Glacial isostatic adjustment in the static gravity field of Fennoscandia

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Abstract

In the central part of Fennoscandia, the crust is currently rising, because of the delayed response of the viscous mantle to melting of the Late Pleistocene ice sheet. This process, called Glacial Isostatic Adjustment (GIA), causes a negative anomaly in the present-day static gravity field as isostatic equilibrium has not been reached yet. Several studies have tried to use this anomaly as a constraint on models of GIA, but the uncertainty in crustal and upper mantle structures has not been fully taken into account. Therefore, our aim is to revisit this using improved crustal models and compensation techniques. We find that in contrast with other studies, the effect of crustal anomalies on the gravity field cannot be effectively removed, because of uncertainties in the crustal and upper mantle density models. Our second aim is to estimate the effects on geophysical models, which assume isostatic equilibrium, after correcting the observed gravity field with numerical models for GIA. We show that correcting for GIA in geophysical modelling can give changes of several kilometer in the thickness of structural layers of modeled lithosphere, which is a small but significant correction. Correcting the gravity field for GIA prior to assuming isostatic equilibrium and inferring density anomalies might be relevant in other areas with ongoing postglacial rebound such as North America and the polar regions.

1. Introduction

The static gravity field of Fennoscandia is known with high accuracy (± 2 mGal) at a wavelength of 100–200 km [Mayer-Gürr et al., 2012]. Figure 1 shows the satellite-derived free-air anomaly observed in the Fennoscandian region. The negative gravity anomaly observed in the Bay of Bothnia (for geographical references, see Figure 2) is linked to the Glacial Isostatic Adjustment (GIA) in that area [Heiskanen and Vening Meinesz, 1958; Balling, 1980; Sjöberg et al., 1994]. However, the characteristic elliptic shape of GIA is not immediately visible in the static gravity field. Other sources contribute to the gravity field, such as density anomalies in the crust or mantle, crustal thickening, or even a long-wavelength feature related to the Icelandic mantle plume [Marquart, 1989; Steffen and Wu, 2011]. The question that arises is whether, given the contribution of all these sources, it is possible to extract information from static gravity field observations to constrain GIA modeling. In the following, the free-air gravity anomaly that exists at present due to incomplete rebound will be referred to as the GIA gravity signal or gravity anomaly.

Several studies have aimed to separate the GIA gravity signal from the observed gravity field. However, the conclusions of these studies differ significantly. Heiskanen and Vening Meinesz [1958] estimated a contribution of −20 mGal to the gravity anomaly but stated that local density anomalies made the problem very complex. Honkasolo [1964] could not find a spatial correlation between the gravity anomaly and the region of GIA induced uplift. Balling [1980] concluded that −15 to −20 mGal in the area could be explained by postglacial uplift after removing the gravity anomaly correlated with topography. This number was subsequently used to constrain mantle viscosities in GIA modeling [Peltier and Wu, 1982; Wu and Peltier, 1983; Mitrovica and Peltier, 1989]. However, Anderson [1984] suggested that crustal structures explain a large part of the observed gravity anomaly and should be accounted for. After removing the effect of crustal thickening, the negative anomaly in the gravity field was almost completely removed [Marquart, 1989], supporting the conclusion that crustal thickening is dominant in the gravity field. A more elaborate method [Sjöberg et al., 1994] improved on this study and concluded that postglacial uplift was responsible for −28 mGal and crustal thickening for −12 mGal in the observed gravity anomaly. In a recent study, Sjöberg and Bagherbandi [2013] apply bandpass filtering to the spherical harmonic coefficients, assuming the GIA signal resides...
mainly between coefficients of degree and order 10 and 70. They obtain a negative geoid anomaly of $-12\,\text{m}$, which corresponds to a negative free-air anomaly of $-30\,\text{mGal}$ in central Fennoscandia.

The largest limitation of most of these studies was the use of a homogeneous crust [Kakkuri and Wang, 1998]. With the increasing number of seismic measurements in the area, it was possible to construct an improved crustal density model. With this improved crustal model, Kakkuri and Wang [1998] concluded that the negative gravity anomaly produced by crustal effects was actually twice as large as the observed gravity anomaly and the density variations in the upper mantle made a dominant contribution to the observed gravity field. A more recent study [Kaban et al., 2010], with improved crustal models shows a positive residual gravity anomaly in the Bay of Bothnia after the removal of crustal effects, leaving no contribution for the GIA gravity signal. Ebbing [2007] proposed a model for the lithosphere and mantle in which crustal thickness variations were compensated by intracrustal densities to achieve isostatic equilibrium. This model leaves up to $-20\,\text{mGal}$ unexplained in the area of GIA.

Figure 1. The free-air gravity anomaly (mGal) from the GOC03S [Mayer-Gürr et al., 2012] model up to spherical harmonic degree and order 180.

Figure 2. These figures represent the datasets used in this study. (a) The topography of Northwestern Europe with the locations of the regional relative sea level (RSL) observation sites (red diamonds) that are used in this study. (b) The location of the GPS stations with the value of vertical motion [Lidberg et al., 2007]. The white circles are error estimates. Red bars show uplift and blue bars represent measured subsidence. (c) Gravity rate estimated from GRACE data for the period January 2003 to December 2011 after smoothing with 400 km Gaussian filter halfwidth.
Figure 3. Misfit with respect to the selected (see Appendix B) relative sea level data (RSL) of Tushingham and Peltier [1992] in Fennoscandia for the GIA model with varying upper and lower mantle viscosity with three different lithosphere thicknesses: (a) 80 km, (b) 115 km, and (c) 150 km. Parameter combinations for which the misfit falls within the 95% confidence region are shown by circles. The dashed red area represents the range of viscosities of VM2 [Peltier, 1996], and the green solid line represents the VM5a [Peltier and Drummond, 2008] viscosity model. Viscosities are in Pa s.

Given these different conclusions, the question is still open on whether it is possible to separate the gravity effect caused by GIA from other sources. This study explores the possibilities to observe and separate the GIA gravity signal from static gravity field observations using improved crustal models of Fennoscandia and various assumptions of isostatic compensation. Furthermore, we will investigate the opposite question, whether the effect of a GIA correction in a geophysical model, which assumes isostasy, is significant.

2. GIA Model

The magnitude and uncertainty of the GIA gravity effect is computed for comparison with the uncertainty in the gravity field models introduced by current density models of the crust. The GIA model is based on the normal mode method [Wu and Peltier, 1982] for a multilayer model [Vermeersen and Sabadini, 1997] with self-consistent sea levels [Mitrovica and Peltier, 1991]. Details are given in van der Wal et al. [2009]. Geocenter motion has been added but not rotational feedback, as this latter effect is quite small in regional studies. The ice model ICE-5G is used as ice loading model [Peltier, 2004], representing one full glacial cycle up to 122 ka. Adding another glacial cycle does not significantly change the selection of preferred GIA models. Also, the remaining gravity anomaly and deflection of the lithosphere in Fennoscandia are not affected significantly. To support these statements, a worst-case scenario was modeled. The results in the remainder of this study will be modeled with only one ice cycle. For details, see Appendix A. The GIA models that have the lowest misfit to relative sea level data, GPS-based uplift rates, and GRACE observed gravity change, will be used in the rest of the study.

An incompressible Maxwell rheology is used, where the viscosity is varied in two layers of the mantle separated at the 670 km discontinuity. These layers can be resolved using misfit minimisation of sea level data. Paulson et al. [2007] found a small effect of compressibility on misfit for GRACE data and RSL data in North America. We assume the same holds for Fennoscandia, in which case the choice of preferred models is not affected. Neglecting compressibility does affect the displacement and induced gravity anomaly of our models, but we think that the estimated uncertainties from variations in viscosity are large enough to accommodate the effect of compressibility. Upper and lower mantle viscosities are varied from $2 \times 10^{20}$ Pa s to $512 \times 10^{20}$ Pa s in steps of a factor 2. The elastic lithosphere is modeled with a thickness of 80, 115, and 150 km, to sample the range of plausible values for lithosphere thicknesses in Fennoscandia [Whitehouse, 2009; Steffen and Wu, 2011]. There are three issues and assumptions in our modelling that need to be addressed. First, ICE-5G implicitly uses a certain mantle viscosity profile. However, since ICE-5G is created in a global sense, it is possible that regionally the rheology deviates and regional studies have commonly varied viscosity for a fixed ice loading history [Milne et al., 2001; Wang et al., 2008; Steffen et al., 2010]. Second, the uncertainty in the ice loading history is not known, and third, the mantle viscosity does not only vary in radial directions. To deal with the last two issues, we also compare output from a GIA model that uses an ice model that was created independently from mantle rheology, with our results. Viscosity in the Earth model component of the GIA model varies laterally as well as radially. Misfit analysis in van der Wal et al. [2013] already yielded preferred models, and the model output is used as is.
Figure 4. Misfit with respect to GPS data from the BIFROST project [Lidberg et al., 2007] with three different lithosphere thicknesses: (a) 80 km, (b) 115 km, and (c) 150 km. Viscosities are in Pa s.

Models are searched which minimise a chi-squared misfit with respect the observational data, defined as

\[ \chi^2 = \frac{1}{N} \sum_{i=1}^{N} \left( \frac{o_i - p_i}{\sigma_i} \right)^2 \]  

Here, \( N \) is the number of observations, \( o_i \) are the observations, \( p_i \) are the predicted values from the models interpolated at the measurement locations, and \( \sigma_i \) are the standard deviations corresponding to the observations. Certain models with a different viscosity distribution can result in a misfit which is statistically indistinguishable from the best-fitting model. To obtain a confidence region, we use a contour that contains 95% of the probability distribution of the parameters [Press et al., 1992]. For the 2 degrees-of-freedom in these models, this is represented by a \( \chi^2 \) misfit contour that differs less than 5.99 with the minimal misfit value.

As the first data source, we use the relative sea level (RSL) data from the database of Tushingham and Peltier [1992] to constrain the GIA modelling. Only RSL sites sensitive to Fennoscandia deglaciation are used to obtain an estimate for the regional viscosity profile. The used geological sites are shown in Figure 2a, and the RSL curves can be found in Appendix B. For \( \sigma_i \), both the height errors and timing errors are incorporated in a similar fashion as van der Wal et al. [2011]. One area of low misfit is close to the viscosity profiles of VMSa [Peltier and Drummond, 2008] and VM2 [Peltier, 1996], depicted in Figure 3 by the green line and the red dashed area, respectively. Also, the average viscosity found in the study by Mitrovica [1996] of 0.65–1.10 \( \times 10^{21} \) Pa s is within this region. Furthermore, there are no significant changes between the results of different lithosphere thicknesses. The results show that this particular set of RSL observations is not capable to distinguish between lithosphere thicknesses in our area of interest. Part of the reason is that different subsets of the Fennoscandian RSL data set prefer different lithosphere thicknesses [Steffen et al., 2014] possibly as a result of actual variation in lithosphere thickness. The second region of low misfit, which contains the minimum misfit, is found for models with an upper mantle viscosity of 1.6–3.2 \( \times 10^{21} \) Pa s and a lower mantle viscosity of 5.12 \( \times 10^{22} \) Pa s. These values are situated in the top-right area in Figure 3. It is known that the lower mantle viscosity cannot be constrained by Fennoscandian data; therefore, it will not be discussed further. The upper mantle viscosity is at the high end of the viscosity ranges in most studies for Scandinavia. We do note that some of the RSL curves predicted by the model with high upper-mantle viscosity do not fit well to the data (Appendix B).

This bifurcation in the misfit plots cannot be resolved solely using RSL observations. Therefore, a similar comparison is made between the GIA models and GPS determined uplift rates at 53 sites in Fennoscandia from the BIFROST project; see Table 1 in Lidberg et al. [2007]. Misfit values are computed using equation (1). The preferred misfits are denoted by the solid circles in Figure 4. Here, a confidence region of 99% is chosen.
Figure 5. The standard deviation of the difference between the models and GRACE time-variable gravity observations with three different lithosphere thicknesses: (a) 80 km, (b) 115 km, and (c) 150 km. The circles in the plot denote standard deviations smaller than 0.08 μGal/yr. Viscosities are in Pa s. The black dashed areas are the selected viscosity profiles for model set A and model set B. (9.21 away from the minimal misfit), because of the small uncertainties in the GPS measurements and the large misfit values. This comparison shows a different misfit pattern than for the RSL results, but the bifurcation is still visible. Fewer GIA models offer an explanation for the GPS data, as only two to three models pass the misfit criteria. The misfit becomes smaller with increasing lithosphere thickness.

To complete the GIA model selection, the GIA model results are also compared with gravity change observations of the GRACE satellite mission. We use CSR release 5 solutions from September 2002 to November 2012. The observations are filtered with a Gaussian filter with 400 km halfwidth (see Figure 2c). In this case, the chi-squared misfit of equation (1) could not be used, since there is no meaningful standard deviation corresponding to the GRACE observations in Fennoscandia. Formal errors, computed according to the procedure of Wahre et al. [2004], are largely constant across Fennoscandia, around 0.02 μGal/yr, but systematic errors are not well known. Therefore, the standard deviation of the difference between the observation and model results is shown in Figure 5 instead. The solid circles represent comparisons with a standard deviation less than 0.08 μGal/yr. These figures show similar results as in the RSL and GPS cases. Again two regions where the models represent the observations best can be observed in Figure 5. Furthermore, a slight decrease in standard deviation is noticed with increasing lithosphere thickness. The GRACE data confirm the conclusions from the RSL and GPS measurements but are also not able to distinguish between the two regions of low misfit.

Figure 6. (a) Current free-air gravity anomaly due to GIA (Model B). (b) Remaining uplift for the different GIA models. Black lines represent model A with the thick black line being the best fitting model as described in the text ($\nu_{um} = 8 \times 10^{20}$ Pa s and $\nu_{lm} = 16 \times 10^{20}$ Pa s). Red lines represent model B with the thick red line being the best fitting model ($\nu_{um} = 16 \times 10^{20}$ Pa s and $\nu_{lm} = 256 \times 10^{20}$ Pa s). The blue dashed line represents the GIA model with dry composite rheology with grain size of 10 mm. The green dashed line represents a GIA model with dry composite rheology with a grain size of 4 mm [van der Wal et al., 2013].
By inspecting the RSL, GPS, and GRACE observations in the comparisons with the GIA models results, two sets of GIA models are selected, model set A and model set B. Both models have a lithosphere thickness of 150 km, due to the observed slight decrease in misfit with GPS and GRACE data for an increasing lithosphere thickness. Model set A represents the choice of viscosities that agrees with viscosity profiles VM2 and VMSa. Model set B represents a viscosity profile with larger upper and lower mantle viscosity. The GIA models that are used in the construction of these two models are shown in Figure 5 by the dashed black areas.

They were selected because most of the GIA models are situated in minimum misfit contours of the RSL comparison. The GIA model with viscosity profile, $\nu_{um} = 8 \times 10^{20}$ Pa s and $\nu_{lm} = 32 \times 10^{20}$ Pa s, was added to model set A, because of the evidence in the misfit comparison of GRACE data.

To illustrate the models, Figure 6a shows the free-air gravity anomaly for the GIA model with viscosity profile, $\nu_{um} = 16 \times 10^{20}$ Pa s and $\nu_{lm} = 256 \times 10^{20}$ Pa s (Model B). The shape of the gravity anomaly is similar to the measured uplift rates [Lidberg et al., 2007]. The minimal value of the GIA static gravity anomaly can be found in the Gulf of Bothnia. This extreme is close to where the maximum ice thickness of the Late Pleistocene ice sheet is believed to have been situated. This is also the location of maximum gravity change, according to the GRACE gravity measurements and corroborated by the GPS uplift measurements, which observe maximal uplift in that area. Furthermore, Figure 6b shows the different remaining uplift or deflection of the lithosphere results of the selected models at cross section AA'. A clear distinction between model set A and B is noticed. Model set A has a minimum deflection of $-57$ m ($\pm 20$ m), and a minimum free-air gravity anomaly value is $-8.2$ mGal ($\pm 2.9$ mGal). Model set B shows a minimum deflection of $196$ m ($\pm 21$ m) and a corresponding minimum free-air gravity anomaly value of $-28.3$ mGal ($\pm 3.0$ mGal). These values are constructed by calculating the mean deflection and gravity anomaly of the model set, whereas the uncertainty is found by computing the minimal and maximal deviation from the mean in the selected sets of GIA models.

To show the effect of another ice loading model and 3-D composite mantle rheology, GIA models from van der Wal et al. [2013] are also shown in Figure 6b. The blue dashed line represents a GIA model with dry rheology with grain size of 10 mm and mantle temperatures derived from surface heat-flow data. The green dashed line represents a GIA model with a dry rheology and 4 mm grain size, but with mantle temperatures derived from a global tomography model. The ice loading history is created from a simple ice flow law with ice extent matching geologic evidence. These models provided the best match to the observed maximum uplift rate in Scandinavia. Despite using a 3-D rheology and a different ice loading model, the results are similar to model A.

The bifurcation in the comparison with RSL and uplift rate observations could be resolved by looking at the static gravity field. The gravity anomaly of model set A fits best with the predictions of Balling [1980] and Heiskanen and Vening Meinesz [1958]. The gravity anomalies of model set B are more negative, being in the range of $-25$ and

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**Figure 7.** Sketch illustrating the lithosphere in isostatic equilibrium (black lines) and the lithosphere bended as it is still relaxing from the load of a former ice sheet (red lines). The green-shaded volumes are responsible for the GIA gravity effect.

**Figure 8.** The GIA gravity signal for a flat-plate model [Watts, 2001] (green) and the spherical layered Earth model with varying lithospheric thickness, as described in the text (blue).
Table 1. The GIA Gravity Anomaly Contributions From the Different Lithosphere Models Experiencing the GIA Deflection

<table>
<thead>
<tr>
<th>Density Contrast</th>
<th>Gravity Signal [mGal]</th>
<th>Percentage</th>
</tr>
</thead>
<tbody>
<tr>
<td>GIA modeling</td>
<td>28.7</td>
<td>100</td>
</tr>
<tr>
<td>Multilayer model</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Topographic boundary</td>
<td>21.2</td>
<td>74.0</td>
</tr>
<tr>
<td>Intracrustal boundaries</td>
<td>4.6</td>
<td>16.1</td>
</tr>
<tr>
<td>Moho boundary</td>
<td>3.0</td>
<td>10.4</td>
</tr>
<tr>
<td>LAB −0.1</td>
<td>−0.1</td>
<td>−0.5</td>
</tr>
</tbody>
</table>

aThe simple model [Watts, 2001] is able to describe the gravity signal almost completely. The gravity signal of the multilayer model is created by the sum of effects of uplift of the separate density contrasts.

3. The Relation Between the GIA Gravity Signal and Lithosphere Deflection

In this section, the relation between the bending of the lithosphere due to the ice loading and the GIA gravity effect in the static gravity field is discussed. We use this relation in section 4.4 to correct for the GIA gravity signal. The gravity signal due to the postglacial deflection can be approximated with the following equation [Watts, 2001]:

$$\Delta g_{\text{GIA}} \approx 2\pi G \rho_{\text{mantle}}$$ (2)

Here, $\rho_{\text{mantle}}$ is the mantle density, $y$ is the vertical deflection which depends on the location, and $G$ is the universal gravitational constant. The thickness of the lithosphere drops out of the equations in the approach described in Watts [2001], because only mantle material is removed, but in reality a thicker lithosphere will put the source of that particular mass source deeper. The assumption that the lithosphere thickness has a negligible effect on the GIA gravity signal is tested using a simple model sketched in Figure 7. This model consists of two layers of constant thickness. The upper layer, which simulates the lithosphere, is bent with both its upper boundary ($V_1$) and its lower boundary ($V_2$) having the same deflection according to the GIA model results. The lower layer simulates the asthenosphere, where its lower boundary remains at a constant depth of 300 km. The top layer has a mean density of 2650 kg/m$^3$, and the lower layer has a mean density of 3300 kg/m$^3$. These values are representative for the Fennoscandian area [Ebbing et al., 2012]. Figure 7 shows the deflected lithosphere in its current state and in its fully relaxed form. Due to the mass deficit in the current state, a negative gravity anomaly is present, compared to the fully relaxed state. With the methodology explained in Novák and Grafarend [2006], the gravity effect can be forward modeled. The thickness, $T$, of the elastic layer in the model is varied from 0 to 200 km thickness, which covers most known lithospheric thicknesses. A deflection with a peak value of 245 m is used, resembling the maximal possible deflection of GIA model results in model B with a lithosphere thickness of 80 km in order to obtain the largest gravity signal. Figure 8 shows a small variation between the value from equation (2) at the location of maximal deflection and the modeled results with varying lithosphere thickness. A difference of around 1 mGal is seen when the lithospheric thickness...
increases to 200 km. This is well below the uncertainty in the observed gravity field in Fennoscandia. Thus, equation (2) is a good approximation for correcting the gravity field for GIA in section 4.4.

This simple two-layer model can be expanded to a multilayered Earth model that is used in the GIA modeling of section 4.2, since the total GIA gravity signal is a summation of effects due to the uplift of different density contrasts. The GIA deflection obtained from the GIA modeling was superimposed on each layer of the multilayered Earth model to calculate the contribution to the GIA gravity signal ($\Delta g_{\text{GIA}}$) for that specific layer. The Earth model that is used comprises one sedimentary layer with a density of 2500 kg/m$^3$, three crustal layers with density increasing with depth from 2700 to 2950 kg/m$^3$, and two mantle layers consisting of an elastic lithosphere with a density of 3330 kg/m$^3$, slightly higher than the viscous asthenosphere (3300 kg/m$^3$). Incompressible elastic bending is assumed in the lithosphere. The results of the forward modeling are shown in Table 1 and compared with the gravity anomaly computed by the GIA model described in section 2. The topographic contribution to the GIA gravity signal is the largest, 74.0% (21.2 mGal), because the topographic boundary has the largest density contrast ($\rho_{\text{air}} - \rho_{\text{crust}}$). The second most important density contrast is the Moho. The deflection of this density contrast introduces a contribution to the GIA signal of 10.4% (3.0 mGal). The rest of the GIA gravity signal is produced by the deflection of the internal density contrasts in the crust, added together they contribute 16.1% (4.6 mGal) to the complete GIA gravity signal. The density contrast of the sedimentary layer and upper crustal layer in this Earth model is small, so it will have a small effect in the GIA signal. However, this could be different in areas where the density contrast between sedimentary and crustal layers is significant. The gravity effect due to the lithosphere-asthenosphere boundary (LAB) is opposite in sign due to the inverse density contrast, removing 0.5% of the GIA gravity signal. The complete gravity signal can be almost completely reproduced using equation (2) with the density of the asthenosphere, because when all the different density contrasts are combined, they add up to the asthenosphere density. This also holds for models with continuous radial density distributions.

4. Results and Discussion

First we aim to answer the first research question: Can the GIA gravity signal be isolated from the observed gravity signal? Doing so would require a density model of the lithosphere, based on seismic measurements, to be subtracted from the observed gravity field. It is not possible to separate the GIA gravity signal from gravity observations using only seismic observations as the lithosphere seen by seismic measurements is still deflected due to the former ice load. In theory, the amount of bending of the lithosphere should be visible in the seismic observations, which itself would be a possible constraint on GIA models. Based on model set B, the maximum deflection could be expected to be as large as 217 m. However, in observations of surface topography it is already difficult to observe this amount of bending. The Gulf of Bothnia is lower than its surroundings, but the east-west dimension of the Gulf of Bothnia is much smaller than the predicted wavelength from GIA modeling. The bending of the lithosphere could also be visible at the Moho. However, the current accuracy of the Moho depth determination is up to 5 km [Grad et al., 2009], which causes the GIA deflection to fall within the uncertainty of the observations.

Therefore, the following section will discuss two different approaches to separate the GIA gravity signal from the observed gravity field. First, it is separated using filtering of the spherical harmonic coefficient representation of the gravity field. This idea is based on the belief that GIA can be isolated from other gravity signals based on wavelength separation. The second section (4.2) will assume isostatic compensation of crustal densities determined from seismic measurements to isolate the GIA gravity signal, which is a non-isostatic effect. This will prove to be unsuccessful due to inaccurate knowledge of crustal and upper mantle structures. Therefore, the last two sections discuss the second research question, to quantify the effects of the GIA correction in geophysical modeling. Section 4.3 will show the effect in a regional crustal model using the remaining uplift estimate, and section 4.4 shows the GIA gravity correction on gravity anomalies and its effect on interpretation of the gravity anomaly.

4.1. Separating GIA Gravity Signal Using Spherical Harmonic Coefficients Truncation

Some studies [e.g., Sjöberg et al., 1994] have separated the GIA signal from other gravity anomaly sources by filtering the spherical harmonic series. For example, Sjöberg and Bagherbandi [2013] use the geoid signal in the spherical harmonic domain of degree and order 10–70 (12 m peak value) to constrain remaining uplift in Fennoscandia to 80 m. The idea behind this band-pass filtering of the spherical harmonic coefficients is to
Figure 10. Isostatic gravity anomalies constructed by four different methods: (a) the statistical method of Balling [1980], (b) a local compensation scheme which puts the compensating masses in the upper mantle, (c) Airy compensation, and (d) Pratt compensation [Marquart, 1989].

eliminates short and long wavelength sources from the observed gravity field, which are believed to contain non-GIA sources such as mantle convection and high-resolution topography.

In Figure 9, the degree variance of the gravity field of the Earth is plotted together with a mantle convection model [Tosi, 2007] and the GIA gravity signal. The degree variance is defined in the following way [Rapp, 1982]:

\[ V_l^2 = \sum_{m=0}^{l} (C_{lm}^2 + S_{lm}^2) \]  (3)

The normalized spherical harmonic coefficients are represented by \(C_{lm}\) and \(S_{lm}\), with degree \(l\) and order \(m\). It can be seen that both the convection model as well as the GIA model have most of their energy in the first 30 spherical harmonic degrees. Furthermore, a gravity model of the top 100 km of the Earth is included in Figure 9 based on the topography and crustal masses of CRUST1.0, which are compensated by a local isostasy scheme where the density of the mantle is changed to obtain isostatic equilibrium. The spectrum of the GIA signal overlaps with that of the crustal and topographical masses for the higher degree spherical harmonic coefficients. Both results show that it is not possible to separate the GIA gravity signal simply by applying a band-pass filter in the spherical harmonic domain, since it overlaps with the signal of mantle convection and crustal masses.

This can be seen in more detail by looking at the effect of the choice of truncation limits. After inspecting the gravity signal of the GIA model results, we conclude that the spherical harmonic domain of degree and order 4–45 contains most of the Fennoscandian GIA signal. Adding coefficients with lower or higher degree and order will change the modeled GIA gravity anomaly by not more than 1 mGal. However, other studies state a spherical harmonic domain of 10 to 70 degree and order [Bjerhammar et al., 1980]. For example, the estimated −12 m geoid in Sjöberg and Bagherbandi [2013] changes as a function of lower degree boundary.
Figure 11. Isostatic gravity anomalies of Fennoscandia where the crustal structures are compensated by density differences in the mantle for different crustal models: (a) CRUST1.0, (b) CRUST2.0, and (c) EPCrust. The compensation depth is 300 km.

For the domain 5–70, the peak geoid undulation is −20 m, domain 8–70 gives −15 m, and domain 11–70 results in −8 m. This quick investigation shows that any lower degree truncation includes a signal from mantle convection or removes the GIA signal of interest. Therefore, we conclude that these low degree signals must be taken into account when studying the GIA effect in the static gravity field.

4.2. Separating GIA Gravity Signal Using Isostasy

Section 3 showed that GIA gravity signal cannot be separated using solely geometric models of the crust because GIA is superimposed in the current observed crustal geometry. Also, the GIA gravity signal cannot be separated by using only spherical harmonic band filtering. Therefore, a different approach is needed to separate the GIA gravity signal from the observations. Since GIA is a non-isostatic effect, this section studies if it is possible to separate the signal from isostatic gravity effects.

The isostatic anomalies are constructed by compensating known masses, which can be topography or crustal structures such that the total mass up to the compensation depth is equal for all locations. Both cases are discussed in the following. First, four different methods from literature are used to generate different topographic isostatic anomalies, where topographic masses are compensated in four different ways. The spread between the models gives information about the uncertainty in the isostatic anomaly introduced by the choice of compensation method. The methods are (a) a statistical technique, which assumes a linear regression between the topography and the gravity anomaly \cite{Balling1980}, (b) a local isostatic compensation scheme, where the compensating masses for the topography are modeled by density anomalies in the upper mantle defined by the CRUST1.0 Moho and a lower boundary of 100 km depth, (c) Airy compensation ($T = 30$ km and a density contrast of $650$ kg/m$^3$), and (d) Pratt compensation using the geoid undulation \cite{Marquart1989}. The computed isostatic geoid undulation of this last method is converted into a gravity anomaly for a consistent comparison. The isostatic anomalies are computed with the global gravity field model GOCO03S \cite{Mayer-Gürr2012} and topography from the ETOPO2v2 model. Figure 10 shows the isostatic anomaly constructed by the four methods. Some local differences are visible, but the overall correlation is high. The observed negative anomaly at the the center of the Gulf of Bothnia is between $-14$ mGal and $-20$ mGal, and the largest negative anomaly is located south of the Åland Islands. The standard deviation of these different solutions is between 5 and 10 mGal in the area of the Fennoscandian post-glacial uplift. The uncertainty in the choice of compensation is small enough that the GIA gravity signals from different models can be distinguished. However, this kind of compensation neglects the fact that crustal structures should also be compensated.

To compensate the crust, a similar approach is followed as the compensation of the topography, shown in Figure 10b. Every mass column is given a total mass given by a reference crust of 27 km thickness and a density of 2800 kg/m$^3$ and a mantle density of 3420 kg/m$^3$. Anomalous masses in the crust are corrected by changing the density in the upper mantle of the model. The forward modeled gravity field of this model is subtracted from the gravity observation, resulting in a crustal isostatic gravity anomaly that consists of non-isostatic effects.
Figure 12. The LCB thickness in Fennoscandia from Ebbing et al. [2012], (a) uncorrected and (b) corrected for GIA effect. The red hatched area represents the location of a negative LCB thickness due to the GIA correction.

To see the magnitude of the uncertainty introduced by unknown crustal structure, three crustal density models based on seismic data are employed here: the global CRUST2.0 [Bassin et al., 2000], the improved CRUST1.0 [Laske et al., 2013], and the regional model EPCrust [Molinari and Morelli, 2011]. The area outside EPCrust is filled in by CRUST2.0 such that it can be used as a global model in the forward modeling scheme. The density models are forward modeled up to degree and order 45 in spherical harmonic coefficients as beyond degree 45 there is little GIA signal (section 4.1). The resulting isostatic gravity anomalies are shown in Figure 11. In all three models, a different anomaly is seen in the region of expected maximum GIA uplift. The CRUST2.0 result shows a large low gravity anomaly where the Baltic craton is expected, whereas EPCrust shows some correlation with the Moho geometry from Grad et al. [2009], which is used in the construction of EPCrust. CRUST1.0 shows a long wavelength east-west gradient in the anomaly. The variation of the anomaly in the area of maximum uplift ranges from −50 to 60 mGal in between the three crustal models. A spread of 110 mGal is well above the uncertainty in the GIA models, and therefore, it will not be possible to select between competing GIA models such as model set A or model set B. This illustrates that the uncertainty in density models of the crust and associated compensation in the upper mantle is still too large to extract the GIA gravity signal from the gravity observations even without consideration of the mantle convection signal.

4.3. Correcting Crustal Models for the Deflection Due to GIA

This section will investigate the second research question: What effect does the correction for GIA have on crustal density interpretations? An earlier study already showed the significance of correcting for GIA [McKenzie, 2010]. To quantify this effect, the study of Ebbing et al. [2012] is chosen as example. In this model, the assumption of equal mass columns is made to constrain the thickness of a high density lower crustal body (LCB) located at the bottom of the crust. If the GIA signal is not taken into account, the crustal masses are overcompensated in the areas where ongoing postglacial rebound is present, as explained in the previous part. This overcompensation results in a too large thickness of the LCB. The thickness of the LCB is obtained by setting the mass of every column equal to that of a reference mass column, as stated in Ebbing et al. [2012]:

\[
\rho_{\text{topo}} D_{\text{topo}} + \sum_{i=1}^{3} \rho_{c_i} D_{c_i} + \rho_{\text{LCB}} D_{\text{LCB}} + \rho_{\text{mantle}} D_{\text{mantle}} + \rho_{\text{asth}} D_{\text{asth}} = \sum_{i=1}^{5} \rho_{\text{ref}_i} D_{\text{ref}_i}
\]  

(4)

Here, \(D\) is the thickness of the layer, given in the subscript, and \(\rho\) is the density of that layer. The density model consists of a topographic layer, three crustal layers plus the LCB, the upper mantle lithosphere, and the asthenosphere. The reference model used by Ebbing et al. [2012] consists of five layers. He deduced the density of those layers from crustal observations in the coastal areas (\(h_{\text{topo}} \approx 0\) around Fennoscandia.

Correcting for the remaining GIA uplift can be done by introducing the GIA deflection to all the crustal layers in the model of Ebbing et al. [2012]. As a result of this correction, extra asthenosphere mass per column is
introduced. The amount of extra mass is location dependent and described by the shape of the GIA deflection. Equation (4) has to be changed to account for additional mass that simulates the non-isostatic effect of GIA. This is done by introducing an extra layer of mass ($\rho_{\text{asth}} D_{\text{GIA}}$) in the density model. The relation becomes

$$\rho_{\text{topo}} D_{\text{topo}} + \sum_{i=1}^{3} \rho_{c_3} D_{c_3}^{\text{new}} + \rho_{\text{LCB}} D_{\text{LCB}}^{\text{new}} + \rho_{\text{mantine}} D_{\text{mantine}} + \rho_{\text{asth}} D_{\text{asth}} + \rho_{\text{asth}} D_{\text{GIA}} = \sum_{i=1}^{5} \rho_{\text{ref}_i} D_{\text{ref}_i}$$

(5)

To find the new thickness of the LCB layer, equation (4) is subtracted from (5), resulting in

$$\left(\rho_{c_3} D_{c_3}^{\text{new}} + \rho_{\text{LCB}} D_{\text{LCB}}^{\text{new}}\right) - \left(\rho_{c_3} D_{c_3}^{\text{old}} + \rho_{\text{LCB}} D_{\text{LCB}}^{\text{old}}\right) + \rho_{\text{asth}} D_{\text{GIA}} = 0$$

(6)

All layers keep the same thickness except the LCB and the lower crust. The sum of thicknesses of the third crustal layer, and the LCB is chosen equal for both the old model and the new model. This means only the upper bound of the LCB is changed. The relation for the new LCB thickness can be found:

$$D_{\text{LCB}}^{\text{new}} = D_{\text{LCB}}^{\text{old}} - \frac{\rho_{\text{asth}}}{\rho_{\text{LCB}} - \rho_{c_3}} D_{\text{GIA}}$$

(7)

Thus, the geometry of the new LCB depends on the amount of GIA displacement remaining and the densities of the three layers. The new LCB thickness depends largely on the difference in density of the LCB and lower crustal layer. If these densities are close, the LCB becomes even thinner.

For the correction, we use the GIA model with maximum deflection to show the largest effect. This model with viscosity structure of $v_{um} = 16 \times 10^{20} \text{ Pa s}$ and $v_{lm} = 256 \times 10^{20} \text{ Pa s}$ has a maximum deflection of 217 m. Correcting the density model with this GIA model according to equation (7) results in a reduction of the thickness of the LCB of 3.2 km using the density values of Ebbing et al. [2012]. This is 6.4% of the total thickness of the LCB layer, which is small but can be considered to be within the resolution of seismic measurements. Forward modeling the new density model with the corrected LCB produces a decreased gravity anomaly due to the smaller LCB layer compared with the initial model, by a value of 30 mGal. This roughly agrees with the GIA gravity signal caused by the subsided lithosphere.

Figure 12 shows the differences between the uncorrected and corrected LCB thickness in the Fennoscandian area. The shape is similar, but there are some subtle changes. Around the Gulf of Bothnia the changes are largest, corresponding to the location of maximum uplift. The large LCB thickness in west and mid Finland is still present, but the GIA correction has reduced the variation of the LCB thickness in other regions. Overall, the GIA correction does not alter the conclusions made by Ebbing et al. [2012]. Still, we suggest to apply a correction with the best fitting GIA model, because the uncertainty in the GIA model is small enough that crustal density estimates can be improved.

4.4. Correcting Observed Gravity Anomalies for the Gravity Anomaly Due to GIA

In the previous section, the effect of correcting for the remaining uplift was studied. Gravity anomalies in Scandinavia can also be corrected for GIA, in which case the corrected gravity anomaly represents the future gravity field, when the lithosphere has fully relaxed from the melting of the ice sheet. Studies that use gravity data, together with the assumption of isostatic equilibrium to constrain density structures, mistakenly compensate the deflected lithosphere due to GIA by increasing mass below the lithosphere. Section 3 showed the GIA gravity signal induced by the deflection of the lithosphere. The observed free-air gravity anomaly can be corrected for this GIA gravity signal in the following way:

$$g_{\text{corrected,FA}} = g_{\text{FA}} - \Delta g_{\text{GIA}}$$

(8)

The GIA gravity signal, $\Delta g_{\text{GIA}}$, can be calculated from GIA modeling or by using the GIA deflection according to equation (2); $g_{\text{FA}}$ is the observed free-air anomaly from GOCC03S [Mayer-Gürr et al., 2012] seen in Figure 1. Figure 13 shows the free-air anomaly after correcting for the GIA effect. The large negative (blue) anomaly disappears after the GIA correction, which indicates that it is mainly caused by postglacial rebound. The positive gravity anomalies in the South Scandes are slightly enhanced, which reduces the need for a thick crustal root to support the mountains in order to match the observed, GIA corrected, gravity anomaly. It should be stressed that this correction only makes sense when the assumption of isostatic equilibrium is made in the modeling approach.
The Bouguer anomaly can also be corrected for the GIA gravity effect; however, the topography that should be used is that of the fully relaxed topography. The GIA corrected Bouguer anomaly becomes

$$\Delta g_{\text{corrected},B} = \Delta g_{\text{corrected,FA}} - 2\pi G \rho_{\text{ch}} h_{\text{corrected}} \quad (9)$$

with $h_{\text{corrected}} = h_{\text{obs}} - h_{\text{GIA deflection}}$, where the GIA deflection is obtained from GIA modeling. The corrected Bouguer anomaly is close to the uncorrected Bouguer anomaly, because GIA is removed in both the gravity signal and the topography. Most of the GIA signal is situated in the displacement of topography, as discussed in section 3.

5. Conclusions

The goal of this study was to answer two questions. First, can the GIA gravity effect be separated from the observed static gravity field? This question is answered by applying different techniques from the literature to isolate the GIA signal from the static gravity field.

The GIA gravity signal is generated by the deflection of the isostatic lithosphere, pushing away mantle material. From the layered lithosphere modeling, it is found that 74% of the GIA gravity signal is generated by the deflection of the topography. The remainder of the signal is due to the uplift of the remaining density contrasts in the crust.

Separating the GIA gravity effect by considering a particular bandwidth in the spherical harmonic domain is unsuccessful, due to the overlap in bandwidth of dynamic deep mantle anomalies, GIA, and signals from the crust. Furthermore, separating the GIA gravity signal using models for the crust in combination with isostatic compensation in the mantle was investigated. Several different methods to achieve isostatic compensation have been pursued, and it is shown that the uncertainty due to the compensation techniques is small (within 5–10 mGal). However, the uncertainty in crustal models based on seismic observations is still too large to separate GIA from density anomalies in the crust and upper mantle. It is concluded that contrary to earlier findings, it is not possible to extract the GIA signal from the observed gravity field with sufficient accuracy to be useful as constraint in GIA modeling.

Current GIA models have less uncertainty than the global and regional density models, such as CRUST2.0, CRUST1.0, and EPCrust, for gravity field modeling purposes. Therefore, it is suggested to correct gravity observations and models using the results from GIA modeling. Such a correction is necessary if the modeling assumes that all topography and crustal masses are isostatically compensated, as the deflections introduced by GIA are non-isostatic. Which brings us to the second research question: what is the effect of the GIA correction in geophysical models constrained by gravity?

A simple correction is done by superimposing the lithosphere deflection, obtained from GIA modeling, onto topography and density layers. The corrected density models have effects in geophysical modeling of the crust and lithosphere (up to 5–10% in thickness of lower crustal layer in the example study). Gravity observations can be corrected by using the simple relations in equations (2) and (8). It will be interesting to see if GIA models in other regions, such as North America and polar regions, are accurate enough to apply a GIA correction there.

Appendix A: Extra Glacial Period

As shown by Johnston and Lambeck [1999], Kaufmann et al. [2002], and Kaufmann and Wu [2002], adding another ice loading cycle affects estimates using RSL curves and other observables. Therefore, the used data sets, regional RSL curves, GPS uplift rates, and GRACE gravity change, are compared with GIA models using two ice cycles to see if it affects the conclusions in this study. The extra ice loading cycle before ICE-5G is
modeled with a linear growth rate beginning at 227 ka, obtaining a maximum at 135 ka, in height identical to the height at the glacial maximum modeled by ICE-5G. After the penultimate glacial maximum, melting occurs until zero ice mass is reached at the beginning of the ICE-5G ice model (122 ka). No interglacial period is chosen, because literature is still relatively unclear about the length of this period [Svendsen et al., 2004; Penaud et al., 2008]. Furthermore, neglecting an interglacial period will have the most effect on the state of the GIA models, because less relaxation is obtained prior to the last glacial cycle. The ice loading history is used in the GIA modeling using similar settings as in section 2.

The resulting misfit plots are shown in Figure A1 for a lithosphere thickness of 150 km. The comparisons of the models with other lithosphere thicknesses give similar deviations and result in similar selection of viscosity profiles. Overall, the introduction of the extra ice loading cycle does not significantly change the RSL, GPS uplift, and GRACE gravity change, compared to the results in Figures 3–5. When inspecting Figure A1 in more detail, it can be noticed that there are small changes in the misfit plots. These changes, however, do not change the conclusions for selecting the GIA models, A and B.

The effect of the extra ice cycle is also inspected for the modeled remaining uplift. Figure A2 shows the similar cross-section as in Figure 6 of model set A and B. The black (model set A) and red (model set B) lines are the results obtained using only ICE-5G. The results using an extra ice loading cycle are included in blue for model set A and green for model set B. For model A, an extra deflection of 6 m is observed at maximum deflection, which is around 5% of the total value. Model B shows a bit more variation, especially the GIA model with upper mantle and lower mantle viscosities of $32 \times 10^{20}$ Pa s and $512 \times 10^{20}$ Pa s, respectively. The larger viscosity shows more sensitivity to earlier ice loading changes, whereas low viscosity does not “remember” these loads. With the exception of the mentioned model, the differences are small and similar to model set A. The mentioned GIA model has a difference of 18 m, which is 7.8% of its total maximum deflection. Inspecting the gravity field differences give similar results, within 10% of the deflection value. It can be concluded that previous ice loading cycles have a small effect on the Fennoscandian GIA modeling done in this study. Nevertheless, for more accurate GIA modeling, extra ice loading histories should be modeled.
Figure B1. The 15 RSL curves from Tushingham and Peltier [1992] with two best-fit results from GIA modelling, Model A (red) and Model B (blue).

Appendix B: Data Used

This section presents the RSL data used in the GIA modelling. The RSL curves are a selection from the data set of Tushingham and Peltier [1992]. Only sites in Fennoscandia are used and for which a long time series is available. This results in 15 sites with the locations shown in Figure 2a. The data with its spatial and time error estimates are plotted in Figure B1 together with RSL predictions from the two best fitting GIA models from the discussed model set A and B.

References


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