. Note that, since a 400-km Gaussian smoothing has been applied to all datasets, the resulting signal is characterized by longer wavelengths and smaller amplitudes than the original signal.

In this first measurement of present-day GIA over the whole of Antarctica, we find the signal to be concentrated between the Weddell Embayment and the Ross Ice Shelf, with the largest magnitudes over the Weddel Embayment. This pattern is in very good agreement with model results obtained with the ice history IJ05 (Ivins and James, 2005) and glaciological models (Le Meur and Huybrechts, 1996; Huybrechts, 2002). In addition, we obtain a larger signal over the Antarctic Peninsula, which could reflect the impact of lateral heterogeneities in the Earth structure (Kaufmann et al., 2005; Wang and Wu, 2006), or past ice sheet change not fully constrained by current glacial–geological data (Bentley et al., 2006). Our GIA signal is significantly different from model results obtained with ice model ICE-5G (VM2) (Peltier, 2004), which produces the largest uplift next to the Ross Ice Shelf. In most of East Antarctica we obtain no relevant GIA signal over land, meaning that the observed mass and elevation changes can be entirely explained by variations in the surface firn thickness (i.e. related to variable accumulation). Two exceptions are represented by the Philippi and Totten glaciers, where a positive GIA signal is compatible with glaciological model results (Huybrechts, 2002).

Our estimate for the GIA impact on GRACE-derived estimates of mass balance amounts to 100 ± 67 Gt/yr, which agrees within one standard deviation with both our IJ05 model results (110 ± 30 Gt/yr,

depending on the viscosity profile) and our ICE-5G model results (151 Gt/yr with a simplified version of viscosity model VM2). Note that our results for IJ05 include a correction of 30 Gt/yr due to the effect of ice changes outside the Antarctic continent, not accounted for in the IJ05 ice history, that we have computed from ICE-5G (VM2). Our estimate is considerably smaller than the value of 176 ± 72 Gt/yr obtained from model results by Velicogna and Wahr (2006), who considered both IJ05 and ICE-5G ice histories in combination with a wide range of earth parameters.

Having obtained a solution for GIA, it is possible to subtract it from the GRACE and ICESat trends, and repeat the combination to solve for changes in ice and firn thickness. In Fig. 4 we show results of firn depth changes, which are concentrated in the areas where the largest ICESat signal is observed. The magnitudes are limited to a few centimetres per year, and are therefore compatible with results from climate modelling (Helsen et al., 2008). In spite of the relatively conservative ICESat error, which dominates Fig. 4b, the largest signals are statistically significant. In Fig. 5 we show results of ice thickness changes, which in the ASE are as large as -8 cm/yr and highly significant. The dynamic thickening of Kamb Ice Stream is barely visible (max 0.8 cm/yr), but statistically significant, while ice thinning over the Peninsula is poorly constrained, due to both the large uncertainty caused by low density of the ICESat measurements and the low resolution of GRACE. Note that, since our surface density model separates a priori areas dominated by variable accumulation from those dominated by ice dynamics, this distinction is maintained in the solutions presented in Figs. 4–5. However, due to the effect of smoothing, results from neighbouring regions are partially overlapping.

5. Discussion and conclusions

We can exploit the fact that our GIA results have been derived from the combination of satellite observations in the spatial domain to compare the obtained pattern with forward model results based on

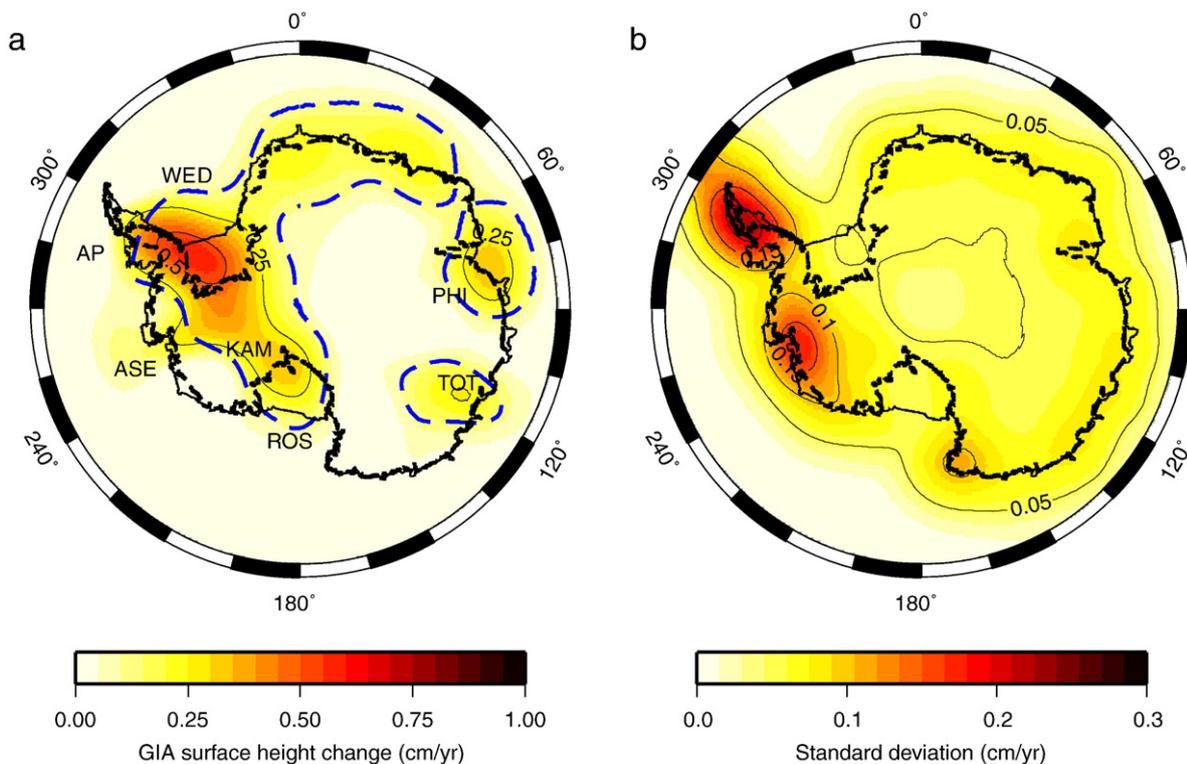


Fig. 3. GIA surface elevation changes from the ICESat-GRACE combination. A blue dashed contour bounds the areas where the GIA results are statistically significant above 2σ (95% confidence level). ROS, Ross Ice Shelf; KAM, Kamb Ice Stream; ASE, Amundsen Sea Embayment; AP, Antarctic Peninsula; WED, Weddell Embayment; PHI, TOT, Philippi and Totten glaciers.

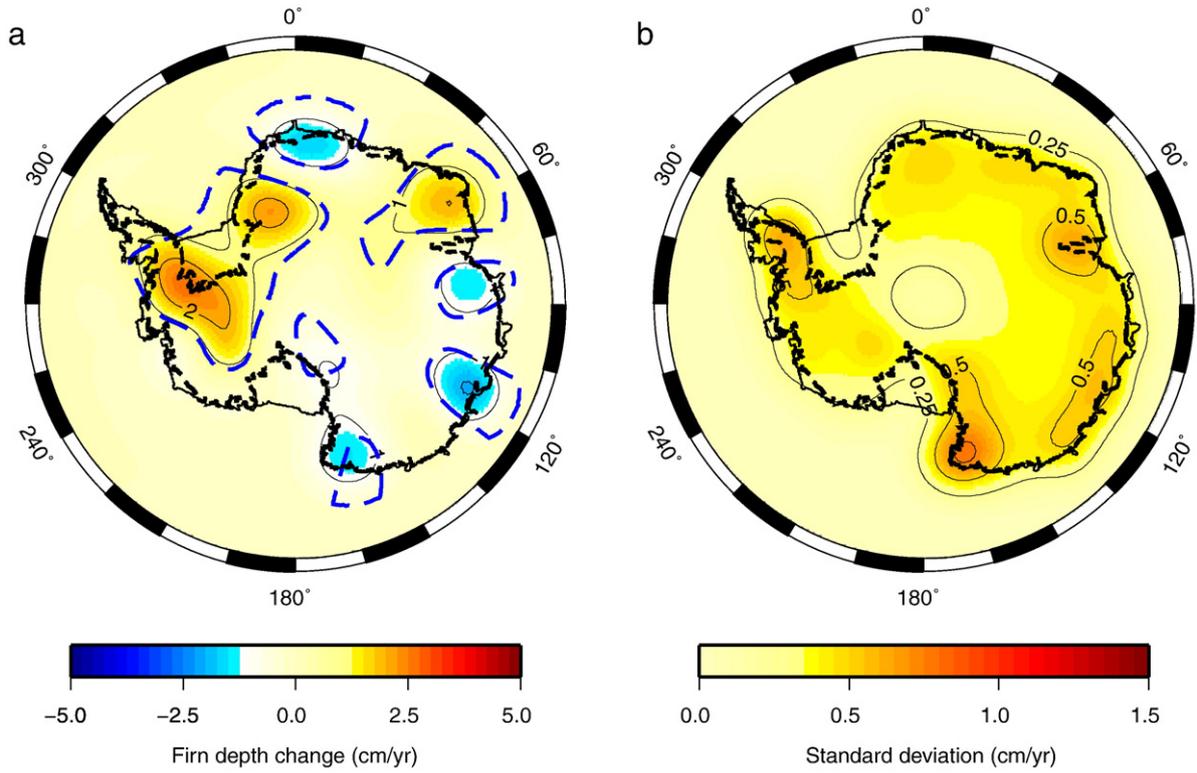


Fig. 4. Firn depth changes from the ICESat-GRACE combination. Blue dashed contour as in Fig. 3.

existing ice history reconstructions. In Fig. 6, we show our numerical results obtained for a PREM-stratified (Dziewonski & Anderson, 1981), incompressible Earth with Maxwell rheology, and by using

ice histories IJ05 and ICE-5G. As far as the viscosity profile is concerned, for ICE-5G we have adopted the prescribed VM2 model (Peltier, 2004), while for IJ05 we have chosen a thinner (65 km)

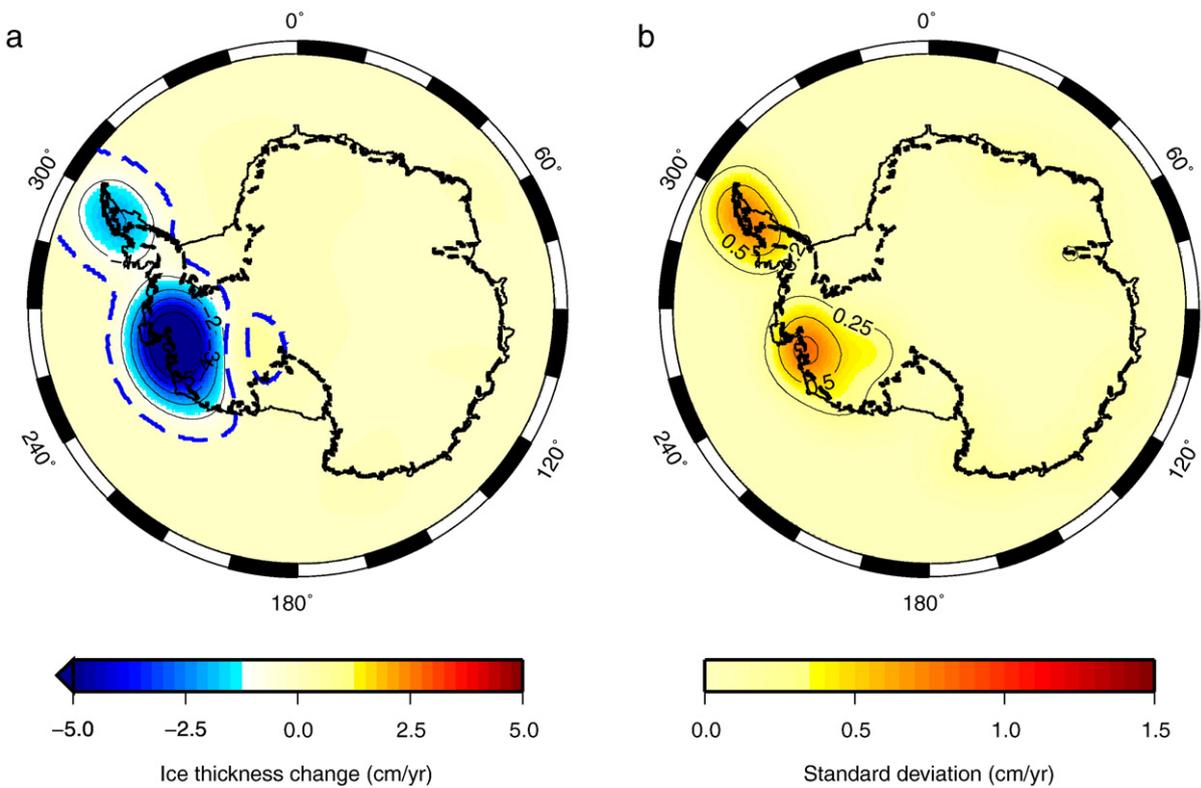


Fig. 5. Ice thickness changes from the ICESat-GRACE combination. Blue dashed contour as in Fig. 3.

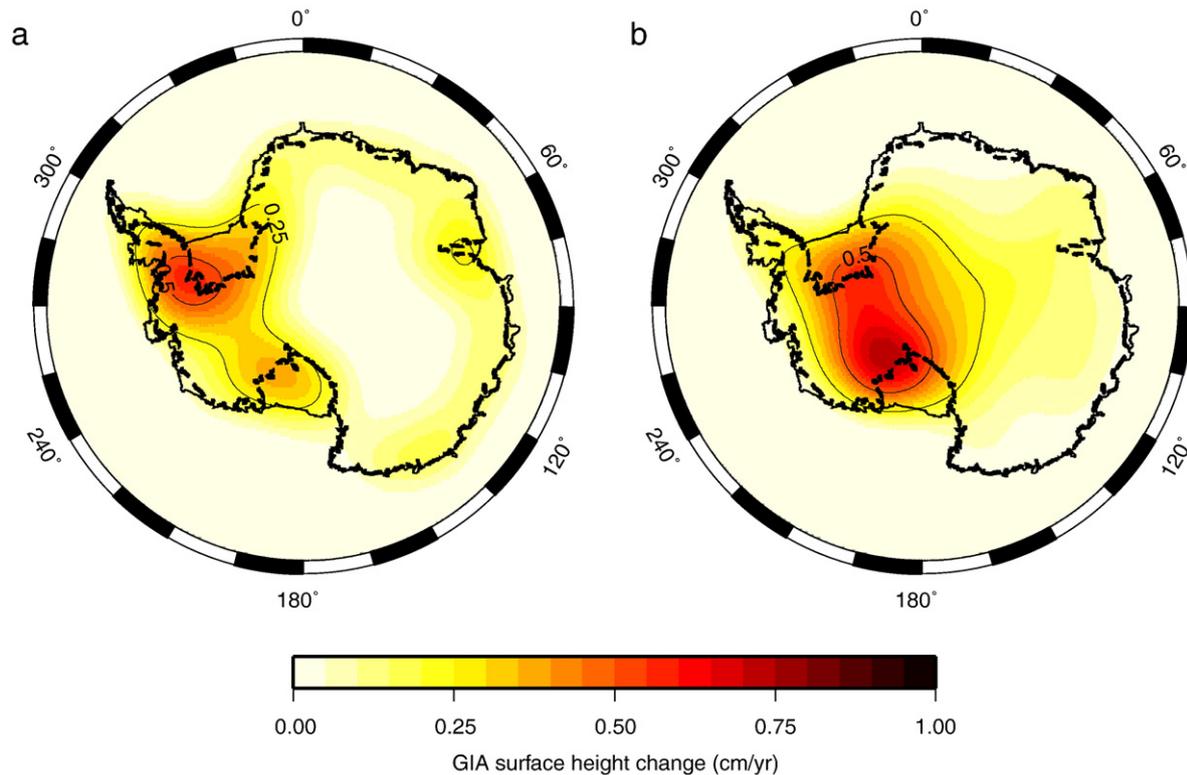


Fig. 6. GIA surface elevation changes from forward model results, based on ice histories (a) IJ05 (Ivins & James, 2005), and (b) ICE-5G (VM2) (Peltier, 2004). Both maps have been smoothed by means of a Gaussian filter with a half-width of 400 km.

elastic lithosphere, a viscosity of $5 \cdot 10^{20}$ Pas in the upper mantle and a viscosity of 10^{22} Pas in the lower mantle (Ivins and James, 2005). The comparison with Fig. 3a shows a very good agreement between our GIA results and what we obtain from IJ05 ice history, both with respect to the spatial distribution of the signal and its magnitude. The level of agreement is remarkable, if we consider that our results have been obtained from direct satellite observations, and therefore represent current dynamics, while IJ05 ice history is based on the glacio-geological record of past changes. In spite of this agreement, our mass estimate for the whole continent (100 ± 67 Gt/yr) shows a large error, which is mainly originating from noise in the satellite measurements: over the WAIS, because of uncertainties in the ICESat measurements over the areas where the largest discharge is currently taking place, and over the East Antarctic Ice Shelf (EAIS), because of the cumulative effect of the GRACE noise over its large surface. However, most of the noise is concentrated over areas where the GIA signal is almost null: when we limit the mass change estimate to regions where our GIA results are statistically significant above the 95% confidence level (i.e., bounded by the blue dashed contours in Fig. 3a), we obtain a mass change of 80 ± 24 Gt/yr.

Apart from the impact of data noise and uncertainties in the density models, a variety of error sources can affect our final results. The main issues are: GRACE and ICESat (post-) processing strategies, the limited temporal resolution and the spatial interpolation of ICESat data, and the correction for variable snow accumulation. As far as the choice of a specific set of GRACE products is concerned, we have verified that the solution discussed here is compatible to within one standard deviation of results obtained by using the GRACE monthly fields provided by the GeoForschungsZentrum Potsdam (GFZ). A further comparison between the various approaches used to process GRACE measurements is beyond the scope of this paper and is the object of current and future studies. The spatial interpolation of ICESat data remains a challenging issue, since most Antarctic mass loss occurs as a result of the discharge of fast and narrow glaciers, which

may not be sufficiently sampled by the relatively sparse ICESat measurements in the coastal areas; however, ICESat does detect large and well-defined basin-scale changes surrounding such outlets, thereby detecting the major extent of ice loss over the five years considered here. Another issue regarding the elevation trends obtained from ICESat is that measurement campaigns, each lasting about one month, only occur 2–3 times per year, which limits the possibility to separate seasonal signals from the secular trend, as discussed by Gunter et al. (2009); however, a large part of our GIA signal is located over the two main ice shelves and is therefore only marginally influenced by firn depth variations over grounded ice.

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