Mass loss of the Greenland from GRACE time-variable gravity measurements

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Abstract:

The Gravity Recovery and Climate Experiment (GRACE) satellite data is used to estimate the rate of ice mass variability over Greenland. To do this, monthly GRACE level 2 Release-04 (RL04) data from three different processing centers, Center for Space Research (CSR), German Research Center for Geosciences (GFZ) and Jet Propulsion Laboratories (JPL) were used during the period April 2002 to February 2010. It should be noted that some months are missing for all three data sets. Results of computations provide a mass decrease of −163 ± 20 Gigaton per year (Gt/yr) based on CSR-RL04 data, −161 ± 21 Gt/yr based on GFZ-RL04 data and −84 ± 26 Gt/yr based on JPL RL04.1. The results are derived by the application of a non-isotropic filter whose degree of smoothing corresponds to a Gaussian filter with a radius of 340 km. Striping effects in the GRACE data, C20 effect, and leakage effects are taken into the consideration in the computations. There is some significant spread of the results among different processing centers of GRACE solutions; however, estimates achieved in this study are in agreement with the results obtained from alternative GRACE solutions.

Keywords: GRACE gravity mission, Greenland, Ice mass loss, leakage effects, non-isotropic filter.

1. Introduction:

Satellite gravity missions have been providing valuable information regarding Earth's gravity field. Due to global coverage of satellite missions, they provide an excellent tool for mapping the gravity field over large areas. GRACE mission not only maps the Earth's static gravity field but it also provides temporal variations of Earth's gravity field to a scale of several hundred kilometers and with a period of around one month. Thanks to the GRACE mission, changes in the gravity field which are caused by the redistribution of mass within the Earth and on or above the Earth's surface could be detected. In recent years, several research groups have used GRACE data to estimate the rate of ice mass change over Greenland. The main advantage of GRACE is that it is sensitive to the entire ice body.GRACE is not only sensitive to the entire ice body, but to all mass changes. And for the separation of these different mass changes, different geophysical models or ancillary data are necessary.

Baur et al. (2009a) used the monthly GRACE solutions Release-04 provided by GRACE processing centers of CSR (University of Texas), GFZ (Potsdam) and JPL (California) for the period 2002 to 2008. They estimated an average value of −162 ± 11 Gt/yr of the Greenland ice mass change for all available monthly GRACE data. Velicogna (2009) estimated a decrease of the Greenland ice mass of −230 ± 33 Gt/yr using the CSR (RL04) monthly solutions between 2002 and 2009. Another estimate by the same author amounts to −227 ± 33 Gt/yr for the Greenland ice body using the CSR monthly solutions Release 01 (RL01) during 2002 to 2006 (Velicogna and Wahr, 2006a). Wouters et al. (2008) estimated a value of −179 ± 25 Gt/yr for the Greenland ice mass loss. They used the CSR (RL04) monthly solutions from 2003 to 2008. Luthcke et al. (2006) used raw GRACE KBRR (K-Band Range and Range rate) data and their estimate was −101 ± 16 Gt/yr of the Greenland ice mass loss from 2003 to 2005. Using the CSR monthly solutions RL01 during 2002–2005, Chen et al. (2006) computed a decrease of ice mass of –219 ± 21 Gt/yr for the Greenland. Ramillien et al. (2006) used the same period as Chen et al. (2006) but using the GRGS/CNES GRACE solutions, and they estimated –109 ± 9 Gt/yr mass loss for the Greenland ice sheet.

It is clear that the ice mass estimates using monthly GRACE solutions are not all in agreement and the results differ significantly. The large differences in the estimates can partly be attributed to the different observation periods used combined with the large variability in Greenland's mass balance, but they are mainly due to the different methods used. Besides differences introduced by the different groups processing the raw data, they can be caused by truncating GRACE monthly coefficients differently, using different filters and different smoothing radii, and from failing to restore power lost by smoothing.

In this study, we estimate the Greenland ice mass change based on monthly GRACE solutions provided by the three different processing centers; CSR, JPL and GFZ during April 2002 to February 2010. The latest release (RL04) is used with improved geophysical signal models and data processing techniques resulting to smallest error among other releases (Bettadpur, 2007). To decorrelate the GRACE data, a filtering technique based on non-isotropic filter with different smoothing characteristics is applied. Note that all of the results reported above are based on isotropic filters. Spatial leakage effects are also accounted for in the computations. A software package has been developed in order to estimate the time series of mass changes using time-variable GRACE gravity measurements.

2. Surface mass change estimation from GRACE

The GRACE twin satellites launched in March 2002 and jointly implemented by the US National Aeronautics and Space Administration (NASA) and German Aerospace Center

(DLR) (Tapley et al., 2004a). GRACE measures Earth gravity changes with unprecedented accuracy by tracking the changes in the distance between the two satellites and combining these measurements with data from on-board accelerometers and Global Positioning System (GPS) receivers. GRACE data are used to determine monthly spherical harmonic coefficients of the Earth's gravity field. Each field consists of gravity field normalized (Stokes) coefficients, C_{lm} and S_{lm}, up to degree and order (*l, m*) 120 in JPL and GFZ products and 60 in CSR products (Tapley et al., 2004b). Using the static 30-day fully normalized spherical harmonic coefficients, one can estimate monthly local changes in surface mass changes (Wahr et al., 1998). The mass changes can be assumed in a very thin layer of water concentrated at the surface with a variable thickness. This assumption is not far from reality as changes in water storage in hydrologic reservoirs, by moving ocean, atmospheric and cryospheric masses, and by exchange among these reservoirs causes monthly changes in gravity signals (Chambers, 2007). The vertical extent of the water is much smaller than the horizontal scales of the changes and this vertical column height is called equivalent water thickness. Mass variations are modeled as surface density variations $\Delta\sigma$ (the unit of $\Delta\sigma$ is mass/surface area) in a spherical layer. The mass change within a given region is expressed as the following integral over the unit sphere:

$$\Delta m = \int \Delta\sigma(\varphi,\lambda)\tau(\varphi,\lambda)\cos\varphi \mathrm{d}\varphi \mathrm{d}\lambda \tag{1}$$

where φ and λ are the latitude and longitude of the point of interest and

$$\tau(\varphi,\lambda) = \begin{cases} 0 & \text{outside the region} \\ 1 & \text{inside the region} \end{cases} \tag{2}$$

To estimate mass changes according to Eq.(1) from GRACE monthly gravity field solutions, one must estimate the satellite derived surface density changes. Invoking the multiple expansions in spherical coordinates and using surface harmonic functions, the surface density variations can be generally expressed as follow (Wahr et.al., 1998):

$$\Delta\sigma(\varphi,\lambda)=a\rho_{\mathrm{w}}\sum_{l=0}^{\infty}\sum_{m=0}^{l}\bar{P}_{lm}(\sin\varphi)\left[\Delta\hat{C}_{lm}\cos m\lambda+\Delta\hat{S}_{lm}\sin m\lambda\right] \tag{3}$$

where a is the major semi axis of a reference ellipsoid and $\bar{P}_{lm}$ is the normalized associated Legendre function of the first kind. ρ_{w} is the mass-density of freshwater (assumed throughout this paper to be 1000 kg/m3), and is included here so that $\Delta\hat{C}_{lm}$ and $\Delta\hat{S}_{lm}$ are dimensionless. It should be stated here that $\Delta\sigma/\rho_{\mathrm{w}}$ transforms surface mass-densities to equivalent water thickness values. Typically, $\Delta\hat{C}_{lm}$ and $\Delta\hat{S}_{lm}$ are residuals with respect to a background model or set of models. It can be shown that there is a simple relation between $\Delta\hat{C}_{lm}$, $\Delta\hat{S}_{lm}$ and ΔC_{lm}, ΔS_{lm} as (Wahr et al., 1998):

$$\begin{Bmatrix}\Delta\hat{C}_{lm}\\ \Delta\hat{S}_{lm}\end{Bmatrix}=\frac{\rho_{\mathrm{ave}}}{3\rho_{\mathrm{w}}}\frac{2l+1}{1+k_l}\begin{Bmatrix}\Delta C_{lm}\\ \Delta S_{lm}\end{Bmatrix} \tag{4}$$

where ΔC_{lm} and ΔS_{lm} are time-variable components of the GRACE observed Stokes coefficients for some month of degree and order (l, m) or as changes relative to the mean of the monthly solutions. Also k_l is the load Love number of degree l which is given in Wahr et al. (1998) and ρ_{ave} is the average mass-density of the solid Earth (5517 kg/m3 in this study). Therefore, monthly spherical harmonic coefficients of the Earth's gravity field can be used to estimate monthly local changes in surface mass density (Wahr et al., 1998):

$$\Delta\sigma(\varphi,\lambda)=\frac{a\rho_{\mathrm{ave}}}{3}\sum_{l=0}^{\infty}\sum_{m=0}^{l}\frac{2l+1}{1+k_l}\bar{P}_{lm}(\sin\varphi)\left[\Delta C_{lm}\cos m\lambda+\Delta S_{lm}\sin m\lambda\right] \tag{5}$$

Crucial for a reliable estimate of secular mass changes from GRACE monthly solutions is the ability to correct for the systematic errors in the surface mass density computation as discussed below.

Due to the nature of the measurement technique in GRACE and mission geometry, the monthly spherical harmonic coefficients are contaminated with short-wavelength noises (Kusche et al., 2009). The correlated and resolution dependent noise in the coefficients are not white-noise on the sphere and it has usually 'striping' pattern. The noise is significant when one is interested in signals of geographical extension of a few hundreds km or in using the higher degree coefficients (short-wavelengths). Generally, there are two different kind of filters to the removal of the noise; isotropic and non- isotropic filters. An isotropic filter has weight depending only on the degree l, whereas a non-isotropic filter has weight depending on the degree and order (l, m). Non-isotropic filters are used in this study since the GRACE noise structure mainly manifests itself as near north-south "stripes" and it has a non-isotropic nature. Non-isotropic filters have been demonstrated to be useful in many specific applications. Kusche (2007) devised a non-isotropic filter algorithm, which is similar to a Tikhonov-type regularization of the original normal equation system based on a systematic error covariance matrix computed from the GRACE orbits, and an a priori signal covariance matrix in the spherical harmonic domain. Using this filter, Kusche et al. (2009) has analyzed GRACE RL04 monthly gravity solutions in three different smoothing degrees of corresponding Gaussian Radius (CR): CR1 = 530 km, CR2 = 340 km and CR3 = 240 km. This will be called corresponding radius (CR) throughout this paper. The results presented by Kusche et al. (2009) were in good agreement with mass anomalies derived from a global hydrological model. In this study we use the decorrelation and smoothing method of Kusche et al. (2009) to correct monthly GRACE RL04 gravity models. The three above-mentioned corresponding radii are used. Moreover, Kusche (2007, 2009) presented two strategies to infer this corresponding radius with significantly differing results. In this study, the strategy is based on comparing the 'isotropic part' of the anisotropic decorrelation filter with the

Gaussian in terms of matching the particular spectral degree where the filter weight drops to 0.5.

Due to the GRACE orbit geometry and the separation length between its satellites, the lowest-degree zonal harmonics, C_{20} (or in another format as J_2) cannot be well determined from the GRACE data (Tapley et al. 2004b). The C_{20} estimates from GRACE also are well-known to be affected by significant long-period tidal aliases. Therefore, some of the previous studies excluded the C_{20} value in the estimation of surface mass density (e.g. Wahr et al., 2004). The replacement of the GRACE C_{20} coefficient by its estimate from Satellite Laser Ranging (SLR) improves the estimation of mass variations from GRACE (Chen et al. 2005), The SLR time series are also more precise, with about a third of the noise of the GRACE time series. Therefore, the monthly SLR estimates for C_{20} coefficient are used to replace the estimates from GRACE in this study. The SLR time series for C_{20} coefficient are taken from J. Ries (personal communication, 2010).

The mass redistribution over land and ocean causes the mass changes detected by the GRACE data. For a reliable estimate of secular mass changes over Greenland one needs to correct for leakage effects. The leakage effect originates, for example, from limited spatial resolution and imperfect reduction of satellite measurement errors (Swenson et al., 2002). On the one hand, mass change at a place outside Greenland propagates into a signal spreading over Greenland and has an impact on the Greenland mass change estimates. On the other hand, mass change over Greenland propagates into a signal spreading over outside the Greenland area. These are called leakage in and leakage out effects, respectively. Generally speaking, signals spread spatially and leak to the surrounding region, and theoretically over the entire Earth. The leakage effects are a major obstacle to deriving reliable estimate of secular mass changes. The leakage out signals must be returned to the area concerned and the leakage in signals must be removed from the interested area. Not correcting for the leakage effects is one of the reasons for the differences among GRACE-derived mass change estimates. Baur et al. (2009a) investigations showed, without taking leakage effects into account, ice mass change estimate over Greenland could be reduced at most by a factor of 2. In this study, the leakage effects are estimated based on a two-step procedure as follows and we only use GRACE data for leakage effects computations. Step 1: The coefficients associated with the leakage effects are calculated using Eq.(6) below and integrating $\Delta\sigma$ (φ, λ), only on the area concerned, which is the boundary of the region function τ in Eq.(2) shown in Fig. 1:

$$\begin{Bmatrix} \Delta C_{lm} \\ \Delta S_{lm} \end{Bmatrix} = \frac{3}{4\pi a \rho_{\mathrm{ave}}} \frac{1+k_l}{2l+1} \int_0^{2\pi} \mathrm{d}\lambda \int_0^{\pi} \cos\varphi \mathrm{d}\varphi \times \Delta\sigma(\varphi,\lambda)\, \tilde{P}_{lm}(\sin\varphi) \begin{Bmatrix} \cos m\lambda \\ \sin m\lambda \end{Bmatrix} \tag{6}$$

There may be two candidates for the input data, i.e. $\Delta\sigma\ \varphi\lambda\ (\ )$, in Eq.(6): the first candidate could be calculated from global hydrological model values (e.g. Global Land Data Assimilation System (GLDAS), Rodell et al., 2004a) and the second candidate could be derived from GRACE data alone. The second candidate is used in this study. Step 2: Using Eq.(5) and the coefficients derived from Eq.(6), the leakage signal is computed only for points outside the area concerned. To remove the leakage effects from the estimated surface mass density, the leakage in signals must be subtracted and the leakage out signals must be added to the estimation. The sources generating leakage in signals could be from all over the world however the impact reduces with increasing distances. This is because the leaking signals follow the Newton's law of gravitation. The strongest signals on Greenland can be caused by Alaska, Fennoscandia and the Canadian Shield. These three sources are also used in Baur et al. (2009a) investigations. We also use the same extended area chosen by Baur et al. (2009a) for the computation of the leakage out effects. This is the area that the most delineation (the leakage out effects) is already included (see Fig. 1). The extended area is the mass change counter line according to −7 cm isoline expressed in terms of equivalent water thickness. The difference between the mass change within the extended area and the result over Greenland is the leakage out effect. The procedure is the same for calculation of the leakage in effect as the leakage out from Region 1 to Region 2 is the same as leakage in to Region 2 from Region 1. However, the computations must be repeated for each individual source leaking into area of interest. It should be noted that the treatment of leakage might be responsible for some parts of the differences in mass change estimations. Closer sources of leaking signals than Canadian Shield, Alaska and Fennoscandia are not considered in this study.

In early 2007, reprocessed GRACE RL04 time-variable gravity fields with improved background geophysical models and data processing techniques were released (Bettadpur, 2007). These data have significantly smaller predicted errors than the previous release. For example, the estimated error in the monthly calculation of ocean mass is 3.9 mm of water thickness for RL01 accumulated to degree/order 60 with no smoothing and 1.3 mm after 300 km smoothing. The cumulative error of RL04 coefficients for the same global ocean basin is only 1.1 mm with no smoothing and 0.5 mm with 300 km smoothing (see Chambers, 2009). Thus, unsmoothed RL04 coefficients give a comparable result to smoothed RL01 coefficients, when averages are computed over areas as large as the global ocean. Also, atmospheric and oceanic mass changes have been largely removed from RL04 using numerical model predictions, so that variations over time scales of months to years should reflect primarily un-modeled effects such as snow/ice mass changes (including polar ice body and mountain glaciers), plus other geophysical signals such as postglacial rebound (PGR) and co-seismic and post-seismic deformation. This makes it almost impossible to separate between gravity signals caused by ice-mass variations and signals caused by PGR. In the estimation for the ice mass-change rates in this study, the contaminating factors like the effects of variations in atmospheric mass and the solid Earth contribution from high-latitude PGR are not applied.

The continental hydrology is another important signal that is not included in the background models. The atmospheric effects are negligible for Greenland on the long term trend (Velicogna and Wahr, 2006a, b). We also chose not to apply the correction for the PGR signal, considering the total uncertainty in the PGR estimations (Velicogna and Wahr, 2006a, b). It is left to interested readers to choose their preferred PGR model. There are still some open issues in the context of PGR signal solutions. Nevertheless, it should be stated here that the PGR signal for the entire Greenland is computed to about −7.4 Gt/yr with a standard deviation of ±19 Gt/yr (Velicogna and Wahr, 2006b). It should be noted that this estimation could be at the very lower end of PGR. When comparing to the ice-mass estimates the PGR signal is more than one order of magnitude smaller. To complete this section, Fig. 2 shows a summary of the computational steps in this study to estimate the ice-mass change trends over Greenland.

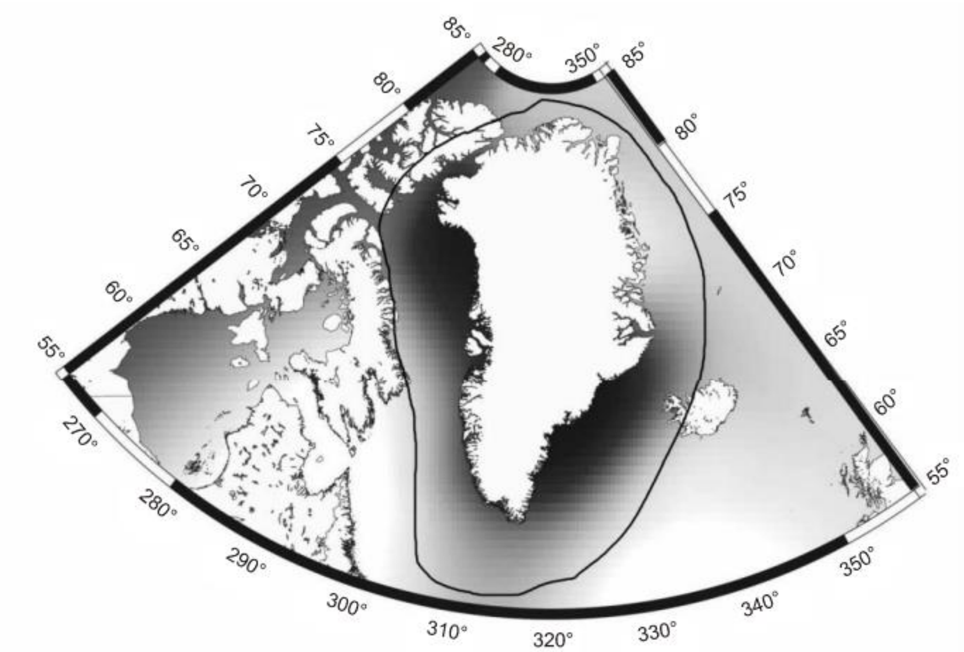

Fig. 1. GRACE-derived equivalent water thickness variations from April 2002 to February 2010 in cm. CSR monthly solution and a corresponding radius of 340 km are chosen. The boundary of the region function τ in Eq.(2) is the white-colored Greenland ice sheet. The extended area used in the leakage effects computations is the mass change contour line according to −7 cm isoline shown approximately by the closed solid line.

3. Numerical investigations

We estimate the secular trend in Greenland ice mass rate using more than 8 years of GRACE level 2 RL04 data. Three independently estimated series of monthly GRACE solutions by CSR, GFZ and JPL processing centers during the period April 2002 to February 2010 are used. The data sets are shortly described in Table 1. RL04 coefficients are distributed on the level-2 data archives as GSM files (GSM is a file extension). The GSM files contain spherical harmonic coefficients representing the gravity field of the Earth. As mentioned in Section 2, monthly solutions of the GRACE when computing ice mass rates include an unphysical striping error pattern which can be considered as noises and must be decorrelated/filtered. They have been filtered using Kusche et al. (2009) method in the three different corresponding radii (530 km, 340 km, and 240 km). The monthly SLR estimates for C_{20} coefficient are used to replace the estimates from GRACE to complete the data edition step.

In order to estimate the Greenland ice mass rate, the surface mass density changes (in terms of equivalent water thickness for each month) is computed using Eq.(5). To do this, the time-mean of the coefficients from April 2002 to February 2010 is computed and by removing the mean from monthly spherical harmonic coefficients, the monthly coefficients anomalies ΔC_{lm} and ΔS_{lm} are determined. Using the coefficients anomalies and applying Eq.(5) on a 1° × 1° grid, one can estimate monthly mass variability over Greenland and its surrounding (see also Chen et al., 2006). Our preliminary investigations showed that the choice of this grid size was sufficient and did not influence significantly the ice mass change estimations.

Next step is to form an approximate estimate of total mass change for each month, by summing over grid elements with cosine latitude weighting. In this estimation, the leakage effects are corrected. According to the described algorithm in Section 2, the leakage effects have been accounted for the monthly mass change estimations. Fig. 3a–c shows Greenland monthly mass changes estimated in Gt among calibrated error bars from three GRACE data sets released by GFZ, JPL and CSR. All three datasets have been filtered in the three different corresponding radii (530 km, 340 km, and 240 km).

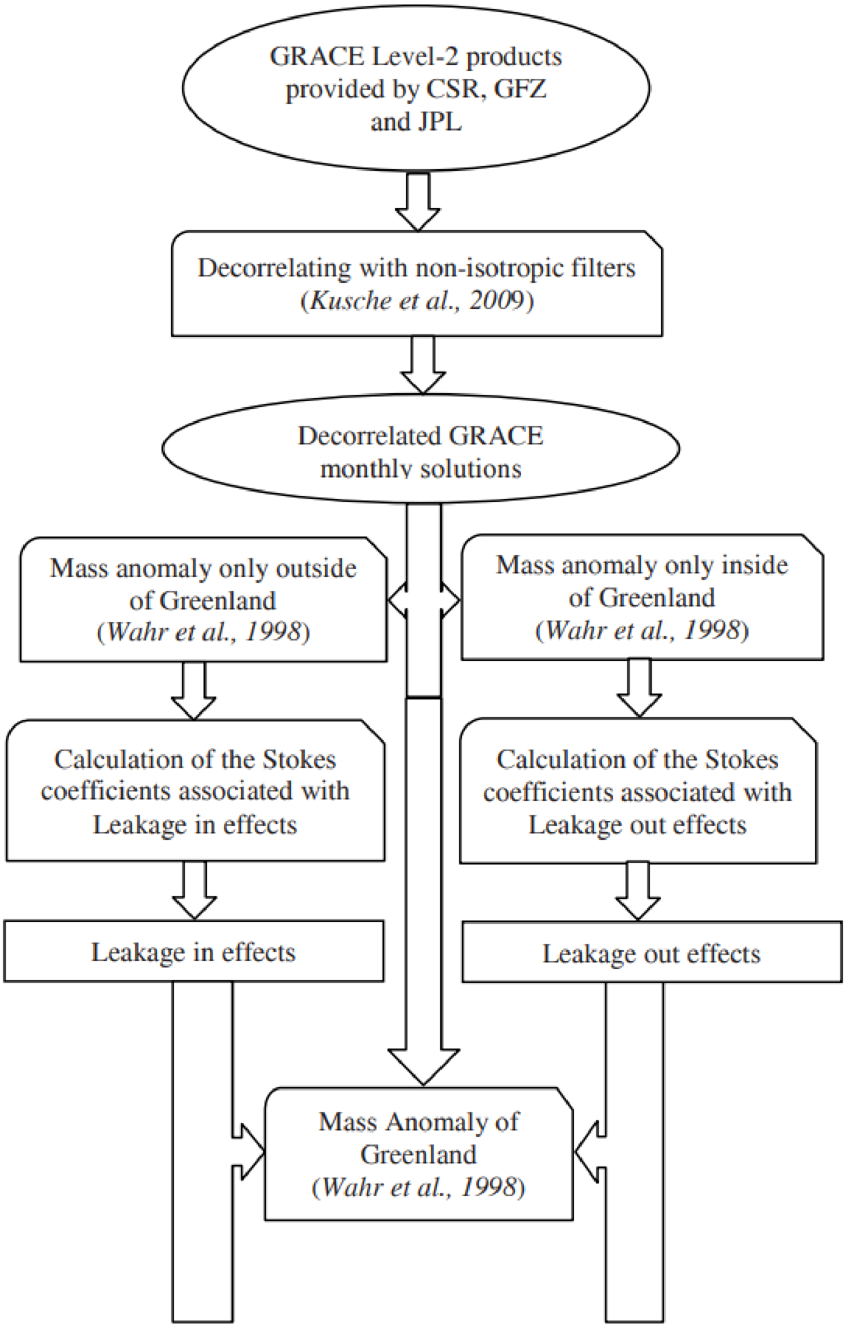

Fig. 2. The procedure for estimating the Greenland ice mass change by GRACE time variable gravity measurements.

Table 1. Description of the monthly GRACE solutions used in the computations. For all three data sets, the months June–July 2002 and June 2003 are missing due to missing accelerometer data. $(l, m)_{max}$ is the maximum degree and order of the spherical harmonic expansions as delivered by the processing centers.

	CSR	GFZ	JPL
Release	04	04	04.1
Epochs	92	85	88
Start	04-2002	08-2002	04-2002
End	02-2010	01-2010	11-2009
$(l, m)_{max}$	60	120	120

To detect the secular trend and periodic variations in the monthly mass anomalies, a general expression of the following form can be used:

$$f\left(\varphi, \lambda, t\right) = \mathrm{A} + \mathrm{Bt} + \sum_i C_i \cos\left(\omega_i t\right) + D_i \sin\left(\omega_i t\right) + \varepsilon \tag{8}$$

Here, the value of the considered functional f (the ice mass anomaly, here) at a selected location $\left(\varphi, \lambda\right)$ and time t is approximated by a static value A, and its secular (B) and periodic (with amplitude C_i and D_i of typical angular frequencies ω_i) variations. The variable ε characterizes noise and un-modeled effects. This model is applied to a time series of grids from which Fig. 4 is derived. A crucial problem which arises in the determination of secular trends and periodic variations is that whether all components have to be modeled simultaneously or not. Ignoring some systematic components contained in the data or including some components into the model which are not contained in the data, might cause a bias to the estimated parameters. To explore annual, semiannual and other periodic variations one can use the Akaike's Information Criterion (AIC) (Akaike, 1973). It should be noted that there are other criteria to choose amongst the candidate models. The Bayesian Information Criterion (BIC), Cross Validation (CV) approach and Hypothesis Testing are among them (see also Baur et al., 2009b). The AIC value is calculated according to the following equation (Yamamoto et. al., 2008):

$$\mathrm{AIC} = -2\left(\frac{1}{\sqrt{2\pi\sigma^2}} \sum_{\mathrm{i=1}}^{\mathrm{N}} \left(x_{\mathrm{obs}}^{(i)} - x_{\mathrm{calc}}^{(i)} \right)^2 - \frac{N}{2} \ln 2\pi\sigma^2 - k \right) \tag{9}$$

where N is the number of observed data, σ is the standard deviation of the fitting. computed from $(n^{-1} \sum_{i=1}^{n} \hat{e}_i^2)^{1/2}$, where $\hat{e}_i$ are the estimates residuals for a particular candidate model., $x_{\mathrm{obs}}^{(i)} - x_{\mathrm{calc}}^{(i)}$ is the difference between the observed and calculated mass anomaly value at the time point i, and k is the total number of estimated parameters used for fitting. The second term in right hand side of Eq. (9) accounts for the criterion of a good statistical fit. A better fit yields a smaller AIC value among other models considered.

Table 2. Akaike's Information Criterion (*AIC*) values and the corresponding trend (in brackets, the unit is Gt/yr) estimated using three corresponding radii (*CR*) 240 km, 340 km, and 530 km for the Greenland ice mass change and GRACE solutions provided by CSR, GFZ and JPL.

CR [km]	Bias + Trend	Bias + Trend + Annual Variations	Bias + Trend + Annual + Semi-Annual Variations	Bias + Trend + Annual + Semi-Annual + Seasonal Variations
	CSR			
240	−2654 (−169 ± 45)	−3085 (−170 ± 47)	−2300 (−171 ± 47)	−2229 (−171 ± 46)
340	−3352 (−162 ± 21)	−3486 (−163 ± 20)	−2670 (−163 ± 21)	−2583 (−163 ± 21)
530	−3920 (−147 ± 10)	−3950 (−148 ± 11)	−2958 (−149 ± 11)	−2838 (149 ± 10)
	GFZ			
240	−2018 (−172 ± 55)	−1938 (−170 ± 51)	−2075 (−171 ± 53)	−2108 (−171 ± 53)
340	−2163 (−159 ± 20)	−1921 (−160 ± 21)	−2101 (−160 ± 21)	−2247 (−161 ± 21)
530	−2631 (−146 ± 9)	−2348 (−146 ± 9)	−2513 (−146 ± 8)	−2683 (−147 ± 8)
	JPL			
240	−3660 (−91 ± 72)	−1560 (−91 ± 71)	−1379 (−92 ± 70)	−1271 (−92 ± 71)
340	−4292 (−84 ± 26)	−1553 (−85 ± 27)	−1542 (−85 ± 27)	−1448 (−85 ± 26)
530	−4721 (−82 ± 10)	−1696 (−83 ± 11)	−1769 (−83 ± 11)	−1659 (−84 ± 10)

Table 3. Greenland secular ice mass change estimated from GRACE monthly gravity field solutions provided by CSR using Bias + Trend + Annual Variations model, GFZ using Bias + Trend + Annual + Semi-Annual + Seasonal Variations model and JPL using Bias + Trend model. Three filtering corresponding radii 240 km, 340 km, and 530 km are used. The unit is Gt/yr.

	530 km	340 km	240 km
CSR	−148 ± 11	−163 ± 20	−170 ± 47
GFZ	−147 ± 8	−161 ± 21	−171 ± 53
JPL	−82 ± 10	−84 ± 26	−91 ± 72

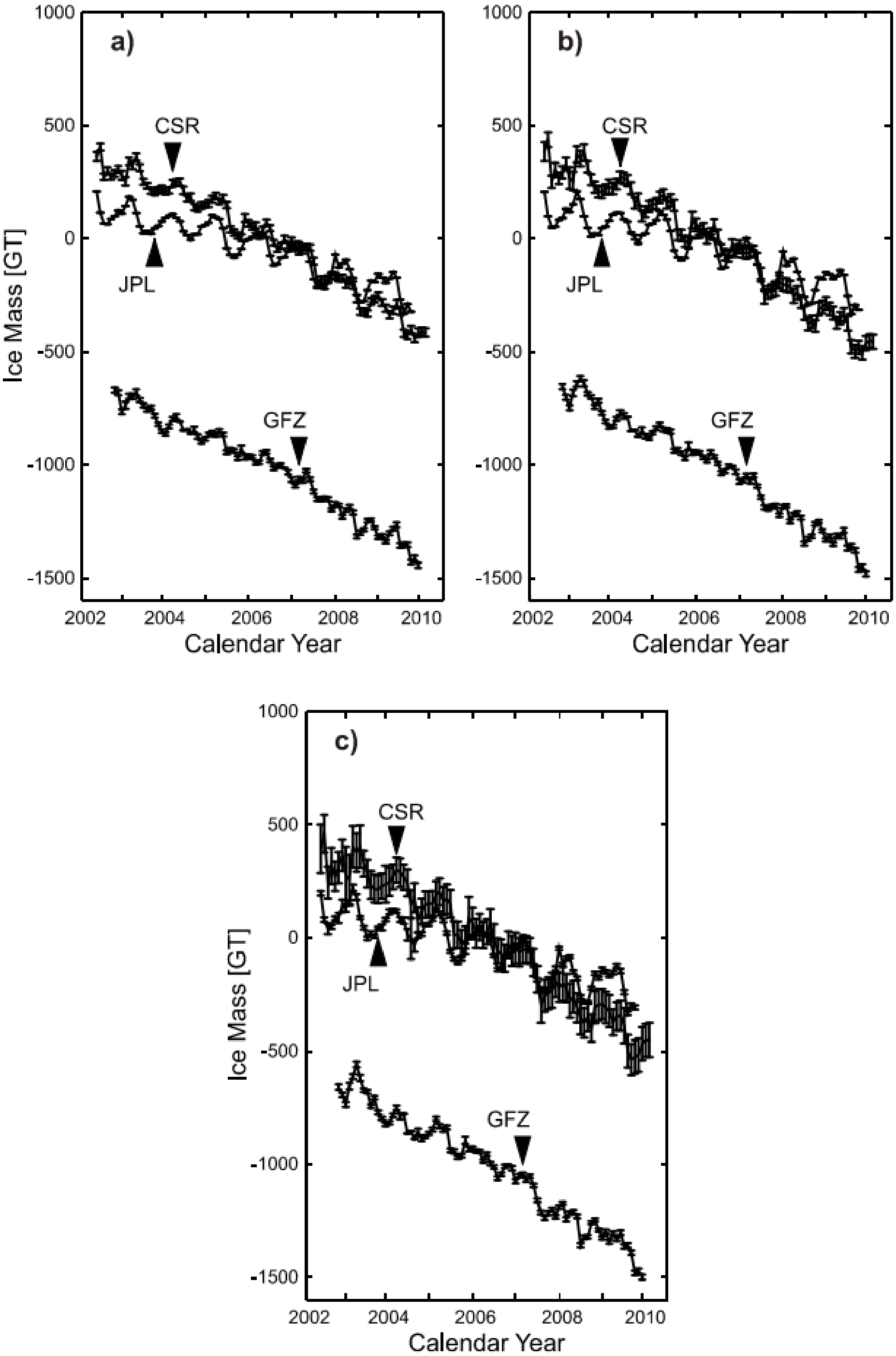

Fig. 3. Estimated Greenland monthly mass change with error bars from April 2002 to February 2010. GRACE datasets are from CSR, GFZ and JPL processing centers. Data have been decorrelated and filtered by *Kusche et al. (2009)* method in three corresponding radii: **a)** 530 km, **b)** 340 km and **c)** 240 km. The calibrated error bars include only the contributions from 1σ uncertainties in the GRACE gravity field.

To explore the periodic variations, the AIC values for the Greenland ice mass estimates are calculated. Table 2 shows the results of AIC calculations. Three corresponding radii 530 km, 340 km, and 240 km are used. Three estimations are made using filtered GRACE time series of CSR, GFZ and JPL. According to Eq.(8), different candidate models including bias, trend, annual, semi-annual and seasonal variations (3-months period) components are examined. Considering the AIC criterion, the optimum fit amongst the candidate models is chosen and using un-weighted least squares method, a secular trend for Greenland ice mass change is estimated. GRACE gravity field solutions provided by CSR, GFZ and JPL from April 2002 to February 2010 are used. Depending on the corresponding radii, different results are achieved.

Finally, Table 3 lists the estimated mass change trends. These estimates are based on the candidate models chosen according to AIC values. Section 4 details the description of the calculation and used models.

4. Discussion:

Our ice-mass change estimates from GRACE monthly solutions with application to Greenland are summarized in Fig. 3. Important features in Fig. 3 are: 1) GRACE level 2 Release-04 datasets from three processing centers of CSR, GFZ and JPL are used to compute the Greenland mass changes; 2) non- isotropic filter in three different corresponding radii are used to decorrelate high frequency GRACE measurements provided by high degree terms and order of the Stock’s coefficients; 3) a method based on potential forward modeling is applied to estimate leakage in and out effects; and 4) un-weighted least squares method is used to estimate a bias, trend and four annual and semiannual terms as well as seasonal variations for the Greenland mass changes.

Several studies to estimate Greenland mass balance used monthly solutions of one GRACE processing center to calculate the secular trend for the mass change. Analysis of the gravity field solutions by CSR, GFZ and JPL reveals a significant spread (see also Sandberg

Sørensen and Forsberg, 2010). This is clearly shown in Fig. 3. Trend patterns in Fig. 3 reveal that the JPL mass change estimates over Greenland is significantly smaller in magnitude.

The GRACE noise structure mainly manifests itself as near north-south stripes and therefore we chose to decorrelate all GRACE products. Several recent studies use Gaussian isotropic filters and just one smoothing radius besides that a variety of alternative smoothing kernels have been proposed. Choosing an effective smoothing radius is critical for processing and understanding GRACE-observed time variable gravity. This effective smoothing radius represents the spatial resolution of the GRACE data, which is a key indicator of the quality of the GRACE data, and has implications for its utility in a range of applications. The spatial resolution also plays an important role in correctly interpreting GRACE observed mass change (Wahr et al., 2004; Tapley et al., 2004b; Rodell et al., 2004b). As shown in Fig. 3, if the spatial radius is small, the error among mass change estimates is large and the derived mass change fields may be overly noisy, while if large, the error among mass change estimate is small and the derived fields may be overly smoothed. Successfully determining the effective spatial radius requires either a priori knowledge of the spatial extent of the true signal, or significant experience or intuition regarding what it might be. As shown in the study by Swenson et al. (2003), the minimum of the sum of GRACE measurement errors and the leakage errors typically occurs in the range 200 km to 600 km. As such, the non-isotropic filter for three corresponding radii of 240 km, 340 km, and 530 km is applied in our investigation. Note that the errors shown in Fig. 3 are estimated based solely on calibrated standard deviations released by GRACE data centers and the fact that different smoothing radii of non-isotropic filter affect the error of estimations.

To take into account the leakage effects, several studies used the principle of rescaling the mass change estimates within an area of interest in order to restore realistic signals. They consider the total leakage signal in the target area altogether (see e.g. Velicogna and Wahr, 2005). Opposed to this principle, the method used in this study separates the leakage signal into the leakage in and leakage out effects. In fact, the method is twofold. First step is to compute leakage in signals which are spread in the area of interest from other areas and second step is to compute leakage out signals which are spread out of the area of interest. The strongest signals leaking into the Greenland are located in Alaska, Fennoscandia and the Canadian ice sheet (see e.g. Baur et al., 2009a). The most dominant leakage in effect is the significant positive signal from the Canadian Shield. The correction for leakage in effects is performed by algebraic subtraction. As an example, the average leakage in and leakage out effects for GFZ monthly gravity solutions and smoothing degree of corresponding Gaussian radius of 340 km is estimated to 13.7 Gt and 40.9 Gt, respectively. The values for CSR and JPL monthly gravity solutions with the same smoothing degree are 7.7 Gt and 17 Gt, and 11 Gt and 33.3 Gt, respectively.

Finally, to estimate the secular trend over Greenland using un-weighted least squares method, we firstly calculate the AIC values. The philosophy behind the AIC values is to select among candidate models relative to each other according to maximum log likelihood. The Maximum Likelihood Estimator (MLE) is an alternative to the Minimum Variance Unbiased Estimator (MVUE). In contrast to the MVUE, the MLE almost always exists and can be computed. For this reason, the MLE is one of the most common estimator procedure used in practice. In order to specify temporal characteristics adequately, AIC values for different candidate models of i) bias and trend, ii) bias, trend and annual variations, iii) bias, trend, annual and semi-annual variations, and iv) bias, trend, annual, semi-annual and seasonal variations are calculated. As it is shown in Table 2, the least squares model including the bias, trend and annual variations has the smallest AIC value for CSR, except for the products filtered by CR1 = 530 km, which just the bias and trend has the smallest value. However, the least squares model including the bias, trend and annual variations has the second smallest value. Therefore, it is decided to use the least squares model including the bias, trend and annual variations for the final computations of secular trend over Greenland with CSR products (see Table 3). In the case of GFZ products, the least squares model including bias, trend, annual, semi-annual and seasonal variation components yields the smallest AIC values. This candidate model is used to estimate the final ice mass change over Greenland with GFZ products (see Table 3). Table 2 shows that the least squares model containing the bias and trend parameters have the smallest AIC value for JPL products. Final estimates of Greenland secular ice mass change is based on the AIC criteria with smallest AIC value. The annual ice mass loss for Greenland becomes –163 ± 20 Gt/yr for CSR, –161 ± 21 Gt/yr for GFZ and –84 ± 26 Gt/yr for JPL monthly gravity solutions. These results are based on the corresponding radius of 340 km. Note that these values are free of any PGR corrections. PGR signals are more than one order of magnitude smaller than ice mass loss signals. The results of the individual contribution from the three data processing centers and with different corresponding radii are shown in Table 3. The errors listed in Table 3 take into account the errors of the least squares adjustments of the mathematical model which is used to detect the secular trend and periodic variations in the monthly mass anomalies, the leakage effects and the gravity field error. In estimation of these errors, the PGR effects are not applied.

Table 4. The Greenland mass balance from GRACE monthly gravity field solutions provided by JPL. Smoothing radius of 340 km used for these computations. The numbers in columns 2–6 are the average values of the total mass with respect to 2002–2009 mean. A-M-J is the April-May-June total mass and A-S-O is August-September-October total mass. Total masses are in Gt.

	A-M-J	A-S-O	Summer Loss	Winter Gain	Net
2002	316	110	–206	---	---
2003	293	48	–245	183	–62
2004	161	13	–148	113	–35
2005	157	–109	–266	144	–122
2006	34	–167	–201	143	–58
2007	–44	–326	–282	123	–159
2008	–175	–425	–250	151	–99
2009	–252	–447	–195	173	–22

The temporal evaluation of Greenland's mass balance (see Fig. 3) shows that mass increases slowly between October and April (see also Wouters et al., 2008). It also shows mass decrease between May and September. The difference between the April-May-June and August-September-October mean mass over Greenland is calculated in Table 4. We chose to only present the results of JPL GRACE solution in Table 4. The pattern is almost the same for the GFZ and CSR solutions. April-May-June (A-M-J) manifests the beginning of the melt season while August-September-October (A-S-O) indicates the end of the melt season. In Table 4, average values of the total mass with respect to the 2002–2009 mean are given for A-M-J and A-S-O. Winter gain, summer loss and net balance are also listed. The summer loss is calculated from the comparison of A-S-O average value with respect to the A-M-J value at the same year and the winter loss is calculated from the comparison of A-S-O average value in preceding year with the A-M-J value in the next year. The summer ice loss values are different over the years, with a maximum in 2007 in which −282 Gt was lost (see also Wouters et al., 2008). The summer ice mass losses are somehow compensated by ice mass increase in the preceding winter seasons. Similar trends with different magnitudes are observed for 2003, 2005 and 2009 with −245 Gt, −266 Gt and −195 Gt ice mass losses during summer. Largest winter ice mass increase is observed for 2003 (183 Gt). It compensates the mass loss in coming summer (see also Wouters et al., 2008). In the winter 2008–2009, the total Greenland ice mass gain was calculated only 173 Gt, and with ice mass loss −195 Gt, resulting in a net mass loss of −22 Gt for the whole 2008–2009. The net balance for 2007–2008 is comparable with the result of 2008–2009, but the net for the entire 2006–2007 is the largest net mass loss of −159 Gt. The summers of 2003, 2005 and 2007 are observed to be the three warmest years since 1961 (see Hanna et al., 2009). The GRACE indicates large mass losses in these three years. This can be an indication of a strong correlation between summer temperature increase and the amount of ice mass loss observed by GRACE.

To show secular mass changes over Greenland, the changes in the equivalent water thickness between April 2002 and February 2010 are depicted in Fig. 4. This estimation is derived from CSR GRACE monthly solutions applying Eq.(8), including bias, trend, and annual variations. Decorrelated models and the non-isotropic filter with corresponding radius of 340 km are chosen. The general pattern of mass change shows thinning along the coast and a slight growth in the inland regions. Pronounced trends are found in the coastal zones from the east to the southeast. Widespread mass loss in the northwestern coastal zones is also observed. The interior parts of Greenland shows less negative trend and the northern and northeastern parts show the least negative trends.

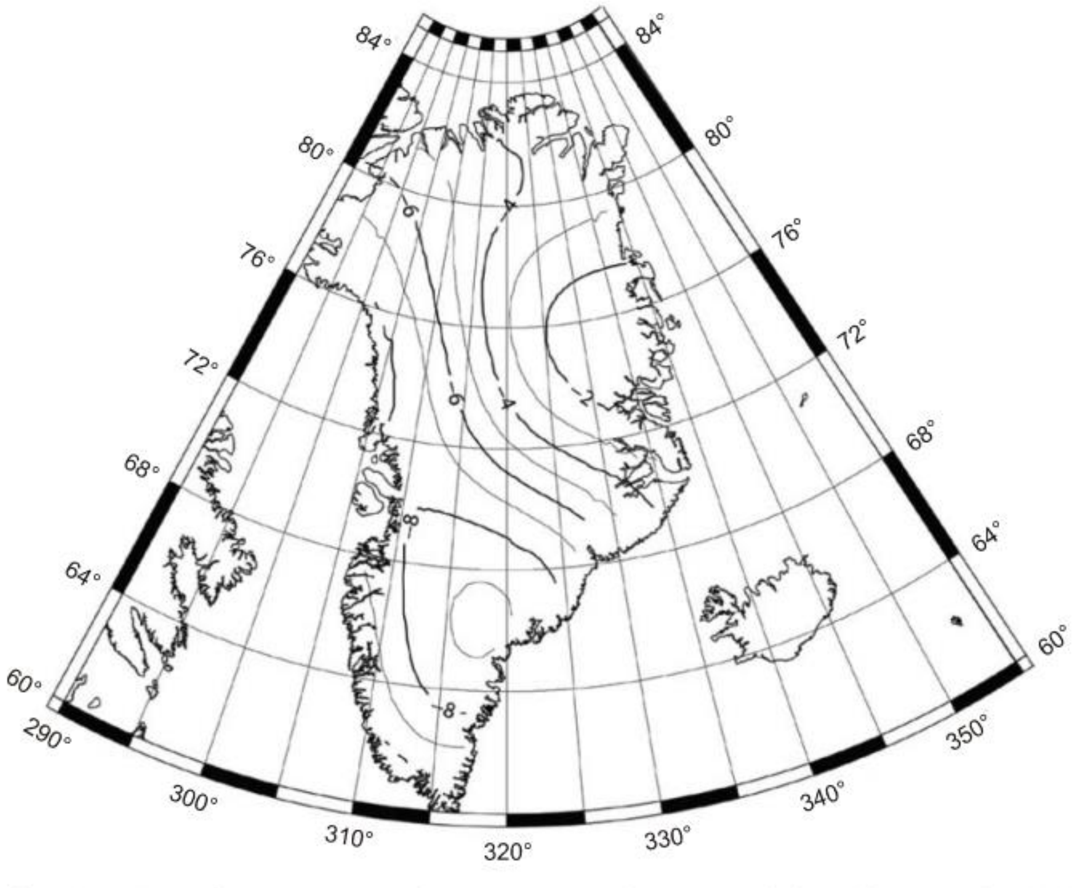

Fig. 4. GRACE-derived secular equivalent water thickness variations from April 2002 to February 2010 (in cm). CSR monthly solution and a corresponding radius of 340 km are chosen.

5. Conclusions:

The GRACE twin satellites have been providing a continuous record of the Earth's gravity field over more than 8 years, which offers an excellent tool to study mass changes over large areas. Acceleration of mass loss over Greenland is reported in several studies consistent with increased global warming in recent years, and indicates that Greenland is a major contributor to recent global sea level rise (Solomon et al., 2007). Accurate estimates of the Greenland ice mass change, accompanied by its error estimation, improve uncertainties in sea level change studies. Given the size and shape and complexity of the Greenland ice body, it makes it difficult to measure ice mass change in the Greenland. A variety of techniques are used to estimate Greenland ice mass balance each of which with limitations and uncertainties. The spherical harmonic coefficients of monthly solutions given by GRACE twin satellites allow regional estimation of Greenland ice mass balance. In contrast to most other techniques, GRACE measures Greenland mass variability over the entire ice sheet. Furthermore, to obtain this mass variability, the process is less ambiguous for GRACE as the relationship between gravity and mass variability follows directly from Newton's law. The main disadvantage of GRACE models for obtaining the Greenland mass change is errors caused from mismodeled postglacial rebound. GRACE is unable to separate gravitational effects of the Greenland ice sheet from those of the underlying solid Earth.

Our GRACE estimate of the total Greenland mass loss using GRACE level 2 RL04 data from the three different processing centers of CSR, JPL and GFZ during April 2002 to February 2010 is:

- −163 ± 20 Gt/yr based on CSR-RL04 using Bias + Trend + Annual Variations model,

• −161 ± 21 Gt/yr based on GFZ-RL04 using Bias + Trend + Annual + Semi-Annual + Seasonal Variations model, and

• −84 ± 26 Gt/yr based on JPL-RL04.1 using Bias + Trend model.

These values are estimated with corresponding radius of 340 km. Even though the total mass losses are very different, the same pattern is seen in the three mass change models.

Ice mass change estimates of Greenland, in this study, are in agreement with other recent GRACE studies, however, it should be noted that each study is characterized by its observation period, individual analysis method and different monthly gravity solutions. Therefore, it would be very difficult to compare different GRACE studies, objectively.

Our GRACE estimates showed that the ice mass loss was not constant and trends were increasingly negative as suggested originally by Velicogna and Wahr (2006a), but now supported by various criteria and using longer periods for the GRACE data. Although the mass loss of the ice sheet is not constant, we decided to represent the GRACE observations by a linear trend.We cannot observe a significant acceleration term over whole data series. This trend is also pointed out by various recent research studies as reported in the introduction. This emphasizes the need for continuous observation of ice mass balance and extending time series of ice mass changes estimated from the Earth's gravity signal provided by GRACE and future missions.

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References

Akaike H., 1973. Information theory and an extension of maximum likelihood principle. In: Petrov B.N. and Csáki F. (Eds.), 2nd International Symposium on Information Theory. Akadémiai Kaidó, Budapest, Hungary, 267–281.

Baur O., Kuhn M. and Featherstone W.E., 2009a. GRACE-derived ice-mass variations over Greenland by accounting for leakage effects. J. Geophys. Res., 114, B06407, DOI: 10.1029/2008JB006239.

Baur O., Kuhn M. and Featherstone W.E., 2009b. GRACE-Derived Linear and Non-Linear Secular Mass Variations over Greenland. Publication No. 213, The Institute for Geoscience Research (TIGeR), Curtin University of Technology, Australia.

Bettadpur S., 2007. CSR Level-2 Processing Standards Document for Product Release 04. Report GRACE 327-742. Center for Space Research, University of Texas at Austin, Austin, TX.

Chambers D.P., 2007. Effects of ice melting on GRACE observations of ocean mass trends. Geophys. Res. Lett., 34, L05610, DOI: 10.1029/2006GL029171.

Chambers D.P., 2009. Calculating trends from GRACE in the presence of large changes in continental ice storage and ocean mass. Geophys. J. Int., 176, 415–419, DOI: 10.1111/j.1365-246X.2008.04012.x.

Chen J.L., Wilson C.R. and Tapley B.D., 2006. Satellite gravity measurements confirm accelerated melting of Greenland ice sheet. Science, 313, 1958–1960, DOI: 10.1029/science.1129007.

Chen J.L., Wilson C.R., Famiglietti J.S. and Rodell M., 2005. Spatial sensitivity of the Gravity Recovery and Climate Experiment (GRACE) time-variable gravity observations. J. Geophys. Res., 110, B08408, DOI: 10.1029/2004JB003536.

Hanna E.P., Cappelen X., Fettweis P., Huybrechts P., Luckman A. and Ribergaard M., 2009. Hydrologic response of the Greenland ice sheet: The role of oceanographic warming. Hydrol. Process., 23, 7–30, DOI: 10.1002/hyp.7090.

Kusche J., 2007. Approximate decorrelation and non-isotropic smoothing of time-variable GRACE type gravity fields. J. Geodesy, 81, 733–749.

Kusche J., Schmidt R., Petrovic S. and Rietbroek R., 2009. Decorrelated GRACE time-variable gravity solutions by GFZ, and their validation using a hydrological model. J. Geodesy, 83, 903–913.

Luthcke S.B., Zwally H.J., Abdalati W., Rowlands D.D., Ray R.D., Nerem R.S., Lemoine F.G., McCarthy J.J. and Chinn D.S., 2006. Recent Greenland ice mass loss by drainage system from satellite gravity observations. Science, 314, 1286–1289, DOI: 10.1126/Science.1130776.

Ramillien G., Lombard A., Cazenave A., Ivins E.R., Llubes M., Remy F. and Biancale R., 2006. Interannual variations of the mass balance of the Antarctica and Greenland ice sheets from GRACE. Glob. Planet. Change, 53, 198–208, DOI: 10.1016/j.gloplacha.

Rodell M., Houser P.R., Jambor U., Gottschalck J., Mitchell K., Meng C.-J., Arsenault K., Cosgrove B., Radakovich J., Bosilovich M., Entin J.K., Walker J.P., Lohmann D. and Toll D., 2004a. The global land data assimilation system. Bull. Amer. Meteorol. Soc., 85, 381–394.

Rodell M., Famiglietti J.S., Chen J.L., Seneviratne S.I., Viterbo P., Holl S. and Wilson C.R., 2004b. Basin scale estimates of evapotranspiration using GRACE and other observations. Geophys. Res. Lett., 31, L20504, DOI: 10.1029/2004GL020873.

Sandberg Sørensen L.S. and Forsberg R., 2010. Greenland ice sheet mass loss from GRACE monthly models. In: Mertikas S.P. (Ed.), Gravity, Geoid and Earth Observation. International Association of Geodesy Symposia 135, 527–532, Springer-Verlag, Berlin, Heidelberg, Germany, DOI: 10.1007/978-3-642-10634-7_70.

Solomon S., Qin D., Manning M., Chen Z., Marquis M., Averyt K.B., Tignor M. and Miller H.L., 2007. Climate Change 2007 - The Physical Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the IPCC. Cambridge University Press, Cambridge, U.K.

Swenson S. and Wahr J, 2002. Methods for inferring regional surface-mass anomalies from Gravity Recovery and Climate Experiment (GRACE) measurements time-variable gravity. J. Geophys. Res., 107, B92193, DOI: 10.1029/2001JB000576.

Swenson S., Wahr J. and Milly P.C.D., 2003. Estimated accuracies of regional water storage variations inferred from the Gravity Recovery and Climate Experiment (GRACE). Water Resour. Res., 39, 1223, DOI: 10.1029/2002WR001808.

Tapley B.D., Bettadpur S., Watkins M.M. and Reigber C., 2004a. The Gravity Recovery and Climate Experiment: Mission overview and early results. Geophys. Res. Lett., 31, L09607, DOI: 10.1029/2004GL019920.

Tapley B.D., Bettadpur S., Ries J.C., Thompson P.F. and Watkins M.M., 2004b. GRACE measurements of mass variability in the Earth system. Science, 305, 503–505.

Velicogna I., 2009. Increasing rates of ice mass loss from the Greenland and Antarctic ice sheets revealed by GRACE. Geophys. Res. Lett., 36, L19503, DOI: 10.1029/2009GL040222.

Veliconga I. and Wahr J., 2005. Greenland mass balance from GRACE. Geophys. Res. Lett., 32, L18505, DOI: 10.1029/2005GL023955.

Velicogna I. and Wahr J., 2006a. Acceleration of Greenland ice mass loss in spring 2004. Nature, 443, 329–331. Velicogna I. and Wahr J., 2006b. Measurements of time-variable gravity show mass loss in Antarctica. Science, 311, 1754–1756, DOI: 10.1126/science. 1123785.

Wahr J., Molenaar M. and Bryan F., 1998. Time variability of the Earth's gravity field: Hydrological and oceanic effects and their possible detection using GRACE. J. Geophys. Res., 103(B12), 30205–30229.

Wahr J., Swenson S., Zlotnicki V. and Velicogana I., 2004. Time-variable gravity from GRACE: First results. Geophys. Res. Lett., 31, L11501, DOI: 10.1029/2004GL019779.

Wouters B., Chambers D.P. and Schrama E.J.O., 2008. GRACE observes small-scale mass loss in Greenland. Geophys. Res. Lett., 35, L20501, DOI: 10.1029/2008GL034816.

Yamamoto K., Fukuda Y., Doi K. and Motoyama H., 2008. Interpretation of the GRACE-derived mass trend in Enderby Land, Antarctica. Polar Sci., 2, 267-276, DOI: 10.1016 /j.polar.2008.10.001.