Greenland mass balance from GRACE

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[1] We use 22 monthly GRACE (Gravity Recovery and Climate Experiment) gravity fields to estimate the linear trend in Greenland ice mass during 2002–2004. We recover a decrease in total ice mass of 82 ± 28 km^3 of ice per year, consistent with estimates from other techniques. Our uncertainty estimate is dominated by the effects of GRACE measurement errors and errors in our post glacial rebound (PG) correction. The main advantages of GRACE are that it is sensitive to the entire ice sheet, and that it provides mass estimates with only minimal use of supporting physical assumptions or ancillary data. **Citation:** Velicogna, I., and J. Wahr (2005), Greenland mass balance from GRACE, *Geophys. Res. Lett.*, *32*, L18505, doi:10.1029/2005GL023955.

1. Introduction

[2] Greenland is one of the largest reservoirs of fresh water on Earth. Its large accumulation rate and significant runoff and meltwater generation, make the Greenland ice sheet a highly dynamic place. Rapid changes in the Greenland ice sheet could impact global sea level and hold the potential of altering the North Atlantic thermocline circulation and global climate.

[3] The Greenland contribution to sea level rise between 1990 and 2100 has been projected to be between −0.02 and +0.09 cm [*Church et al.*, 2001]. Better estimates of its present-day contributions would permit more accurate projections. The dramatic thinning observed in the 1990s at low elevations [*Krabbe et al.*, 2000], and the increased mass loss in more recent years [*Box et al.*, 2004; *Krabbe et al.*, 2004; *Regno*, 2005] make monitoring the ice sheet particularly timely.

[4] In this paper we estimate the secular trend in Greenland mass during 2002–2004 using satellite time variable gravity measurements from GRACE. This is the first measure of mass change over the entire Greenland ice sheet from space.

2. Initial GRACE Estimates

[5] The GRACE mission, administered by NASA and Deutsche Zestful für Luff-una Ramparts, was launched in March, 2002. Its goal is to map the Earth's gravity field to high accuracy every 30 days during its 8–9 year lifetime. GRACE consists of two identical satellites in identical orbits, separated by ∼220 km. The satellites use microwaves to monitor their separation distance. On-board accelerations and GAPS receivers detect non-gravitational accelerations and geocentric orbital motion.

[6] GRACE personnel use these data to determine ∼monthly spherical harmonic coefficients of the Earth's gravity field. To date, twenty-two fields have been released to users, corresponding roughly to Apr/May, Aug, Sept, Oct, Nov, 2002; Feb, Mar, Apr, May, Jul, Aug, Sep, Oct, Nov, Dec, 2003; and Jan, Feb, Mar, Apr, May, Jun, Jul, 2004. Each field consists of gravity field (Stokes) coefficients, C_{lm} and S_{lm}, up to l, $m \leq 120$. These are the coefficients in a spherical harmonic expansion of the geoid [see, e.g., *Wahr et al.*, 1998]. The subscripts l and m are the degree and order of the spherical harmonic, and the horizontal scale is ≈20,000/l km. The GRACE C_{20} coefficients show anomalously large variability in the first few months, so we replace them with values derived by *Cheng and Tapley* [2004] from satellite laser ranging data. The Stokes coefficients can be used to solve for monthly gravity field changes, and thus to estimate monthly variations in the Earth's surface mass distribution. The GRACE fields can currently provide high-latitude (60° and above) estimates of monthly mass changes to accuracies of 10 mm in equivalent water thickness when averaged over discs of radius 600–700 km and larger. *Wahr et al* [2004] and *Tapley et al.* [2004] show some initial results.

[7] Monthly Greenland mass variability is estimated from the GRACE Stokes coefficients by removing the 22-month mean, and applying equation (2) of *Swenson et al.* [2003] to the residuals. We use an averaging function that minimizes the combined measurement error and signal leakage, constructed assuming the GRACE measurement errors are 40 × the baseline error estimate. GRACE does not recover l = 1 coefficients, so we remove l = 1 harmonics from the averaging function. We scale the averaging function so that if it is applied to a uniform mass change of 1 cm water thickness over all regions within a few hundred km of the Greenland coast, but zero in the interior, it returns an average Greenland value of 1 cm. This scaling is motivated by laser altimeter data suggesting that the largest Greenland mass changes are concentrated at the edges [*Krabbe et al.*, 2004]. The scaled averaging function is shown in Figure 1. If, instead, we had scaled the averaging function so that a 1 cm mass change spread evenly over all Greenland returned an average value of 1 cm, the averaging function - and all linear trend results shown below - would be reduced by 20%. However it is not realistic to assume that the ice sheet is thinning uniformly.

[8] Results for the monthly Greenland mass estimates are indicated by the asterisks in Figure 2. The error bars are obtained by convolving our Greenland averaging function with uncertainty estimates for the GRACE Stokes coefficients. Our method of estimating those

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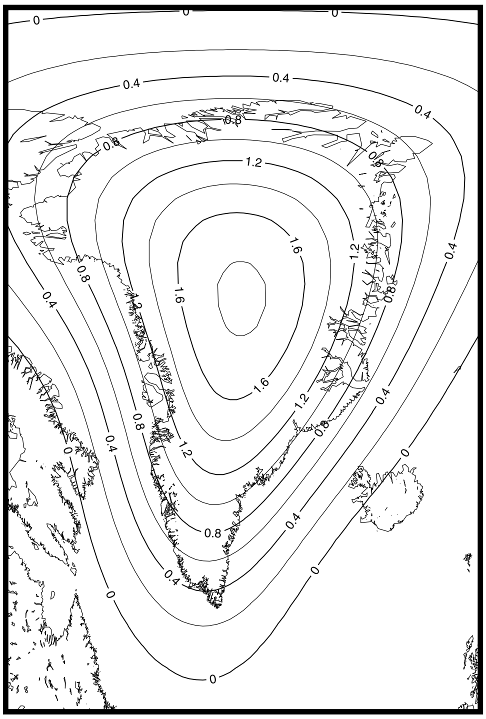

Figure 1. The averaging kernel used to estimate the change in total Greenland mass.

uncertainties is described by *Velicogna et al.* [2005]. The measurement errors are largely uncorrelated between months (S. Bettadpur, personal communication, 2004).

3. Removing Leakage

[9] The averaging kernel extends beyond the boundaries of Greenland. Thus, geophysical signals outside Greenland leak into our estimates. The leakage is more of a problem than it appears from Figure 1; our omission of $l = 1$ terms means the averaging kernel extends around the globe, though with an amplitude too small to see in Figure 1.

[10] We evaluate several possible sources of leakage. To estimate the contamination from continental hydrology outside Greenland, we use monthly, global water storage fields from the Noah Land Surface Model produced with the Global Land Data Assimilation System (GLDAS) [*Rodell et al.*, 2004]. We also compute water storage leakage using the Climate Prediction Center (CPC) model of Y. Fan and H. van den Dool (The Climate Prediction Center global monthly soil moisture data set at 1/2° resolution for 1948 to present, submitted to *Journal of Geophysical Research*, 2004). We adopt the GLDAS results because they display more realistic snowpack results. We use the difference between the GLDAS and CPC estimates as a measure of uncertainty in the hydrology leakage.

[11] Contamination from the ocean is estimated using a JPL version of the ECCO general circulation model [*Lee et al.*, 2002]. To mimic the process used by the GRACE project to de-alias the raw data, we remove the output of a barotropic ocean model [*Ali and Zlotnicki*, 2003] from the ECCO results. We use the barotropic Arctic Ocean model of *Stepanov and Hughes* [2004] to estimate leakage from the Arctic Ocean, which is not included in the ECCO or barotropic models. For each hydrology and ocean model we use output for the same months as the GRACE fields.

[12] We estimate the leakage from these models by calculating their contributions to the Stokes coefficients, and applying the same averaging function used for the GRACE fields. The circles in Figure 2 show the GRACE results after removing the leakage estimates. The amplitude of the leakage is small, approximately equal to the GRACE error bars.

4. The Trend

[13] The results in Figure 2 show a clear trend, superimposed on short-period variability. The short-period terms have been discussed by *Velicogna et al.* [2005]. To recover the trend we simultaneously fit a trend and annually and semiannually varying terms to the monthly GRACE-minus-leakage results shown in Figure 2. We obtain a mass decrease of 40 ± 11 mm/yr of equivalent water thickness, as shown in Figure 2. We convert that trend to a rate of change in total Greenland ice mass, by multiplying by 1.98×10^6 km^2, the ice covered area of Greenland [*Loveland et al.*, 2000] and using an ice density of 917 kg/m^3, to obtain a mass decrease of 87 ± 23 km^3/yr.

[14] The uncertainty is computed by assuming the errors in the monthly solutions are the root-sum-squares (RSS) of the measurement uncertainties (the error bars in Figure 2) and the uncertainties in the leakage estimates. The leakage uncertainties are computed as the differences between the GLDAS and CPC leakage estimates. No attempt is made to assess ocean model errors; the linear trend from the ocean leakage estimates is only half the trend from the hydrology estimates.

[15] Uncertainties related to atmospheric mass fluctuations have not been included in our error estimates because we believe they are small. GRACE data processing employs ECMWF fields to remove atmospheric contributions to gravity prior to solving for the Stokes' coefficients. We evaluate the errors in the atmospheric fields by comparing the ECMWF pressure fields with pressure observations

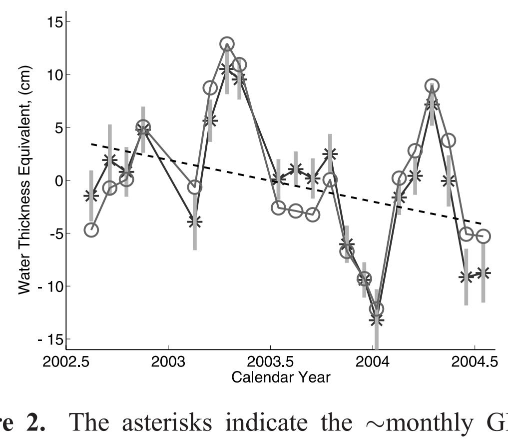

Figure 2. The asterisks indicate the ~monthly GRACE estimates of total Greenland mass variability, in cm of equivalent water thickness. Each value includes an error bar. The circles show the GRACE results, after removing the estimated hydrology + ocean leakage. Also shown is the line that best fits the circles.

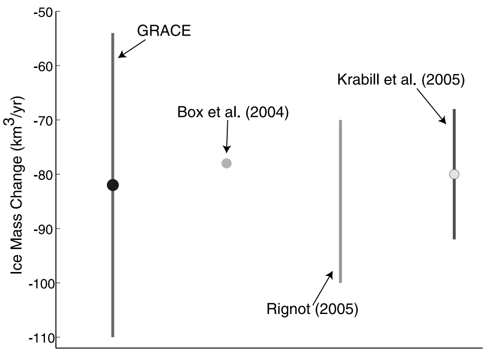

Figure 3. Compares the GRACE estimate of the decrease in total Greenland mass, with three recent, published estimates, derived independently of GRACE. Volumes are in km^3 of ice per year. The vertical lines indicate uncertainties.

from meteorological stations in the World Meteorological Organization catalog and from the AWMO Greenland weather stations (K. Steffen, personal communication, 2005). The variance of atmospheric errors is less than 3% of the GRACE variance.

5. The PG Signal

[16] GRACE cannot distinguish between secular gravity signals caused by present-day changes in Greenland ice and those caused by PG: the viscoelastic response of the solid Earth to glacial unloading over the last several thousand years. To isolate the present-day ice mass signal we independently estimate the PG secular gravity signal and remove it from the GRACE trend given in Section 4. We use the ICE-5G global ice history model of *Peltier* [2004], convolved with viscoelastic Green's functions for an incompressible Earth, computed as described by *Velicogna and Wahr* [2002b]. We use *Peltier*'s [1996] 2-layer approximation to the VM2 mantle viscosity profile used to derive ICE-5G (i.e. 3.6×10^{21} Pa-s below 1170 km depth, and 0.9×10^{21} Pa-s between 1170 km and the bottom of the lithosphere at 120 km depth). We compute the trends in the Stokes coefficients for this model, and convolve them with the Figure 1 averaging function. We obtain an apparent ice mass decrease of 5 km^3/yr.

[17] A decrease in apparent mass seems incompatible with the expectation that the Earth beneath Greenland should be rebounding upward due to the Holocene removal of Greenland ice. But Greenland lies outside the forebulge of the Pleistocene ice sheet in northern Canada, and so there is subsidence caused by the removal of that ice sheet. The 5 km^3/yr estimate results from the near-cancellation of signals caused by the Greenland and non-Greenland ice deglaciation histories, each with an amplitude of about 22 km^3/yr.

[18] There are three general sources of error in our PG estimates: the ice history, the viscosity profile, and physical and numerical approximations in the model. To estimate the possible effects of errors in the Greenland component of ICE-5G, we replace it with model GREEN1 from *Fleming and Lambeck* [2004], and repeat the viscoelastic convolution. We use the difference of the two secular gravity field predictions as the level of uncertainty due to errors in the Greenland deglaciation model. To estimate the effects of errors in the deglaciation history outside Greenland, we apply a similar procedure, replacing the non-Greenland components of ICE-5G with those of ICE-3G [*Tushingham and Peltier*, 1991].

[19] To estimate the uncertainty caused by errors in the viscosity profile, we reconvolve the ICE-5G ice model with a variety of 2-layer viscosity profiles that have been adopted in previous PG studies. We take the difference between extreme cases as a measure of the uncertainty. The lower mantle viscosity, for example, is chosen to vary between 2.0×10^{21} and 50.0×10^{21} Pa-s, and the depth to the lower mantle is allowed to vary between 660 km and 1170 km. It is inconsistent to change the viscosity profile without also changing ICE-5G, since ICE-5G was constructed to match observations when convolved with VM2 Green's functions. But in the absence of an alternative, we adopt this method as a means of obtaining an upper bound uncertainty.

[20] The most questionable geophysical approximation in our model is probably our omission of compressibility, which is likely to affect our predicted gravity field rates by $\sim$10% [*Tamisiea et al.*, 2002]. We approximate this error as 10% of the contribution from the Greenland deglaciation alone, rather than as 10% of the (much smaller) total contribution. To be conservative, we assume that unknown errors related to numerical approximations are 20% of the contribution from the Greenland deglaciation.

[21] The total PG error is thus the sum of errors from: (1) the Greenland ice model; (2) the ice model outside Greenland; (3) the viscosity profile; (4) the omission of compressibility; and (5) possible numerical problems. We estimate these introduce errors into our GRACE-minus-PG Greenland mass estimate of: (1) ±5 km^3/yr; (2) ±4 km^3/yr; (3) ±17 km^3/yr; (4) ±2 km^3/yr; and (5) ±5 km^3/yr. The viscosity profile uncertainty causes by far the largest error. Computing the RSS of these individual uncertainties to get the total PG uncertainty, we obtain a PG contribution to the GRACE estimate of ice mass decrease of 5 ± 19 km^3/yr, which must be subtracted from the GRACE results. When we do this, and take the RSS of the errors in the GRACE fit and in the PG correction, we obtain our final estimate of the decrease in total Greenland mass between the summers of 2002 and 2004: 82 ± 28 km^3/yr.

6. Summary and Discussion

[22] Greenland has been a major contributor to recent global sea level change. Greenland mass variability has a strong dynamic component that is difficult to measure given the size and complexity of the ice sheet. Previous estimates of mass variability have been obtained using a variety of techniques, each of which has intrinsic limitations and uncertainties. A problem common to these techniques is their difficulty in monitoring the entire ice sheet.

[23] *Krabbe et al.* [2004] used airborne laser altimeter measurements to estimate a mass loss of 80 ± 12 km^3/yr during 1997–2003. They converted altimeter surface-elevation rates into mass change rates by multiplying by the density of ice. Where altimeter measurements were not available they estimated mass variability using

a monthly version of a degree-day runoff/retention model, with ECMWF surface air temperature and precipitation/evaporation values. This method of supplementing the altimeter coverage includes dynamical ice flow contributions only from glaciers the authors identified as major contributors.

[24] *Box et al.* [2004] estimated a mass loss of 78 km^3/yr during 1991–2000, comparing modeled accumulation-minus-melt with an estimate of mass discharge based on steady state conditions. Hence their estimate is biased by an uncertainty in mass discharge.

[25] *Regno* [2005] compared measured ice flux with observed accumulation minus modeled melt estimates, to obtain a mass loss of between 70 and 100 km^3/yr in 2000, and several 10 km^3/yr more in 2004 due to the acceleration of several glaciers. His estimate does not include all glaciers along the west coast of Greenland.

[26] In contrast, GRACE measures mass changes over the entire ice sheet. Furthermore, the process of inferring mass variability is less ambiguous for GRACE than for most other techniques; the relationship between gravity and mass follows directly from Newton's Law of Gravity. For altimetry, for example, the transformation of elevation differences into mass variability requires knowledge of firn density, which is usually poorly known. This accentuates the problem of determining linear trends over short time spans since high frequency variability in accumulation has more impact on elevations than on mass (fluctuations in firn density are smaller at high frequencies). The main disadvantage of GRACE for obtaining total Greenland variability is its inability to separate gravitational effects of the Greenland ice sheet from those of the underlying solid Earth. This causes errors from mismodeled PG to be a more serious problem than for other techniques.

[27] Our GRACE estimate of the total Greenland mass loss between the summers of 2002 and 2004, is 82 ± 28 km^3/yr. This should not be interpreted as the long-term mass imbalance, since the GRACE data were acquired over just two years. The results, though, are consistent with the three independent estimates described above (see Figure 3), even though the estimates correspond to different time spans. Note that all estimates indicate the ice sheet is losing mass.

[28] The uncertainty of our GRACE result is dominated about equally by the effects of GRACE measurement errors and errors in the PG estimate. The effects of measurement errors will decrease as more GRACE data are acquired. The effects of PG errors, though, will remain unchanged. PG errors are secular, so will not degrade attempts to infer such things as changes in the mass loss rates from one year to another - a difficult thing for altimeters to determine because of the density limitations. But the PG errors do define a limit to the accuracy in the secular trend achievable with GRACE alone. One way to alleviate this problem in the future is to combine GRACE with altimetry data and GAPS measurements [*Velicogna and Wahr*, 2002a].

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