Determination of the Earth’s structure in Fennoscandia from GRACE and implications for the optimal post-processing of GRACE data

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SUMMARY
Analysis of data from the Gravity Recovery and Climate Experiment (GRACE) satellite mission allows us to identify regions of long-term mass changes such as the areas of Glacial Isostatic Adjustment (GIA) in North America and Fennoscandia. As there are now more than 7 yr of data available, the determined trends are robust enough for the inference of viscosity structure of the Earth’s mantle. In this study, we focus on the Fennoscandian rebound area as there are abundant high-quality terrestrial data to use as ground-truth.

In the first step, GRACE data are taken to fix the optimal radial (1-D) viscosity profile and the lithospheric thickness combination, which are needed as background parameters in 3-D earth modelling. The results are in basic agreement with results based upon relative sea level and GPS data, showing a lithospheric thickness in Fennoscandia between 90 and 160 km and an upper mantle viscosity of about $2 \times 10^{20}$ Pa s. The lower mantle viscosity is poorly resolved, however. In the second step, GRACE data are used to constrain the 3-D viscosity using spherical finite element modelling. In this case, the results also agree with past investigations, but GRACE data alone cannot discriminate between lateral heterogeneities in the mantle that are thermal in origin from those due to changes in chemical composition. More notably, we treat in detail GRACE-related questions such as implementation of an adequate Level-2 filter technique and identification of the best reduction method for hydrological mass change signals. It turns out that the Gaussian filter technique is the best for this type of investigation. Even the best global hydrology models used in GRACE investigations still fail to improve the mismatches—thus one should be careful not to blindly use them for ‘improving’ GIA models in North America or other centres of rebound.

In conclusion, our study clearly shows that GRACE data greatly complement the study of GIA. As there are new GRACE releases in progress, and also in light of a new generation of ice history models, GRACE is likely to sharpen our insights concerning earth structure and rheology within the next few years.

Key words: Numerical solutions; Satellite geodesy; Gravity anomalies and Earth structure; Composition of the mantle; Dynamics of lithosphere and mantle; Europe.

1 INTRODUCTION
Glacial isostatic adjustment (GIA) is a well-known phenomenon in North America and Fennoscandia. This ongoing process, induced by the rebound of the Earth’s surface after the former Late Weichselian glaciation, is observed by geological (relative sea level (RSL)) and geodetic data (tilt, GPS, absolute and relative gravity measurements, tide-gauges, free-air gravity, low-order satellite gravity change). These data have been successfully used to determine the Earth’s rheological structure beneath these regions.

Recently, new data have become available from satellite missions such as GRACE (Gravity Recovery and Climate Experiment). The focus of this paper is on using the GRACE monthly solutions to infer mantle rheology and ice history in Fennoscandia as there are plenty of high-quality terrestrial data to use as ground-truth. As different processing centres give different GRACE solutions and also different processing techniques can alter the solution, it is not clear how this would affect the interpretation. This paper will show that despite of the differences in GRACE solutions, a best viscosity profile and ice history model can be inferred, and that it is possible to evaluate the different processing techniques. Finally, a combined GRACE solution is also proposed.
There have been similar studies with GRACE data (see Ivins & Wolf 2008, for an overview). In an earlier study, Velicogna & Wahr (2002) compared simulated GRACE measurements and earth models. They found that by using GRACE data alone, it should be possible to obtain useful information about the Earth’s viscosity structure. They suggested to use GRACE measurements, in combination with other GIA data, in inversions to improve the Earth’s viscosity profile. Van der Wal et al. (2008) investigated the uncertainties introduced by various post-processing methods of GRACE data on the secular gravity rates in North America. The authors showed that the largest uncertainty is due to hydrology which can significantly affect the estimation of secular gravity rate and thus the interpretation of ice thickness or mantle viscosity in GIA studies. Both van der Wal et al. (2008) and Tamisiea et al. (2007) found that the estimated secular gravity rate in North America showed two distinct peaks west and south east of Hudson Bay. However, van der Wal et al. (2008) found the maximum south east of the Hudson Bay, while Tamisiea et al. (2007) located the maximum in the west. This discrepancy is related to several differences in the analysis, such as different GRACE solutions, time spans and further data reductions including that with hydrological models. Nevertheless, both results confirm geological evidence that the ancient Laurentian ice complex was composed of two large domes (Dyke et al. 2003).

Tamisiea et al. (2007) also determined an earth model that produces a so-called preferred GIA prediction that can be removed from the GRACE trend. The lithospheric thickness is 120 km, upper-mantle viscosity $\eta_{LM} = 8 \times 10^{20}$ Pa s and lower-mantle viscosity $\eta_{UM} = 3 \times 10^{21}$ Pa s.

In a general study, Paulson et al. (2007) used 53 months (between April 2002 and December 2006) of GRACE data with post-processing described in Swenson & Wahr (2006) to infer the mantle viscosity structure. They combined GRACE-derived secular gravity changes near Hudson Bay, and geological measurements of RSL changes over the last 10,000 a in the same region in a Monte Carlo inversion, as suggested by Velicogna & Wahr (2002). Assuming a lithospheric thickness of 120 km, they found that the viscosities in the upper and lower mantle have values of $5.3 \times 10^{20}$ and $2.3 \times 10^{21}$ Pa s, respectively. Other models that provide a reasonable fit to the data may allow a weaker upper mantle compensated by a stronger lower mantle. They also found that the GRACE and RSL data used in their study cannot resolve more than two layers in the upper 1800 km of the mantle.

Both Tamisiea et al. (2007) and Paulson et al. (2007) applied GLDAS global hydrology model (Rodell et al. 2004) to reduce the disturbing signals from hydrology. In fact, the various Level-2 GRACE monthly solutions contain mass signals from the hydrosphere, cryosphere and geosphere, in addition to unremoved effects and/or artefacts from pre-processing. Thus, depending on the region and the temporal signature of the effect of interest, different dedicated post-processing techniques are required to clearly determine the signal of interest (Steffen et al. 2009c). However, Barletta & Bordoni (2009) did not use any hydrology model in their extraction of GIA signal with the weighted mass trends technique.

A comparison of several studies (Steffen et al. 2008, 2009a,b,c) related to the secular trend in Fennoscandia shows that after 7 yr of GRACE in orbit, the trend is both stable and consistent with patterns generated by terrestrial sources, such as GPS (Lidberg et al. 2007) and AG measurements (Gitlein et al. 2008). The GIA pattern in Fennoscandia can be loosely described as an ellipse with maximum amplitude in the northern Gulf of Bothnia (Fig. 1). As the GRACE-derived gravity change is in general agreement with this configuration, the aim of this paper is to use the robust trend to infer the lithospheric thickness and mantle viscosity in Fennoscandia for 1-D and 3-D earth models. So far, such an investigation has been done by Paulson et al. (2007) for North America, but only in combination with RSL data for 1-D earth models. For Fennoscandia, this investigation with GRACE data is new, particularly with respect to 3-D earth models. In addition to the determination of radial earth

Figure 1. Results of terrestrial measurements presenting the GIA pattern in Fennoscandia: (a) Fig. 3 of Steffen et al. (2009b) showing the uplift contour lines (in $\mu$Gal yr$^{-1}$) after Ekman (1996) converted from geometrical height changes to gravity changes by applying the factor $-2.04 \mu$Gal cm$^{-1}$, which is in accordance with Ekman & Mäkinen (1996). Red dots mark distribution of absolute gravimetry stations observed within a multi-national cooperation during annual campaigns since 2003 by FG5 absolute gravimeters. (b) GPS-derived uplift velocity field after Lidberg et al. (2007). Black dots indicate GPS stations.
parameters, we analyse the capability of GRACE to determine the 3-D structure and the chemical composition of the mantle. The effect of lateral viscosity variation can be significant especially for horizontal motion (Kaufmann et al. 2005). There exist several investigations to North America (Wu 2005, 2006; Wang & Wu 2006a,b; Wang et al. 2008; Wu & Wang 2008), but only Wang et al. (2008) and Wu & Wang (2008) used GRACE to constrain mantle rheology. In turn, with both the 1-D and 3-D modelling we are able to qualitatively state which GRACE solution fits best and which analysing approach, including filtering techniques and discussion of global hydrology models, probably leads to the best results. Therefore, we use different GRACE solutions, apply selected filtering techniques and investigate contributions of two global hydrology models.

Several ice models are available (e.g. ICE-3G, ICE-4G, ICE-5G, RSES) for the 1-D and 3-D modelling. Previous studies already showed that the predictions from these ice models fit the RSL data reasonably well (e.g. Lambeck et al. 1990, 1998; Peltier 2004; Steffen & Kaufmann 2005). Here we will also use the GRACE data to evaluate the relative merits of Fennoscandian ice models.

In the next two sections we discuss our post-processing of GRACE Level-2 data and briefly introduce our earth and ice model parameters, respectively. This is followed by the presentation and discussion of the results and finally, we summarize our main findings.

2 GRACE DATA AND ANALYSIS

GRACE monthly, weekly and 10-days solutions are provided by several analysis centres, such as the three main analysis centres Center for Space Research (CSR) in University of Texas at Austin, Jet Propulsion Laboratory (JPL) in Pasadena, and the Helmholtz-Zentrum Potsdam, Deutsches GeoForschungsZentrum (GFZ). In addition, there are solutions from the University of Bonn (ITG), the Centre National d’Etudes Spatiales (CNES) in Toulouse and the Technical University Delft (DEOS Mass Transport model release 1, DMT-1). In this study, we determine secular gravity changes in Fennoscandia from the GRACE monthly solutions Release 4 (RL04) provided by CSR and GFZ, owing to the fact that only these two provide monthly solutions with calibrated standard deviations. Furthermore, the DMT-1 data are compared as they provide the solution from another filter technique.

We use 69 monthly gravity field solutions from July 2003 to March 2009, with a gap in January 2004 for GFZ. CSR and TU Delft have provided all solutions of that time span. Each GRACE monthly solution consists of a set of spherical harmonic coefficients \( C_{lm} \) and \( S_{lm} \) up to degree and order 60 (CSR) or 120 (GFZ, DMT-1) with corresponding calibrated errors for CSR and GFZ (GRACE 2009). Each solution centre reduces oceanic and atmospheric contributions as well as tidal effects in a standardized centre-specific processing procedure.

Gravity values \( dg(\varphi, \lambda, t) \) are computed on a \( 1^\circ \times 1^\circ \) grid for each monthly solution. We simultaneously fit a constant, a linear trend, an annual and a 2.5-yr periodicity to the gravity values. The latter is an average of basin-related and filter-dependent frequency analysis defined by Schmidt et al. (2008). Ray et al. (2003) showed that aliasing exists for the S2, K2 and K1 tides, which result in 161 d, 3.7 yr and 7.4 yr periods, respectively. Although the time period covered by the GRACE monthly fields in our study is less than 7.4 yr and thus the contributions from K2 and K1 are poorly retrieved due to their long periods, aliasing from the S2 tide is considered. Hence, the 161-day period is included to reduce effects that may result from an insufficient ocean tide correction, particularly in high latitude areas (Ray et al. 2003). If a Gaussian filter with 400 km radius is used, the accuracy for the secular trend, which results from the regression according to the Gauss–Markov theorem, is about 0.1 \( \mu \)Gal yr\(^{-1} \).

The gravity field variations of the Earth result from the integral effect of oceanic, atmospheric and hydrological mass movements and those caused by the dynamics of the Earth’s interior. In addition, the monthly fields may also contain residual signals that are artefacts of the pre-processing. Thus, adequate processing with appropriate filtering techniques has to be applied to extract the GIA signal. On another note, it is well known that the gravity fields require smoothing to reduce the effects of errors present in short-wavelength components. As the smoothing radius decreases, these errors manifest themselves in maps of surface mass variability as stripes generally oriented north to south (Swenson & Wahr 2006). As a consequence, the spherical coefficients are truncated here at degree and order 50 and then treated by different filtering techniques.

Several filtering techniques, mainly non-isotropic, have been published in the past years (e.g. Han et al. 2005; Chambers 2006; Sasgen et al. 2006; Swenson & Wahr 2006; Kusche 2007; Wouters & Schrama 2007; Davis et al. 2008; Klees et al. 2008; Duan et al. 2009; Kusche et al. 2009; Wu et al. 2009). They have been designed to reduce the stripes, but accepting the risk of removing real signals (Swenson & Wahr 2006). In most applications, the isotropic Gaussian filter is used (Jekeli 1981; Wahr et al. 1998) for the GRACE monthly gravity fields as well as the non-isotropic destriping filter from Swenson & Wahr (2006) or Chambers (2006). Although the Gaussian filter depends on the spherical harmonic degree \( l \) only, the destriping filter presents a non-isotropic filter for decorrelation of GRACE coefficients. The spectral signature of the correlated errors is examined and removed using polynomials, which clearly reduces the presence of stripes in the GRACE gravity fields. After using these filters, stripe artefacts may still be present between \( \pm 45^\circ \) latitude for Gauss-only and between \( \pm 15^\circ \) latitude for Gauss + destriping filter.

Kusche (2007) and Kusche et al. (2009) have presented a more efficient procedure to reduce stripes and spurious patterns, while retaining the signal magnitudes. An approximate decorrelation transformation, similar to a Tikhonov-type regularization of the original normal equation system, is applied to the monthly solutions. This smoothing reduces the noise in the higher frequencies. It takes into account the GRACE orbit/sampling geometry and a priori information of the expected hydrological and ocean signal variability from models. Due to the variation of parameters, three different smoothing degrees named DDK1, DDK2 and DDK3 have been developed. They correspond to Gaussian filter radii of 530, 340 and 240 km, respectively. Monthly models for GFZ, CSR, JPL and CNES as well as 10-days solutions for CNES filtered by this method can be downloaded from the ICGEM website (http://igcem.gfz-potsdam.de/ICGEM/ICGEM.html).

Klees et al. (2008) presented an anisotropic, non-symmetric (ANS) filter technique, which is claimed by the authors to be the optimal filter for the GRACE monthly solutions. The filter incorporates the noise and the full signal variance-covariance matrix to tailor the filter to the error characteristics of a particular monthly solution. It can be treated as extended version of the technique presented in Kusche (2007). The ANS filter effectively acts as a destriping filter with additional 400 km Gaussian filter (Liu et al. 2010) and is used for the DMT-1 monthly solutions.
In this study, the following filters will be used: the Gaussian filter (Wahr et al. 1998), the destriping filter (Swenson & Wahr 2006), the DDKx filter (Kusche 2007; Kusche et al. 2009) and the ANS filter (Klees et al. 2008). Furthermore, several filter radii between 350 and 800 km for the Gaussian and destriping filter will be tested. Fig. 2 gives selected examples of the calculated trend from different GRACE solutions and using different filter techniques.

Inspection of this figure shows that smaller filter radii, 350 or 400 km, reveal more detailed features (compare Figs 2c and d with Figs 2a and b). Furthermore, the maximum value around the Gulf of Bothnia increases (compare Fig. 2a, b and c), giving the predicted uplift value of about 1.2–1.3 $\mu$Gal yr$^{-1}$ (Steffen et al. 2008).

As an example, we show the output of the DDK2 filter for GFZ (Fig. 2g). It agrees well with Fig. 2(d), because DDK2 is comparable to a Gaussian filter of 340 km. CSR with both destriping (DS) filter from Swenson & Wahr (2006) and 400 km Gaussian filter. (f) combination of both centres with Gaussian filter of 400 km. (i) DMT-1 solution with ANS filter (Klees et al. 2008), which is designed to perform like a destriping filter with additional 400 km Gaussian filter. Units in $\mu$Gal yr$^{-1}$.

Figure 2. Selected examples of trend estimates from GRACE monthly solutions in Fennoscandia using different filter techniques and solution centres. GFZ with Gaussian (G) filter of (a) 800 km, (b) 600 km, (c) 400 km, (d) 350 km and (g) filter DDK2 (Kusche 2007; Kusche et al. 2009), which is comparable to a Gaussian filter of 340 km. CSR with (e) Gaussian filter of 400 km and (h) both destriping (DS) filter from Swenson & Wahr (2006) and 400 km Gaussian filter. (f) combination of both centres with Gaussian filter of 400 km. (i) DMT-1 solution with ANS filter (Klees et al. 2008), which is designed to perform like a destriping filter with additional 400 km Gaussian filter. Units in $\mu$Gal yr$^{-1}$.
to a Gaussian filter with 340 km. Small differences occur south of the uplift region, where stripe features are removed. However, the axis of the filtered uplift region is rotated in a SW–NE direction, although the maximum amplitude is unaffected.

Using the destriping filter (Fig. 2f) for CSR, the uplift maximum is slightly smaller but notably shifted to Central Finland. This shift, according to current knowledge, is improbable (cf. Scherneck et al. 2003; Lidberg et al. 2007; Steffen et al. 2008). This shift occurs for all solutions (CSR, GFZ, JPL, CNES, ITG), and the cause is clearly the destripping filter which alters or reduces the NS trend around the Gulf of Bothnia. Thus, for further investigation, we select the Gaussian filter with 400 km filter radius as basis, but will also compare the other filter radii, the DDK Gaussian filter with 400 km filter radius as basis, but will also compare the other filter radii, the DDK filter and the destripping filter to the GIA models. Applying 400 km radius, the maximum area is isolated at the GIA-predicted location and the stripe artefacts are limited in their contribution.

Fig. 2(i) confirms that the ANS-filter acts as a destripping filter with an additional 400 km Gaussian filter. The DMT-1 solution fits in shape to the CSR solution with destripping filter, but the uplift maximum is larger and centred around the Gulf of Bothnia as in Fig. 2(c).

Comparing GFZ and CSR (Figs 2c and e), differences in the magnitude and shape are clearly visible. Although the discrepancies in the secular trend in Fennoscandia among different analysis centres was discussed in several papers by Steffen et al. (2008, 2009a,b,c), full clarity concerning the origin is still elusive. It is important to explore those differences in the post-processing of each analysis centre which are, however, not documented in the handbooks. Nevertheless, both GFZ and CSR fields indicate a trend in geoid that is centred around the Gulf of Bothnia. Thus, we are motivated to explore what information exists for constraining Earth structure. In doing so, we introduce a simple averaging (arithmetic–geometric mean) of both solutions (Fig. 2f). The trend estimates for Fennoscandia of CSR and GFZ are weighted by 50 per cent. Fig. 2(f) clearly shows the averaging. The maximum of GFZ is decreased and the shape is more SW–NE directed, muting the larger amplitudes around the Lofoten northwest of the Norwegian coast apparent in GFZ fields.

Temporal gravity variations can also be related to hydrological signals. Steffen et al. (2009c) summarized and discussed three global hydrology models and their long-term trend in Northern Europe, they are: the Global Land Data Assimilation System (GLDAS, Rodell et al. 2004), the Land Dynamics World (LaDWorld, Milly et al. 2002) and the WaterGap Global Hydrology Model (WGHM, Döll et al. 2003). A comparison among them clearly shows discrepancies between these hydrological models. Although WGHM highlights a positive trend of about 16 mmETWM yr\(^{-1}\) (mm in equivalent thickness of water mass per year) in Central Scandinavia, LaDWorld and GLDAS yield only small long-term trends of less than 10 mmETWM yr\(^{-1}\). (Note that the GIA signal in Fennoscandia is about 35 mmETWM yr\(^{-1}\)). In contrast, all three models show a negative trend in the East European Plains.

In view of the GRACE gravity trend due to GIA, the contribution from all hydrology models is smaller by at least 20 mmETWM yr\(^{-1}\) than the detected GRACE trend signal in Fennoscandia. The hydrological effects derived from LaDWorld and GLDAS in the region of interest are nearly negligible. The larger hydrological effects in the WGHM do not strongly affect the GRACE-derived maximum signal, but slightly alter the shape of the g-dot pattern and may interfere with extraction of earth modelling information. On another note, the strong differences among the long-term hydrological variations from all models requires reconciliation before being used to correct for the long-term trend in GRACE data (Steffen et al. 2009c).

In an attempt to ameliorate corruption from hydrology, we remove the long-term hydrological contribution in Fennoscandia calculated with WGHM and compare that with removal from GLDAS for the selected time span in the GRACE result (Section 5.3).

### 3 GIA MODELLING

We compare the GRACE solutions to model predictions of spherical, compressible, Maxwell-viscoelastic earth models. In Section 5.1, we use a 1-D earth model with the software package ICEAGE (Kauffmann 2004). In Section 5.2, the Coupled Laplace-Finite Element method (Wu 2004) is employed for the 3-D earth models.

#### 3.1 1-D earth model

The calculation method for this approach has been extensively described in Steffen & Kauffmann (2005). In general, this is an iterative procedure in the spectral domain up to degree 192 following the pseudo-spectral approach outlined in Mitrovica et al. (1994) and Mitrovica & Milne (1998). It allows us to derive several quantities of interest, such as relative sea level change, present-day surface motions, time-dependent perturbations of the gravitational field and rotational contributions from the ice-ocean imbalance, using spherically symmetric (1-D), compressible, Maxwell-viscoelastic Earth models. The elastic structure of the models is derived from PREM (Dziewonski & Anderson 1981), and lithospheric thickness is a free parameter that can vary from 60 to 160 km. Mantle viscosity is parametrized in several sub-lithospheric layers with constant viscosity within each layer. The upper mantle viscosity \(\eta_{LM} \) is allowed to vary between \(10^{19}\) and \(4 \times 10^{21}\) Pa s, and the lower mantle viscosity \(\eta_{LM} \) between \(10^{21}\) and \(10^{23}\) Pa s (Table 1). The Earth’s core is assumed to be inviscid, and is incorporated as an inner boundary condition. With this GIA model, we search for the best-fitting 3-layer (lithosphere, upper mantle, lower mantle) earth models in combination with different ice models (Section 3.3) to see which model combination gives the best fit to the GRACE estimated secular rate of gravity change in Fennoscandia. The search involves almost 1100 different earth models.

#### Table 1. Earth models used for the sensitivity analysis.

<table>
<thead>
<tr>
<th>Earth model</th>
<th>LT (km)</th>
<th>LM1 (Pa s)</th>
<th>LM2 (Pa s)</th>
<th>Ice model</th>
</tr>
</thead>
<tbody>
<tr>
<td>1D</td>
<td>60–160</td>
<td>10^{19} to 4 × 10^{21}</td>
<td>10^{21} to 10^{23}</td>
<td>RSES, ICE-3G, ICE-4G, ICE-5G</td>
</tr>
<tr>
<td>3D-RF3S20</td>
<td>115</td>
<td>6 × 10^{20}</td>
<td>3 × 10^{21}</td>
<td>6 × 10^{21}</td>
</tr>
</tbody>
</table>

**Note:** LT, lithospheric thickness; UM, upper mantle viscosity; TZ, transition zone viscosity; LM1, shallow lower mantle viscosity; LM2, deep lower mantle viscosity.
3.2 3-D earth model

The 3-D earth models are spherical, non-rotating, self-gravitating, viscoelastic models with Maxwell rheology, material compressibility and self-gravitating oceans. In this study, the model RF3S20 is used as reference (Wang et al. 2008). It is mainly characterized by a 115-km-thick elastic lithosphere, four mantle layers (upper mantle, transition zone, shallow lower mantle and deep lower mantle, denoted by UM, TZ, LM1, LM2, respectively) and an inviscid core. The assigned background viscosity value for UM and TZ is $6 \times 10^{20}$ Pa s (Table 1). In the lower mantle, the background viscosities for LM1 and LM2 are set to $3 \times 10^{21}$ and $6 \times 10^{21}$ Pa s, respectively. The lateral viscosity perturbation is inferred from the lateral SH velocity anomalies given in seismic tomographic model S20A of Ekström & Dziewonski (1998) by a modified version of Ivins & Sammis (1995) scaling relationship, where $\beta$ is a scaling factor that has been included in this relationship to determine the contribution of thermal effects versus compositional heterogeneity and non-isotropic pre-stress effects on lateral heterogeneity in mantle viscosity. When $\beta = 1$, lateral velocity variations are caused by thermal effects alone. For $\beta = 0$, there is no lateral viscosity variation and the Earth is laterally homogeneous. The value of $\beta$ around 0.2–0.4 is found to explain most of the global relative sea level data, the uplift rates in North America and Fennoscandia and the BIFROST horizontal velocity data (Wang et al. 2008). We assign four different values of $\beta$, [0, 0.2, 0.4, 0.6], which results in four different 3-D earth models. Former tests have shown that $\beta$ factors larger than 0.75 result in the worst misfits, and thus we limit the variation to these four values.

3.3 Ice models

The surface load consists of two contributions, one from the Late Pleistocene ice sheet and the other from the corresponding ocean load mass change. For the Late Pleistocene glacial history, we employ four different global ice models (Table 1). The first model, provided by Kurt Lambeck (see, e.g. Lambeck et al. 1998, 2000) and known as RSES (Kaufmann & Lambeck 2002), is a global ice model comprising Late Pleistocene ice sheets over North America, North Europe, the Barents Sea, Greenland, the British Isles, parts of Siberia, northern Russia, the Alps, and Antarctica. It combines the extent and the melting history from different separate ice models. The Laurentide and Greenland Ice Sheets have been taken from ICE-1 (Peltier & Andrews 1976), the Scandinavian and Barents Sea Ice Sheets from model FBK8 (Lambeck et al. 1998), the British Ice Sheet from model BK4 (Lambeck 1993), the Antarctic Ice Sheet from model ANT3 (Nakada & Lambeck 1988). In addition, we used the global ice models ICE-3G (Tushingham & Peltier 1991), ICE-4G (Peltier 1994) and ICE-5G (Peltier 2004). The extent of all ice sheets at the Last Glacial Maximum (LGM) approximately 21 400 yr BP is shown in Fig. 3. The ice margins of these models are quite similar over northern Europe with only small differences in the southeast, no ice cover over the Kara Sea area and the existence of an ice-bridge between the Scottish and Norwegian ice sheets at LGM for ICE-5G. ICE-4G and ICE-5G contain the most ice, with more than 2500 m over central Fennoscandia, while ICE-3G and RSES are less than 2000 m thick. The ice sheet maxima are centred over the Gulf of Bothnia and central Sweden, but with differing sizes and collapse histories. The models will therefore produce different patterns of rebound when implemented within a GIA model.

4 AREA OF INVESTIGATION

The gravity trend from GRACE is compared to the gravity change calculated from the GIA modelling in a selected range from 2°E to 40°E and 55°N to 70°N (Fig. 4). This area covers the main region of known uplift. The area is not highly influenced by hydrology or ocean changes. In addition, it covers locations of RSL data that have been used in the 1-D modelling by Steffen & Kaufmann (2005). The RSL observations are summarized in Lambeck et al. (1998). We have not further subdivided the area in a central and a far field like it has been done by Steffen & Kaufmann (2005) (triangles for central locations and inverse triangles for peripheral locations) to test for regional differences in the lithospheric thickness. Steffen & Kaufmann (2005) found for the central region a thick lithosphere ($H_L = 160$ km), while the peripheral region yielded a thinner lithosphere ($H_L = 100$ km). Preliminary tests within our analysis have shown that the current GRACE results cannot sufficiently distinguish between the crustal differences. As the subregions are smaller than our area of investigation, small errors in the post-processing lead to erroneous interpretations. Thus, we postpone this investigation to a future paper.

5 RESULTS

The GRACE-derived gravity changes within the area of investigation are compared to the gravity changes calculated with 1-D and 3-D earth models to determine best-fit earth models for Fennoscandia. We determine the least-squares misfit, defined by

\[
\chi = \sqrt{\frac{1}{n} \sum_{i=1}^{n} \left( \frac{o_i - p_i(a_i)}{\Delta o_i} \right)^2}
\]

for each GRACE data set in comparison to each 1-D or 3-D earth model. $n$ represents the number of data points in the $1° \times 1°$ grid of the investigation area (624 for the whole area), $o_i$ the corresponding GRACE observations for this grid, $p_i(a_i)$ the predicted gravity changes for a specific earth model $a_i$ and $\Delta o_i$ the data uncertainty of the GRACE observations.

The minimum value of $\chi$ within the parameter space of the 1-D earth models produces an earth model $a_i$, which fits the GRACE results best. In the ideal situation $\chi = 1$, the model is complete and the observational uncertainties are normally distributed with known standard deviations and uncorrelated. In addition, a confidence parameter

\[
\Psi = \sqrt{\frac{1}{n} \sum_{i=1}^{n} \left( \frac{p_i(a_0) - p_i(a_i)}{\Delta o_i} \right)^2}
\]

is calculated to bracket all earth models that fit the GRACE observations equally well as the best-fit 1-D earth model $a_0$ within the observational uncertainties. For all confidence parameters $\Psi \leq 1$, the predicted gravity change for a specific earth model $p_i(a_i)$ fits the observations as well as that of the best-fit 1-D earth model $p_i(a_0)$ within the $1\sigma$-uncertainty.

5.1 Results for 1-D models

As we compare the GRACE results to predictions from earth models with different ice history, we first discuss the impact of the ice load. This comparison is possible because, as we will see later, the inferred earth model parameters are comparable especially in view of the confidence parameter $\Psi \leq 1$. Fig. 5 summarizes the effects
Figure 3. Comparison of the ice extent at 21 400 yr BP (Last Glacial Maximum) in Fennoscandia from the four different ice models RSES (Lambeck et al. 1998), ICE-3G (Tushingham & Peltier 1991), ICE-4G (Peltier 1994) and ICE-5G (Peltier 2004).

The comparison shows the impact of Gaussian filter radius and ice models on the best misfits of GFZ and CSR GRACE g-dot solutions when compared with predictions from 1-D earth models. In general, the smallest misfits are found for ICE-5G ice history. RSES has higher misfits by 10–15 per cent. The largest misfits, at least twice as large as for ICE-5G, are determined for ICE-3G, while ICE-4G fits are intermediate in quality. Similar results are also found when using the other filter techniques discussed earlier. From this comparison it clearly follows that only ICE-5G and RSES need to be further investigated.

Next, we consider the impact on the inferred rheologic structure. Fig. 6 shows typical misfit maps when the predictions from various earth models with RSES ice history are compared to the gravity-rate-of-change from the GFZ and CSR GRACE gravity data using a Gaussian filter of 400 km. For GFZ (Fig. 6a) we find a model having a lithospheric thickness of \( H_l = 160 \) km, an upper mantle viscosity of \( \eta_{UM} = 4 \times 10^{20} \) Pa s and a lower mantle viscosity of \( \eta_{LM} = 10^{22} \) Pa s; for CSR \( H_l = 160 \) km, \( \eta_{UM} = 2 \times 10^{20} \) and \( \eta_{LM} = 10^{23} \) Pa s, respectively (Fig. 6b). The upper mantle viscosity is well constrained following a narrow band between \( \eta_{UM} = [3–6] \times 10^{20} \) Pa s for GFZ and \( \eta_{UM} = [1–3] \times 10^{20} \) Pa s for CSR.

Lithospheric thickness varies between 120 and 160 km for GFZ and 70 and 160 km for CSR. This variation is probably due to the lateral lithospheric variation under Fennoscandia as seen from seismic observations (cf. Tesauro et al. 2008, 2009). Although the western margin has lower thicknesses (less than 100 km), values as high as 160 km or more are found in the centre of the old continental root (e.g. Martinec & Wolf 2005; Steffen & Kaufmann 2005). This is supported by an examination of the confidence parameter with \( \Psi \leq 2 \). The envelope of solutions allow upper mantle viscosities higher than \( 10^{21} \) Pa s, and these high viscosities simulate thicknesses larger than 160 km. On another note, values of 160 km and larger are probably due to the truncation of signals with short wave-length in the GRACE solution, thus it only sees the signals with long-wavelength. Several studies with terrestrial data
find smaller thicknesses of 70–120 km (e.g. Lambeck et al. 1998; Wieczorkowski et al. 1999; Kaufmann & Wu 2002; Fleming et al. 2003; Martinec & Wolf 2005; Steffen & Kaufmann 2005).

The lower mantle viscosity is larger than the upper mantle viscosity by 1–2 orders of magnitude but is poorly resolved, which confirms earlier findings for Fennoscandia (e.g. Mitrovica 1996; Lambeck et al. 1998; Milne et al. 2001, 2004; Mitrovica & Forte 2004; Klemann & Wolf 2005; Martinec & Wolf 2005; Steffen & Kaufmann 2005; Steffen et al. 2006), whose sensitivity kernel rapidly decreases below 670 km depth.

Tables 2 and 3 summarize the results similar to Fig. 6 for all calculated combinations of solution centres and filter techniques including variation in filter radius. The third, fourth and fifth columns give the confidence range and the best fitting lithospheric thickness, upper and lower mantle viscosities with the value of $\chi$ on column 6. In general, a lithospheric thickness of 160 km is determined for a Gaussian filter radius of 600 km or less, but this result is independent of both the ice load and the filtering technique. Exceptions are found for the DDK× filter and for DMT-1, especially for ICE-5G history. Substantially, outliers are generated if the Gaussian filter is 800 km, for these remove and average far too much of the gravity change signal. The upper mantle viscosity bracketed is $[2–4] \times 10^{20}$ Pa s. The confidence areas become smaller for both larger radii and use of a destriping filter. The lower mantle viscosity is generally unconstrained, but if we ignore the DDK× results, values of $10^{22}$ Pa s are acceptable. The misfit difference between loads ICE-5G and RSES is generally less than 20 per cent and on average is about 15 per cent. Although there are better misfits for ICE-5G, the quality of fits confirms that both ice models are quite reasonable for GIA studies.

It is also interesting to look at the correlation between the GRACE gravity change and the rate of land uplift from the BIFROST GPS studies (Lidberg et al. 2007). Such correlation allows us to address the next question: which post-processing technique is better? This is possible as the change in gravity and uplift due to GIA are approximately proportional to each other (see, e.g. Ekman & Mäkinen 1996). The last column in Table 2 shows that there is good correlation (>88 per cent) if the filter radii is less than 800 km. The correlation increases for smaller filter radii to more than 93 per cent. Comparing the filtering techniques, the GRACE result with Gaussian filter correlates slightly better than the DDK× and much better than the destriping filter. Interestingly, a filter radius of 400 km seems to be most adequate, giving slightly better correlations than 350 or 450 km. Among the GFZ, CSR and DMT-1 solutions, the best correlation is found for the DMT-1 solution with 94.3 per cent.

From the comparison above, the best 1-D earth models determined agree quite well within a small part of the confidence area. Hence, the misfits are probably due to either filtering technique or the appropriate filter radii (Fig. 7). For GFZ, the misfits increase with increasing radius up to 600 km, then decrease with 800 km. In contrast, CSR shows an increase up to 800 km, while the misfits for CSR are smaller by as much as 40 per cent in comparison to GFZ fields. The DDK× also demonstrate this decrease in misfit.
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5.2 Comparison to 3-D models

The comparison of the different ice histories for the 1-D models has shown that RSES and ICE-5G give the best misfit. Thus, we only use these two models as surface load in the 3-D modelling. A cursory comparison revealed to us that among the 3-D model predictions with GRACE-derived g-dots, the greatest misfit reduction is achieved with the RSES model. This is in notable contrast to the results of the 1-D models. However, the difference between the misfits of RSES and ICE-5G is at most 16 per cent and on average about 8 per cent, and again confirms the high quality of both ice models.

Similar to the 1-D investigation, we first discuss the effects of both filtering techniques and GRACE solutions, except here the variation of mantle viscosity is determined completely by the \( \beta \) parameter. Fig. 8 shows the misfit comparison for the RF3S20 3-D model with ICE-5G (as noted earlier, RSES improves the misfits, on average, by 8 per cent) to GRACE uplift trends from GFZ and CSR with Gaussian filter of 350 km radius, destriping and DDK2 filter. As for the same as the one determined with GFZ. The combination of the two centre solutions seems to present a better correlation to BIFROST land uplift rate of up to 94 per cent, and thus such a combination may be appropriate for earth parameter investigations like this. Furthermore, with ICE-5G history, the combination GFZ+CSR using Gaussian filter of 350 km gives the best overall misfit with \( \chi = 1.12 \) (Table 3).

Figure 5. Comparison of best misfits of different GRACE solutions with different Gaussian filter radius after comparison to 1-D earth models with different ice loads. Red symbols: GFZ; blue symbols: CSR. Circles: RSES (Lambeck et al. 1998), squares: ICE-3G (Tushingham & Peltier 1991); crosses (Peltier 1994); diamonds: ICE-5G (Peltier 2004).
Figure 6. Best three-layer 1-D earth models using the ice model RSES and the confidence regions $\Psi \leq 1$ (dark grey) and $\Psi \leq 2$ (light grey) for the GFZ (a) and CSR (b) GRACE results with Gaussian filter radius of 400 km. The best model is marked with diamond, respectively. (A) GRACE data misfit as a function of lithospheric thickness and upper mantle viscosity for a fixed lower mantle viscosity of (a) $10^{22}$ Pa s and (b) $10^{23}$ Pa s. (B) GRACE data misfit as a function of upper and lower mantle viscosities for a fixed lithospheric thickness of 160 km.

The main results also hold for the RSES ice history with different filter radii. Fig. 9 shows the misfit comparison for RF3S20 with the RSES ice model with GRACE solutions and Gaussian radius of about 400 km. Here, we exchanged DDK2 with DMT-1, as the latter is designed for 400 km. The destriping filter has larger misfits than the Gaussian case, and CSR gives smaller misfit than GFZ. We can improve the misfit by combining GFZ and CSR. The misfits of DMT-1 are quite comparable to GFZ with Gaussian filter or to the combination of GFZ and CSR with destriping filter.

Wang et al. (2008) have used the misfits between prediction and observations for different $\beta$ values and for constraining the thermal versus non-thermal (e.g. chemical) contribution to observed seismic velocity variations. In view of such constraint on the thermal contribution, Figs 8 and 9 demonstrate that GRACE data in Fennoscandia reveal no particular preference for any $\beta$ value.
5.3 Hydrological contribution

The determination of the thermal contribution may be improved by removing other long-term gravity signals from the GRACE solutions. As introduced in Section 2, hydrology can be one of these signals. Therefore, we remove the long-term trend calculated with the global hydrology models GLDAS and WGHM from the GRACE results. Fig. 10 is similar to Fig. 9 except that the Gaussian filter of 400 km is used without any destriping, and hydrological contributions are removed from the data. As shown in Fig. 10, $\beta = 0.2$ still produces the poorest misfit, and $\beta = 0.6$ reveals improvement. However, when discussing the improvement on GRACE only, the GFZ misfits decrease when removing the hydrological contribution, while the CSR misfits increase, and this result holds for both of the hydrology models. Thus, with the 3-D modelling it is not possible to distinguish which hydrology model is more appropriately applicable.

To further clarify this, we also determined the 1-D radial profiles for GRACE solutions with removed hydrological contribution as calculated from the two models. A brief comparison to results without hydrological correction yields larger misfits for both models and, in certain cases, extremely different radial profiles that are, according to our current knowledge, absolutely unrealistic. Considering this, we conclude that both 1-D and 3-D models are actually corrupted by hydrological reduction. WGHM and GLDAS must be improved before they can be used in such an analysis.

6 CONCLUSIONS

A robust 7-yr gravity trend from GRACE observation was used to infer the lithospheric thickness and mantle viscosity under Fennoscandia for a 1-D earth model and the value of $\beta$ for 3-D earth models. This is the first investigation that employs GRACE data in combination of 1-D and 3-D earth models for Fennoscandian rebound. Two recent global ice models appear both adequate for our 1-D and 3-D investigation, RSES (Lambeck et al. 2000) and ICE-5G (Peltier 2004).

The radial earth structure in Fennoscandia is characterised by a lithospheric thickness of at least 90 km. In general, the solutions prefer 160 km which is most likely due to fact that short wavelengths are truncated in the satellite data. Upper mantle viscosity is constrained to be in the range $[2-4] \times 10^{20}$ Pa s. Solutions for the lower mantle viscosity are larger by 1–1.5 orders of magnitude, possibly centred about $10^{22}$ Pa s, but GRACE data over Fennoscandia cannot clearly resolve the value in the lower mantle. The Committee on Earth Gravity from Space (1997) suggested that a time
variable satellite gravity mission like GRACE could provide useful information on lower mantle viscosity and Velicogna & Wahr (2002) provided a quantitative forecast. Our results cannot support the optimism of this forecast at the moment, possibly because the investigation area is too small.

The 3-D modelling confirms the main results of the 1-D modelling. Unfortunately, the differentiation of the thermal contribution is quite poorly constrained. There is a slight preference for values of $\beta = 0.4$. An improvement may be possible by removing the hydrological contribution that can be calculated with global hydrology models. However, the corrected trends contain large uncertainties. Thus, one should use these hydrological models more cautiously in the correction of GIA signals in other regions such as North America.

The earth structure is found with different GRACE centre-specific solutions and filter techniques, which allowed a comparison and a determination of the best post-processing approach for an investigation such as this. We conclude that Gaussian filter of radii 350–450 km yield superior fits. We have also found that for studies regarding the earth structure in Fennoscandia, the destriping filter increases the misfit and decreases the correlation with GPS data. This cautions us not to use the destriping filter blindly in the determination of radial earth structure in North America (e.g. Paulson et al. 2007), because it is not clear how the destriping filter affects the outcome. In summary, the selection of a filter is important and that also depends on the area of study as well as the detailed aim of the investigation (see also Steffen et al. 2009c).

Regarding the solution centres, we found that CSR gives better misfits than GFZ, although the best earth models found only differ within small, acceptable ranges. We could improve the misfit by combining both to a new solution GFZ+CSR, which is an arithmetic-geometric mean. The improvement is up to a factor of 2 (for GFZ). The best 1-D model with ICE-5G ice history we find with the combination and a Gaussian filter radius of 350 km is characterised by a lithospheric thickness of 160 km, an upper mantle viscosity of $4 \times 10^{20}$ Pa s and a lower mantle viscosity of $2 \times 10^{22}$ Pa s. The misfit, the lowest of all, is $\chi = 1.12$. The combination of the two centre solutions also presents a better correlation to BIFROST land uplift rate of up to 94 per cent. Assuming that the modelling results, 1-D or 3-D, represent an almost realistic GIA signal when e.g. compared to terrestrial measurements (Fig. 1 and level-2 products will help give more insights concerning these topics within the next few years.

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**Table 3.** Same as Table 2, but for ICE-5G ice history (Peltier 2004) as surface load.

<table>
<thead>
<tr>
<th>Data set</th>
<th>Filter</th>
<th>$H_1$ (km)</th>
<th>$n_{ULM}$ $10^{20}$ (Pa s)</th>
<th>$n_{LM}$ $10^{22}$ (Pa s)</th>
<th>$\chi$</th>
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<td>G450</td>
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<td>3–5 (4)</td>
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<tr>
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<tr>
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<td>2–10 (10)</td>
<td>2.68</td>
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<td>3–5 (4)</td>
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Figure 7. Comparison of best misfits of different GRACE solutions with different filter techniques after comparison to 1-D earth models with RSES (Lambeck et al. 1998, circles) or ICE-5G (Peltier 2004, diamonds) ice load. Red, pink and light green symbols: GFZ with Gaussian, DDKx and destriping filter, respectively; blue, violet and dark green symbols: CSR with Gaussian, DDKx and destriping filter, respectively; yellow: average of CSR and GFZ with Gaussian filter; orange: DMT-1 with ANS filter. DDK1, DDK2 and DDK3 correspond to Gaussian filter radii of 530, 340 and 240 km, respectively.

Figure 8. Comparison of best misfits of different GRACE solutions with different filter techniques after comparison to 3-D earth models with ICE-5G (Peltier 2004) ice load. Red: GFZ; blue: CSR; black: average of CSR and GFZ. Diamonds: Gaussian filter with 350 km radius; dots: destriping filter + Gaussian filter with 350 km radius; crosses: DDK2.
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