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# The effect of signal leakage and glacial isostatic rebound on GRACE-derived ice mass changes in Iceland

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## SUMMARY

Monthly gravity field models from the GRACE satellite mission are widely used to determine ice mass changes of large ice sheets as well as smaller glaciers and ice caps. Here, we investigate in detail the ice mass changes of the Icelandic ice caps as derived from GRACE data. The small size of the Icelandic ice caps, their location close to other rapidly changing ice covered areas and the low viscosity of the mantle below Iceland make this especially challenging. The mass balance of the ice caps is well constrained by field mass balance measurements, making this area ideal for such investigations. We find that the ice mass changes of the Icelandic ice caps derived from GRACE gravity field models are influenced by both the large gravity change signal resulting from ice mass loss in southeast Greenland and the mass redistribution within the Earth mantle due to glacial isostatic adjustment since the Little Ice Age (~1890 AD). To minimize the signal that leaks towards Iceland from Greenland, we employ an independent mass change estimate of the Greenland Ice Sheet derived from satellite laser altimetry. We also estimate the effect of post Little Ice Age glacial isostatic adjustment, from knowledge of the ice history and GPS network constrained crustal deformation data. We find that both the leakage from Greenland and the post Little Ice Age glacial isostatic adjustment are important to take into account, in order to correctly determine Iceland ice mass changes from GRACE, and when applying these an average mass balance of the Icelandic ice caps of  $-11.4 \pm 2.2$  Gt yr<sup>-1</sup> for the period 2003–2010 is found. This number corresponds well with available mass balance measurements.

**Key words:** Inverse theory; Time variable gravity; Global change from geodesy; Glaciology; Arctic region.

## 1 INTRODUCTION

The response of the cryosphere to the changing climate is of great societal importance due to its contribution to sea level rise (e.g. Stocker 2014). Satellite gravity data from the Gravity Recovery and Climate Experiment (GRACE) satellite mission (Tapley *et al.* 2004) has since 2002 provided important and unique information about the mass changes of both the large ice sheets (e.g. Velicogna 2009; Sasgen *et al.* 2012; Barletta *et al.* 2013; Velicogna *et al.* 2014) and smaller ice caps and glaciers (e.g. Chen *et al.* 2007; Luthcke *et al.* 2008; Jacob *et al.* 2012; Gardner *et al.* 2013), including those in Iceland.

The annual mass balance of the ice caps in Iceland has been estimated from field measurements since ~1990 (Björnsson &

Pálsson 2008), which is an ideal data set for validation of the ice mass changes derived from GRACE data. The reported mean annual mass loss of  $11.0 \pm 1.5$  Gt yr<sup>-1</sup> (Björnsson *et al.* 2013) of the ice caps in Iceland for the period 2003–2010 is large enough to be detected by GRACE, and the greatly enhanced mass loss of ~25 Gt in 2010 (Björnsson *et al.* 2013), a consequence of the Eyjafjallajökull eruption, should also be visible in the GRACE data. Some previous studies based on GRACE data (Jacob *et al.* 2012; Gardner *et al.* 2013) have found average mass balance estimates for Iceland which agree with the measured mean mass balance (Björnsson *et al.* 2013), but these have employed models of glacial isostatic adjustment (GIA) that predict little or no current response in Iceland. More recent studies have shown that the GIA response is likely to be substantial (Auriac *et al.* 2013), and this prompts us to take a

closer look at the GRACE signal over Iceland. In particular, we investigate the effects of applying a new local GIA correction for Iceland, and correcting for the disturbance from the close-by Greenland ice mass changes on the GRACE derived mass changes in Iceland.

Vatnajökull, the largest ice cap in Iceland, is situated only  $\sim 700$  km away from the southeastern coast of Greenland, which is an area that has experienced a large mass loss during the GRACE observational period (e.g. Khan *et al.* 2007; Howat *et al.* 2008; Wouters *et al.* 2008; Barletta *et al.* 2013; Velicogna *et al.* 2014). The monthly GRACE gravity models have a restricted spectral resolution, which results in a spatial spreading (leakage) of the gravity signals away from the region of mass change. Therefore, the gravity change signal generated by the mass loss in Greenland and Iceland may overlap, and since deriving mass changes from gravity observations is an ill-posed problem (having more than one solution), it is difficult to separate the mass changes occurring in the two areas; there might be a leakage of mass from one area into another (Baur *et al.* 2009). All methods that derive ice mass changes from gravity changes will be sensitive to this leakage effect, and it is therefore necessary to take leakage into account when using GRACE data, to reliably determine the ice mass changes in Iceland.

In the present study, we show the effect of leakage from the Greenland Ice Sheet on the GRACE-derived mass balance of the Icelandic ice caps. Different approaches to deal with the leakage problem have been proposed (e.g. Baur *et al.* 2009; Schrama & Wouters 2011; Zou and Jin 2014). One way is to forward model and remove the gravity signal of the disturbing sources by use of a priori knowledge. Here, we use an independent mass balance result of the Greenland Ice Sheet (Sørensen *et al.* 2011; Sasgen *et al.* 2012), which is based on data from the Ice, Cloud and land Elevation Satellite (ICESat) mission (Schutz *et al.* 2005) to minimize the gravity trend generated by the mass loss in Greenland, aiming at reducing the issue of mass loss signal leakage between Greenland and Iceland.

As mentioned above, another mass change component which must be considered, when aiming at isolating the gravitational signal from the ice caps in Iceland, is the mass re-distribution within the Earth due to GIA. Since the mantle beneath Iceland has a low viscosity (Barnhoorn *et al.* 2011; Auriac *et al.* 2013), the Earth is currently rebounding from the ice load changes which have occurred since the end of the Little Ice Age (LIA,  $\sim 1890$  AD) (Árnadóttir *et al.* 2009), while the effects of the ice changes on longer time scales are small or negligible here. This means that reliable information about the recent ice history and the Earth viscosity is crucial to estimate the GIA-induced gravity changes that will be observed by GRACE. In this study, we use the estimated mass changes of the Icelandic glaciers since  $\sim 1890$  (Björnsson *et al.* 2013) and the extensive ISNET GPS network data for crustal uplift (Árnadóttir *et al.* 2009) to model the local GIA signal and estimate the mass changes in the mantle associated with it.

We apply two different methods for deriving mass changes from the GRACE gravity models to determine whether the choice of methods affects the results.

## 2 GRACE DATA

We use monthly release 5 (RL05) GRACE gravity fields from the German research centre for Geosciences (GFZ; Dahle *et al.* 2012), spanning the time period January 2003 to April 2011, downloaded

via the <http://podaac.jpl.nasa.gov/GRACE> website. These fields are provided as sets of fully normalized Stokes' coefficients provided up to degree and order 90 but due to the larger uncertainties associated with the higher degree and order coefficients, we use only the coefficients up to degree and order 60.

Due to the single orbital plane of the GRACE satellites, noise manifests itself as along-track stripes in the monthly gravity models. Therefore, it is common practice to filter (or de-stripe) the models prior to any further analysis (e.g. Kusche *et al.* 2009). In this study though, we do not apply any other filter than the truncation at degree and order 60. For the data releases prior to RL05 (the one used here), the three official GRACE processing centres Center for Space Research (CSR), GFZ, and Jet Propulsion Laboratory (JPL) recommended the users to substitute the  $C_{20}$  coefficients in their monthly models with the ones obtained from satellite laser ranging (SLR) satellites (Cheng *et al.* 2013), but for RL05 we use the  $C_{20}$  provided in the GRACE files, as recommended by GFZ. As part of the product generation, atmospheric and oceanic variability is corrected for using a combination of the operational atmospheric fields from ECMWF and a baroclinic ocean model (OMCT). Also, the product is corrected for tides, and a static gravity field is subtracted (Dahle *et al.* 2012).

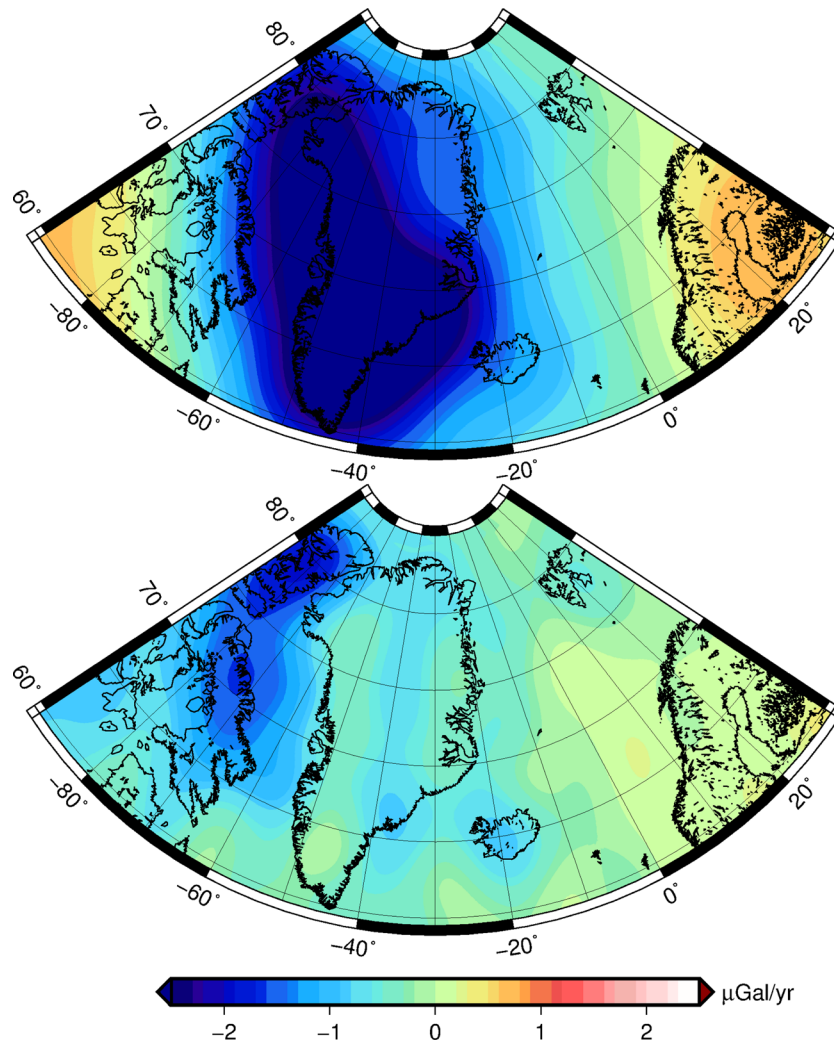
Furthermore, we include degree one coefficients computed as described by Swenson *et al.* (2008) for completeness, although the effect of this correction is negligible for the small area (Iceland) considered here.

## 3 METHODS

Several methods for deriving ice mass changes from GRACE data exist, and here we apply two different methods for studying the signal over Iceland; a conversion and an inversion method which are described in the following sections. We apply the methods on the gravity trend derived from GRACE observations in the time period 2003–2010. The gravity trend is shown in the upper panel in Fig. 1.

### 3.1 Ice mass changes from conversion

A widely used method for deriving mass changes on the Earth's surface from GRACE data is to create monthly surface maps of equivalent water thickness (EWT) from the gravity models following the methodology of Wahr *et al.* (1998). Mass change signals for a given region of interest are obtained by regionally integrating the GRACE data. As GRACE data consist of spherical harmonics, the regional integration function requires expansion into the same spherical harmonics before being applied, which introduces spatial leakage and thus complicates defining a well suited integration region. Our region function which takes a value of one over Iceland is displayed in Fig. 2(a). The GRACE data is also smoothed with a 320 km Gaussian filter before integrating and converting the solution to regional monthly EWT values. We process GRACE data (Section 2) accordingly and fit a linear trend to the monthly results, which leads to a regional mass-balance estimate. We note that different approaches for GRACE data conversion exist and could be utilized here, for example, converting the GRACE data to the spatial domain before applying a regional integration. However, as we will attempt to remove leakage into out domain of interest from Greenland by means of modifying Stokes' coefficients (*cf.* Section 3.3 below), we prefer the above outlined conversion method to only operate on the GRACE data in spherical harmonics space.



**Figure 1.** Change in gravity [ $\mu\text{Gal yr}^{-1}$ ] observed by GRACE (upper figure) and after removal of ICE-5G GIA model and mass changes in Greenland (lower figure).

### 3.2 Ice mass changes from inversion

The other method is the point mass inversion approach, in which mass gravity changes in satellite altitude are associated with point masses on the Earth's surface, introducing restrictions on the spatial extent of the mass loss signal to be the ice covered areas. The solution grid for the ice mass changes applied here is shown in Fig. 2(b) and covers both Greenland, Canadian Arctic, Iceland and Svalbard. All ice covered regions in the proximity of Iceland must be included to avoid a heavy signal leakage from these areas.

The inversion method is essentially based on eqs (1) and (2) and solves for point mass changes  $m_j = \underline{x}$  in pre-determined locations from gravity change observations  $\delta g_i = \underline{y}$ . The method is described in detail in (Forsberg & Reeh 2007; Sørensen & Forsberg 2010; Barletta *et al.* 2013).

$$\underline{y} = \delta g_i = G \sum_j m_j \frac{(h+a) - a \cos \psi_{ij}}{r_{ij}^3} = \mathbf{A} \underline{x}. \quad (1)$$

$G$  is the gravitational constant,  $a$  is the mean radius of the Earth,  $h$  is the height of the observation, and  $r_{ij}$  is the distance and  $\psi_{ij}$  the angle subtended from the centre of the Earth between the observation  $\delta g_{ij}$  and the mass point  $m_j$ . The inversion problem is ill-posed and

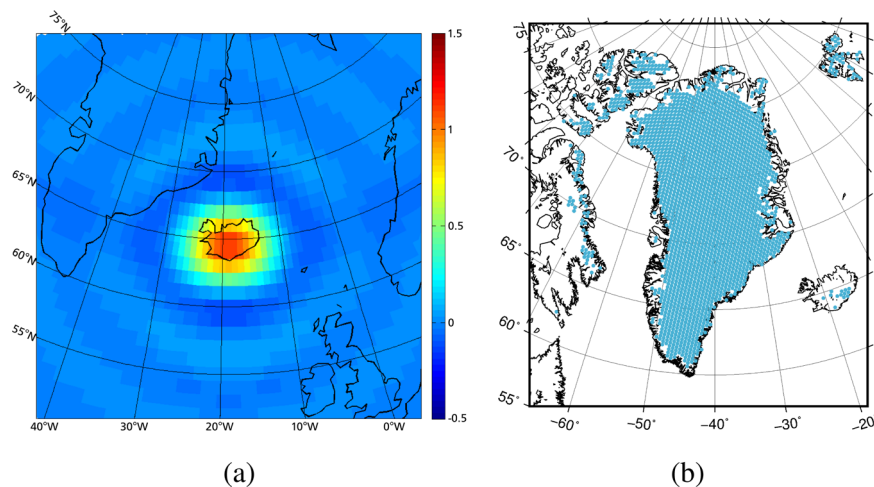
regularization is therefore needed (Tikhonov 1995) and the solution equation is

$$\underline{x} = (\mathbf{A}^T \mathbf{A} + \lambda \mathbf{I})^{-1} \mathbf{A}^T \underline{y}, \quad (2)$$

where  $\lambda$  is the parameter which controls the smoothness of the resulting point mass change solution. We have applied the regularization/smoothing parameter  $\lambda$  that was determined through extensive investigations on synthetic data in the Barletta *et al.* (2013) study (and which is described in detail in the supplement of Barletta *et al.* 2013). The regularization parameter used is 80 000. The total mass change is only slightly affected by the choice of  $\lambda$ , while the spatial distribution of the mass changes depends heavily on this. As our study is focused on total mass changes the choice of  $\lambda$  is less important.

### 3.3 Signal leakage from Greenland

During the last decade, the Greenland Ice Sheet and its surrounding glaciers and ice caps have had a negative mass balance of more than 200 Gt  $\text{yr}^{-1}$  (Sasgen *et al.* 2012; Shepherd *et al.* 2012; Barletta *et al.* 2013; Bolch *et al.* 2013; Gardner *et al.* 2013; Velicogna *et al.* 2014), with a large part of the ice mass loss occurring in



**Figure 2.** (a) Weighting function for integration in the conversion method. (b) Solution grid for the inversion of gravity to ice mass changes.

the southeastern part of Greenland. The associated large negative gravity trend in southeast Greenland spreads over the surrounding areas, and since Iceland is situated close to this area of high mass loss, the gravity signals from the two areas will potentially overlap. Therefore, it is likely that both methods used for deriving mass changes will fail in separating the two contributing areas (Greenland and Iceland), hence this spatial leakage must be considered when using GRACE data to constrain the mass changes.

We aim at minimizing this leakage effect by subtracting the Greenland signal by using an independent data set; the mean mass changes of the Greenland Ice Sheet for the period 2003–2009 derived from ICESat data (Sørensen *et al.* 2011; Sasgen *et al.* 2012). For the purpose of this study, we describe the 5 km grid of mass changes from Sasgen *et al.* (2012), which sums to  $-244 \text{ Gt yr}^{-1}$  over the entire Greenland Ice Sheet and surrounding glaciers and ice caps, as a spherical harmonic expansion up to degree and order 60 to be consistent with the GRACE monthly models. To do this, we use the theory described in Swenson & Wahr (2002), and subtract the resulting change in Stokes' coefficients from the GRACE monthly fields in order to separate the Greenland and Iceland signal and thereby reduce signal leakage.

### 3.4 Glacial isostatic adjustment

Iceland is located in a tectonically very active zone which is associated with a low viscosity upper mantle. A study by Auriac *et al.* (2013), used uplift observed from satellite radar interferometry and viscoelastic modelling to improve estimates of the Earth's rheology beneath Iceland, and they found mantle viscosities ( $\eta$ ) in the range  $4\text{--}10 \times 10^{18} \text{ Pa s}$ .

Such a low viscosity means that the traditional global ice history models, such as, for example, ICE-5G (Peltier 2004) that describes the deglaciation since the Last Glacial Maximum (LGM) in rather large time steps, will predict little or no present-day signal in Iceland. In low-viscosity zones, the recent (post-LIA) ice history might produce a significant present-day mass re-distribution signal, which must be taken into account in the GRACE analysis (e.g. Tamisieva *et al.* 2005; Nield *et al.* 2014).

To model the GIA-induced mass re-distribution for Iceland we employ a fast computing viscoelastic deformable Earth model (Bueler *et al.* 2007) which represents Earth's elastic response with

Green's functions and utilizes a spectral transfer function to estimate viscous mantle flow overlaid by an elastic plate.

Since pre-LIA unloading is of little importance for modelling uplift rates in Iceland (Fleming *et al.* 2007), we drive the Earth model by three different estimated ice unloading histories for Iceland, spanning from the LIA to 2012: (1) an Iceland-wide average annual mass balance is computed by tuning a simple degree-day mass balance model (Marzeion *et al.* 2012) to existing records (Björnsson *et al.* 2013) and forcing the model with weather station data from Stykkishólmur (1890–2010); (2) a local study for South-Iceland (Aðlgeirsdóttir *et al.* 2011) is upscaled to reflect total, countrywide mass changes; (3) the known mass changes since the LIA of Langjökull ice cap (Björnsson *et al.* 2013) are upscaled to reflect total, countrywide mass changes. All three unloading histories are calibrated to reproduce an estimate of the total ice mass loss of 500 Gt in Iceland between 1890 and 2012 (Björnsson & Pálsson 2008; Björnsson *et al.* 2013).

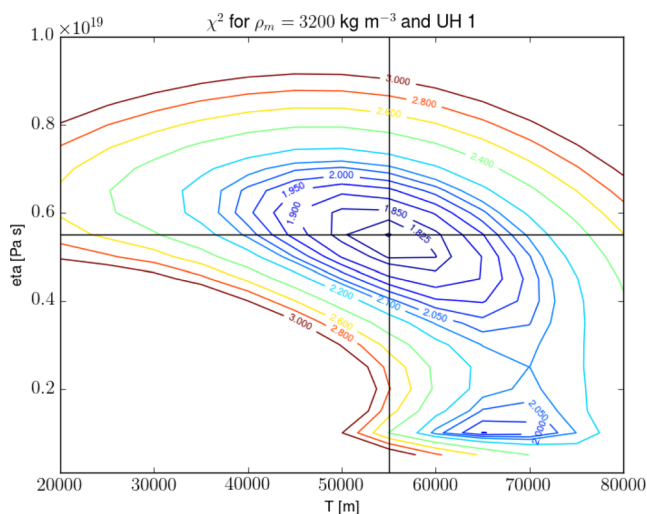
Several parameters within the Earth model require tuning. The elastic behaviour of Earth's crust is represented in the model by defining a flexural rigidity  $D = (ET^3)/[12(1 - \nu^2)]$  with  $E$  being the Young's modulus,  $\nu$  the Poisson's ratio and  $T$  the thickness of the elastic layer. As the model only works with  $D$ , attempting to tune for  $E$ ,  $\nu$ , and  $T$  simultaneously is futile due to the non-unique nature of this formulation. Therefore we set  $E = 40 \text{ GPa}$  and  $\nu = 0.25$  (Árnadóttir *et al.* 2009) and only vary  $T$ . For the viscous mantle flow we vary the mantle viscosity,  $\eta$ , and the mantle density,  $\rho_m$ .

The uplift rates in Iceland are modelled on a 2 km resolution grid by forcing the Earth model with the above mentioned unloading scenarios equally distributed over the 2012 glacier extents. The ISNET GPS reference station network (Árnadóttir *et al.* 2009) is used to assess the success of the modelled vertical uplift rates between 1993 and 2004. Model tests with all unloading scenarios, where the ice-mass loss is distributed over an area close to the glacier margins, were also done. These resulted in a worse fit to the GPS uplift data and thus are omitted here. We utilize the same model performance measure,  $\chi^2$ , as Árnadóttir *et al.* (2009), who reported their best fit being  $\chi^2 = 1.85$ , and we are able to demonstrate a successful reproduction (similar value of  $\chi^2$ ) of the measured uplift rates for Iceland. By minimizing the  $\chi^2$  measure for a given unloading history and set value of mantle density,  $\rho_m$ , we infer the model parameters  $T$  and  $\eta$  as well as the post-LIA GIA



**Table 1.** Results from our GIA modelling are displayed in column 5, the GIA correction to be applied for Iceland. Mantle viscosity ( $\eta$ , column 3) and crustal thickness ( $T$ , column 4) are estimated by minimizing the  $\chi^2$  performance measure for a set unloading history (UH, column 1) and mantle density ( $\rho_m$ , column 2).

UH	$\rho_m$ ( $\text{kg m}^{-3}$ )	$\eta$ ( $1\text{e}18 \text{ Pa s}$ )	$T$ [km]	GIA corr. ( $\text{Gt yr}^{-1}$ )	$\chi^2$
1	3200	5.5	55	5.5	1.79
2	3200	6.5	55	5.6	1.79
3	3200	5.0	55	5.6	1.80
1	3400	5.5	55	5.6	1.85
2	3400	6.5	55	5.6	1.82
3	3400	5.0	55	5.6	1.83
1	3000	5.5	55	5.4	1.80
2	3000	6.5	55	5.5	1.80
3	3000	5.0	55	5.5	1.85



**Figure 3.** Model performance measure  $\chi^2$  as contours for unloading history 1 and mantle density  $\rho_m = 3200 \text{ kg m}^{-3}$  as an example of our earth model tuning. Best fit, that is, lowest  $\chi^2$  value, is marked with black lines.

correction for Iceland. The model results are summarized in Table 1. Model performance ( $\chi^2$ ) for unloading history 1 and mantle density  $\rho_m = 3200 \text{ kg m}^{-3}$  is shown in Fig. 3 as contours with the best fit (lowest  $\chi^2$  value) marked as an example for our model tuning procedure. Estimated values for mantle viscosity are within the range determined by Auriac *et al.* (2013) and the values for the mantle density and thickness of the elastic layer are realistic.

Summarizing our GIA modelling efforts outlined here, we estimate that a post-LIA GIA correction of  $5.5 \pm 1.0 \text{ Gt yr}^{-1}$  should be added to the GRACE results. Since the GIA signal is caused by mass inflow to the area of interest, applying the GIA correction will result in an increased ice mass loss compared to not correcting for GIA.

## 4 GRACE MASS CHANGE RESULTS

The derived gravity trend from GRACE observations in the time period 2003–2010 over a region covering Greenland, Iceland, the Canadian arctic and Scandinavia is shown in the upper panel in Fig. 1. This shows some distinct signals; the mass loss of Greenland is clearly dominating the picture with a strong negative trend in gravity while positive trends are seen over Scandinavia and in the southwestern part of the map.

The lower panel in Fig. 1 shows the same observed trend, after subtracting the gravity trend signal caused by the Greenland mass loss (see Section 3.3) and the ICE-5G GIA model contribution (Peltier 2004). It is seen that a gravity signal over Iceland becomes evident when the leakage from Greenland is removed, and similarly a gravity trend in the Canadian Arctic and over Svalbard stands out. Furthermore, the positive trends seen in Scandinavia and in the southwestern part of the map have been greatly reduced by the subtraction of the ICE-5G GIA correction.

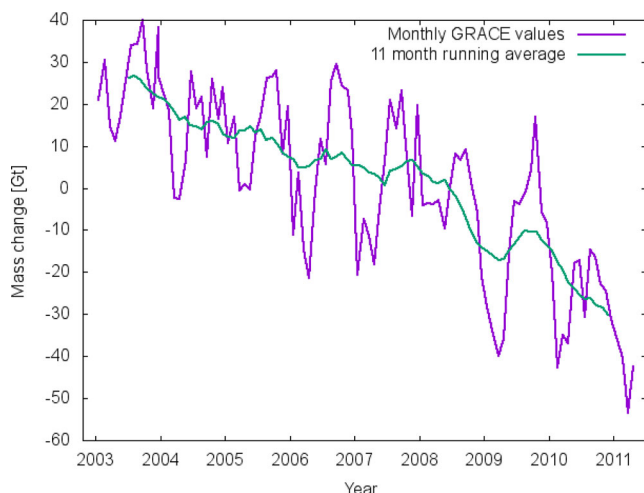
Table 2 shows several GRACE-derived mass balance estimates for Iceland, based on the conversion and inversion methods. Table 2 contains estimates obtained both with and without reducing the leakage from Greenland and correcting for the post-LIA GIA signal. The error estimates of the GRACE derived ice mass loss rates are estimated from running a suite of calculations of the gravity field to solve for mass change rates, including the error estimates that are provided with the data (standard deviations on each Stokes coefficient), and applying different filters on the GRACE data. This resulted in a small error of  $\pm 1 \text{ Gt yr}^{-1}$ , and to this we added a  $\pm 1 \text{ Gt yr}^{-1}$  to account for a possible error from the fact that we cannot be sure that the Greenland contribution is completely removed. Table 2 shows that if the signal leakage from the Greenland mass loss is not accounted for, the inversion method and conversion methods give different mass balance results for Iceland of  $-3.9 \pm 2$  and  $-9.4 \pm 2 \text{ Gt yr}^{-1}$ , respectively. It is also seen that if the Greenland signal is subtracted, the two methods give identical numbers of  $-5.9 \pm 2 \text{ Gt yr}^{-1}$ . The post-LIA correction affects the two methods in the same way, which is why the mass balance estimates from the two methods agree on a mass balance of  $-11.4 \text{ Gt yr}^{-1}$  if both leakage is reduced and post-LIA GIA is corrected for.

### 4.1 Comparison with mass balance measurements

The mean annual Iceland ice mass change results derived from GRACE (2003–2010), that are listed in Table 2, can be directly compared with the mass balance found from field measurements of  $11.0 \pm 1.5 \text{ Gt yr}^{-1}$  for the same time period (Björnsson *et al.* 2013). Each year the winter and summer surface mass balance is measured at  $\sim 100$  chosen survey sites on the three largest ice caps (Björnsson *et al.* 1998, 2002; Jóhannesson *et al.* 2013) that cover about 90 per cent of the total glaciated area in Iceland. The mass balance for each ice cap is estimated from the measurements (uncertainty estimated to be 5–15 per cent) and extrapolation is used

**Table 2.** Mass balance estimates (2003–2010) of the Icelandic ice caps derived from GRACE data using two different methods, with and without correcting for leakage and post-LIA GIA.

Signal from Greenland	GIA correction	Mass balance [ $\text{Gt yr}^{-1}$ ] from inversion	Mass balance [ $\text{Gt yr}^{-1}$ ] from conversion
Not subtracted	ICE-5G	$-3.9 \pm 2.0$	$-9.4 \pm 2.0$
Subtracted	ICE-5G	$-5.9 \pm 2.0$	$-5.9 \pm 2.0$
Subtracted	ICE-5G and post-LIA	$-11.4 \pm 2.2$	$-11.4 \pm 2.2$



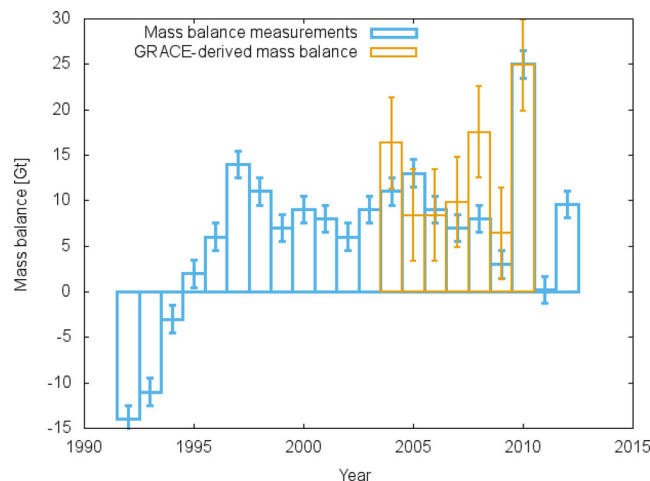
**Figure 4.** Monthly mass changes of Icelandic ice caps derived from GRACE using the inversion method. Also shown is an 11 month running average of the monthly mass changes.

to estimate total mass balance for the missing 10 per cent of the glaciated area.

When leakage from Greenland is minimized and post-LIA GIA is taken into account as described in Section 3.4, both methods result in a mass balance that agrees well with the measured mass balance.

Based on this, we investigate whether it is also possible to extract reliable mass change results from GRACE on an annual basis, as it was possible for the multi-year trend. The monthly mass changes derived from the inversion method after reducing leakage from Greenland is shown in Fig. 4. These monthly mass change estimates exhibit a high (and unrealistic) variability. This is to be expected due to several reasons. The GRACE monthly gravity models are corrected for ocean and atmospheric mass variability via models, but these are not flawless and any errors in these will be mapped directly into the derived ice mass changes. The small magnitude of the Iceland ice mass change signal combined with its geographical location makes it sensitive to such errors. Furthermore, since we have removed only the mean annual mass trend from Greenland for 2003–2009, there will still be some degree of seasonal mass variability coming from this region, possibly influencing the Iceland time-series. One approach to reduce these rapid variations in the mass change time series is to do a running average with a window of a given size. The result of such a running average operation with a window size of 11 months is also shown in Fig. 4. In order to estimate annual mass balance, annual differences over glaciological years are derived from the smoothed curve, and the results are shown together with the mass balance measurement record in Fig. 5. Due to the noisy behaviour of the monthly data, the annual GRACE-derived mass balance estimates are associated with a rather large uncertainty of  $\pm 5$  Gt. For comparison, the uncertainty on the measured mass balances is estimated to be  $\pm 1.5$  Gt. The error on the GRACE-derived yearly mass balance estimates shown in Fig. 5 was chosen conservatively to  $\pm 5$  Gt  $\text{yr}^{-1}$ . This choice was based on a similar approach as described above, adding to this also the uncertainty introduced by varying the temporal smoothing of the time series, and acknowledging that on short time scales, possible errors in atmosphere and ocean models might influence the results.

The unusually large mass loss of 25 Gt in 2010, which is evident in the measured mass balance, is clearly visible in the GRACE-derived mass balance results (Fig. 5). This is also seen as a



**Figure 5.** Yearly mass balance measurements shown with GRACE-derived annual mass balance estimates.

rapid mass loss in the smoothed curve in Fig. 4. The GRACE-derived annual mass balance estimates agree well with the observed mass balance for all the years considered except for 2008. Then GRACE results indicate a large mass loss of  $\sim 17$  Gt but observations show negative mass balance of  $\sim 8$  Gt.

## 5 DISCUSSION

Following the method described in Section 3.3 we aim at removing the gravity signal caused by Greenland ice changes, but it is clear from Fig. 1 (lower) that the gravity signal over Greenland is not zero everywhere after this correction has been applied. This implies that the full mass change signal over Greenland has not been removed by the ICESat-derived mass change. This is expected because the ICESat results are the mean annual mass changes of the period 2003–2009, which is exceeded by more than one year in the GRACE data period investigated here. For 2010, we assume the same mass trend as observed for the period 2003–2009. Also, there are of course uncertainties associated with the ICESat mass change estimate. It is noted that even though the full signal is not removed, the leakage into Iceland is indeed significantly reduced.

We find that the two methods applied are affected differently by the signal leakage from Greenland. The inversion method tends to place too much mass change in Greenland if the leakage is not reduced, resulting in an Icelandic ice mass loss estimate of only  $3.9 \pm 2.0$  Gt  $\text{yr}^{-1}$ . This is probably a consequence of the smoothness constraints in the inversion. The conversion method on the other hand predicts a larger Iceland mass loss prior to correcting for leakage than after; a direct consequence of the gravity trend leakage in Fig. 1 (upper) and the integration area in Fig. 2(a). If the signal leakage from Greenland is successfully removed, we would expect the two methods to provide similar mass change results and this is indeed the case with a mass loss of  $5.9 \pm 2.0$  Gt  $\text{yr}^{-1}$ . The post-LIA correction affects the two methods in the same way, which is why the mass balance estimates from the two methods also agree with a mass loss estimate of  $-11.4 \pm 2.2$  (Table 2) if both leakage is reduced and post-LIA GIA is corrected for.

Based on regional post-LIA GIA modelling we estimate a present-day GIA correction of  $5.5 \pm 1.0$  Gt  $\text{yr}^{-1}$ . This number differs significantly from what others have used. For example, Jacob *et al.* (2012) and Gardner *et al.* (2013) estimate an ‘upper bound’ post-LIA GIA correction for GRACE over Iceland to be

$\sim 0$  Gt yr<sup>-1</sup>. The large difference between the GIA correction derived and applied here and that of Jacob *et al.* (2012), must be caused by differences in ice history and/or mantle rheology. Jacob *et al.* (2012) assumed an ice history of the Icelandic ice caps based on Dyurgerov (2010). The volume of ice added and subtracted during the glaciation and de-glaciation is the same, assuming the mean annual mass loss rates for 1961–2006 of  $-1.96$  km<sup>3</sup> yr<sup>-1</sup> for Iceland during the entire de-glaciation, as described by Dyurgerov (2010) the total mass loss sums up to  $\sim 290$  Gt, which is significantly less than what is used here ( $\sim 500$  Gt from Björnsson *et al.* 2013). Our GIA modelling also demonstrates that using a linear deglaciation since the LIA would result in a significantly worse agreement of predicted uplift rates in comparison with the GPS network data, that is,  $\chi^2$  values ranging between 2.2 and 2.4. Hence a non-linear, realistic unloading history as used in our study is required. Other differences in the GIA computations are the ice geometry of the ice loading/unloading and the Earth models used. For example, the Earth model used in Jacob *et al.* (2012) consists of a lithosphere of 65 km, and an asthenosphere, upper mantle and lower mantle with viscosities of  $1 \times 10^{19}$ ,  $3 \times 10^{20}$  and  $2.4 \times 10^{21}$  Pa s, respectively. These are considerably higher values than Auriac *et al.* (2013) and the present study infer from the local ISNET GPS network data for model validation. Such lower values have been confirmed by a study independent of GIA modelling (Barnhoorn *et al.* 2011).

As discussed above, the mass loss signal of the Icelandic glaciers represents a modest area and mass loss signal to be observed by GRACE, making it particularly sensitive to errors in the background models (e.g. atmosphere and ocean) used in the GRACE processing. This will probably not be manifested in the trend (multi-year average mass balance) while the monthly solutions are likely influenced. Therefore, it is basically more difficult to derive annual mass balance estimates than multi-year. The results shown in Figs 4 and 5 show exactly this, the mass balance estimate for the year 2008 is significantly larger than the measured one. The 2010 signal, however, stands out in the GRACE results and agrees well with the observations.

In our processing we have not taken into account storage and delay of the glacier melt runoff, for example due to a few large hydro-power water reservoirs in Iceland. This will introduce an additional uncertainty in the annual mass loss estimates derived from GRACE. We estimate that the amount of run-off that is delayed by the three largest reservoirs operated by Iceland's national hydropower company, Landsvirkjun, amounts to  $\sim 2$  Gt (see Fljótisdalur, Blanda and Búrfell stations at <http://www.landsvirkjun.com/company/powerstations>). Háslón, serving Fljótisdalur station, is the largest of these water reservoirs and is located close to Vatnajökull, and was taken in use in 2008. The effect of the delayed runoff does not appear as a clear signal in the GRACE results after 2008, which indicates that the size of this uncertainty is not resolved in these observations.

As mentioned in Section 2 we have chosen to not apply a filter or de-stripping approach to the GRACE data. This choice was based on a thorough investigation of the influence that different filters could have on our mass change results. It was concluded that the mass balance derived varied only  $0.5$  Gt yr<sup>-1</sup> when applying the different DDK filters (Kusche *et al.* 2009), probably because these are designed to be mass conserving, and affect the spatial distribution but not the strength of the gravity trend signal. We work on a relatively long time trend, and furthermore we work with Stokes' coefficients up to only degree and order 60, which means that the stripes are reduced due to the stacking of data because the noise is correlated

along the satellite track but is not correlated in time. Lastly, we visually evaluated each of the gravity trend fields generated using the different filters and using no filter, and found the spatial pattern in the trend using no filter was physically most convincing with negative trends located over glaciated areas.

## 6 CONCLUSIONS

We employ monthly GRACE gravity models to infer mass changes of the ice caps in Iceland in the period 2003–2010. Our results show that in order to estimate the local ice mass change correctly it is necessary to carefully correct for both Glacial Isostatic Adjustment and signal leakage from the mass loss in Greenland.

We apply two different methods for deriving ice mass changes from GRACE data and find that by applying these directly on the gravity trend from GRACE, they yield different numbers of  $-3.9 \pm 2.0$  and  $-9.4 \pm 2.0$  Gt yr<sup>-1</sup>. We identify signal leakage from the large mass loss in southeast Greenland to be the primary reason for this discrepancy, and we therefore reduce this leakage by subtracting the Greenland signal by use of an independent mass change estimate derived from ICESat laser altimetry and snow/firn modelling. We find that when correcting for signal leakage both methods obtain the same mass balance for the ice caps in Iceland, but the number of  $-5.9 \pm 2.0$  Gt yr<sup>-1</sup> is only about half of the mass loss found from in-situ measurements ( $11 \pm 1.5$  Gt yr<sup>-1</sup>). We demonstrate that a regional model of post-LIA GIA, based on our current best knowledge of ice history and Earth model parameters, helps to reconcile the GRACE-derived and measured mass balance. Our conclusion that the present-day GIA signal is significant and is indeed important to correct for, is in contrast to other studies (e.g. Jacob *et al.* 2012) that have assumed no GIA signal. The annual mass balance estimates of the Icelandic ice caps are derived from the GRACE data, and the time-series show that these can be significantly affected by modelling errors, although the unusually high mass loss in 2010, due to reduces albedo and enhanced melting as a consequence of the volcanic ash distributed over the glaciers, is clearly visible.

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