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Remote Sensing Observations of Modern-Day Regional Ice Sheet Change

DISSertation

submitted in partial satisfaction of the requirements
for the degree of

DOCTOR OF PHILOSOPHY

in Earth System Science

by

Tyler Clark Sutterley

Dissertation Committee:
Professor Isabella Velicogna, Chair
Professor Eric Rignot
Professor François Primeau

2016
DEDICATION

To my parents and the 2007 UCSD varsity eight
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**ACCOLADES**

Outstanding Contributions to the Department of Earth System Science  
University of California, Irvine  
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Outstanding Contributions to the Department of Earth System Science  
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ABSTRACT OF THE DISSERTATION

Remote Sensing Observations of Modern-Day Regional Ice Sheet Change

By

Tyler Clark Sutterley

Doctor of Philosophy in Earth System Science

University of California, Irvine, 2016

Professor Isabella Velicogna, Chair

The Earth’s great ice sheets are losing mass at accelerating levels, rising global sea levels and posing a significant problem to society. The ice sheets contain enough water to raise sea level by 65 meters, and are the largest reservoirs of freshwater on the planet. Measurements of current ice sheet mass change are important in order to assess their current contribution to sea level rise, and to constrain future projections. There are three general approaches for measuring the current mass balance of ice sheets: the gravimetric method using time-variable gravity measurements, the altimetric method combining surface elevation change measurements with estimates of the density change, and the mass budget method combining rates of mass input from snow and rain with rates of mass output from meltwater runoff, ice discharge and other processes. In this dissertation, we use multiple independent measurements to assess the current uncertainties in mass balance efforts, and to create new estimates of current ice sheet mass change. We investigate key regions of Antarctica, where changes in the ice sheet velocity structure have led to accelerating mass losses. We compile new assessments of the mass change of the Greenland ice sheet, where increased rates of surface runoff and losses from ice sheet dynamics have dramatically shifted the mass balance regime. The work helps constrain estimation errors from GRACE, provides new constraints to ice sheet and glacial isostatic adjustment models, and helps improve our general understanding of the mechanisms driving current ice sheet mass change.
Chapter 1

Introduction

Glaciers and ice sheets are formed by the longterm accumulation of snow, which is compressed into ice over decades and centuries [Cuffey and Paterson, 2010]. The two ice sheets currently on the planet are the Greenland ice sheet (GrIS) and the Antarctic ice sheet (AIS). Antarctica is further divided into the West Antarctic ice sheet (WAIS), the East Antarctic ice sheet (EAIS), and the ice sheet of the Antarctic Peninsula (APIS). Other large ice sheets have been present in the Northern Hemisphere during the glacial periods of the last three million years [Allison et al., 2009]. At the last glacial maximum, the presence of currently extinct ice sheets and an expanded Antarctic ice sheet stored enough water to lower global sea levels between 120 and 135 meters [Alley, 2005, Clark and Mix, 2002]. The records of the past deglaciation periods provide insight to ice sheet behavior and potential sea level rise during periods of global climate change [Alley, 2005, Allison et al., 2009]. The rate of sea level rise during the last deglaciation period averaged approximately 10 mm/yr, but the rate peaked as high as 50 mm/yr during two large meltwater pulses [Alley, 2005]. The Greenland and Antarctic ice sheets currently store approximately 65 meters of sea level equivalent on land [Alley et al., 2007, Alley, 2005, Allison et al., 2009]. Due to their vast size, any small perturbation to either modern-day ice sheet can have a meaningful impact on global sea level [Zwally, 1989]. The current rate of sea level rise has increased since the 1970’s, which is partly due to ice sheet mass loss [Church et al., 2011]. For the period 1993–2008, the global
mean sea level rose at a rate of 3.22 ± 0.41 mm/year with 0.74 ± 0.26 mm/year coming from the ice sheets [Church et al., 2011]. Ice sheet retreat and collapse are major weak points in current predictions and reconstructions of sea level, which highlights the need for precise observations of current ice sheet change [Dutton et al., 2015].

Ice sheets gain mass from snowfall and precipitation, and lose mass by melt, ice discharge and other processes. The ice sheet mass balance is the quantified difference between the mass input by accumulation, and the mass output by ablation [Alley et al., 2007, Alley, 2005]. The mass balance of the ice sheets is determined by three general approaches: the gravimetric method, the mass budget method, and the altimetric method [Allison et al., 2009]. The gravimetric method calculates changes in ice sheet mass at broad spatial scales by measuring temporal changes in the Earth’s gravitational field [Swenson and Wahr, 2002, Velicogna and Wahr, 2006a,b]. The primary source of time-variable gravity measurements is from the Gravity Recovery and Climate Experiment (GRACE) [Tapley, 2004]. GRACE measures the total mass variation of the planet month-by-month. Isolating the mass change of ice sheets requires the removal of all other Earth system components from the GRACE signal [Velicogna and Wahr, 2006a, Wahr, 2015]. The mass budget method (MBM) compares the mass input by snow accumulation with the mass output by surface ablation and ice dynamics [Rignot et al., 2008a, 2011b]. Rates of ice discharge for an outlet glacier are commonly determined through a combination of surface velocity measurements from Synthetic Aperture Radar (SAR) with measurements of ice sheet thickness at a flux gate, typically at the grounding line [Rignot et al., 2008a]. The grounding line of a glacier is the critical point where the flowing ice is lifted off its bed and begins to float [Rignot et al., 2014]. The altimetric method calculates the total change in volume of the ice sheet by repeat measurements of the ice sheet surface elevation [Wingham et al., 2006, Zwally et al., 2005]. Calculating ice sheet mass change from surface elevation measurements requires information of the processes driving the volume change in order to get the effective density of the change [Pritchard et al., 2010, Shepherd et al., 2012]. Most of the surface layer of the ice sheet is composed of a form
of compacted snow known as firn. The density of the total column of snow, firn and ice can vary in time as low-density snow accumulates or melts, firn is compacted by the overlying snow, or several other surface processes [Alley et al., 2007, Pritchard et al., 2010].

Surface mass balance (SMB) and ice dynamics (D) are the two major terms in the ice sheet mass budget (see Equation 1.1). Surface mass balance is the quantified difference between mass inputs from the precipitation of snow and rain (P), and mass losses by sublimation (SU), runoff (RU), and wind scour (WS) [Lenaerts et al., 2012, van den Broeke et al., 2009]. Runoff is the portion of total snowmelt not retained or refrozen within the ice sheet. Wind scour is the erosion and sublimation of wind-blown snow from the ice sheet surface [Das et al., 2013]. Atmospheric temperatures influence the amounts of snowfall and snowmelt, as well as the partition between rain and snow. Ice dynamics affect the rates of ice discharge to the oceans, and transport of ice from the accumulation zones into the ablation zones [Rignot, 2006, Rignot and Kanagaratnam, 2006]. Different factors can influence the regional ice sheet dynamics, such as changes in the integrity of the buttressing ice shelves or changes in basal friction by variations in meltwater [Rignot and Thomas, 2002].

$$MB = \int (SMB - D) \, dt = \int ((P - SU - RU - WS) - D) \, dt$$ (1.1)

The Greenland ice sheet is currently losing mass at an accelerating rate due to a combination of surface mass balance changes, and increased rates of ice discharge into the ocean [Rignot et al., 2011b, van den Broeke et al., 2009]. Major outlet glaciers along the southeast coast of Greenland have been losing mass for over a decade [Rignot and Kanagaratnam, 2006]. The extent of ice mass loss has spread to the northwest coast and inland [Khan et al., 2010, Poinar et al., 2015]. In July of 2012, Greenland experienced an extreme melt event covering over 98% of the ice sheet surface [Nghiem et al., 2012]. Production of meltwater runoff has increased significantly since the late 1990’s, and is a major contributor to the acceleration in ice sheet mass loss [van den Broeke et al., 2009]. However, as most Greenland ice sheet
mass balance records are short, the separation between accelerating mass loss signals and longterm climate variability is currently limited [Wouters et al., 2013].

Antarctica is presently losing mass largely due to changes in the velocity structure of the ice sheet [Rignot et al., 2011b, Velicogna, 2009]. The mass losses are most concentrated on the Antarctic Peninsula and in West Antarctica [Rignot et al., 2008a]. Some major tributary glaciers draining into the Weddell Sea from the Antarctic Peninsula accelerated 2–8 times their previous flow rates after the 2002 collapse of the Larsen B ice shelf [Rignot et al., 2004]. These glaciers continued flowing at accelerated rates years after the collapse [Rignot et al., 2008a]. Pine Island and Thwaites glaciers of the Amundsen Sea Embayment (ASE) in West Antarctica have experienced significant increases in surface velocity, surface thinning, and grounding line retreat since the 1990’s [Rignot et al., 2002, 2014, Pritchard et al., 2009]. The smaller Smith, Pope, Haynes and Kohler glaciers of the ASE are also draining at increased rates compared to the 1970’s [Mouginot et al., 2014]. The change in ASE dynamics likely stems from the advection of warm Circumpolar Deep Water, which has weakened or disintegrated the buttressing peripheral ice shelves [Jacobs et al., 2011].

1.1 Time-Variable Gravity for Ice Sheet Mass Balance

The Earth’s gravitational field varies in time as masses on and within the Earth move and are exchanged between components of the Earth system∗ [Wahr, 2015]. As a result, gravity measurements can give insight on how the Earth’s mass redistributes over time [Wahr et al., 1998, Wahr, 2015]. The basis of using gravity for measuring mass stems from Newton’s Law of Universal Gravitation, which relates the force between all particles and bodies of matter. Newton’s first law states that if the Earth and an orbiting particle of mass \(m\) are separated

∗The science of measuring the shape, rotation and gravitational field of the Earth is known as geodesy. For detailed derivations of the geodesy equations in this chapter as well as more in-depth discussions of physical geodesy, please see Kaula [1966], Hoffman-Wellenhof and Moritz [2005] and Schubert [2015].
by a distance $r$, then the gravitational force $F$ between the Earth and the particle is:

$$F = G \frac{mM}{r^2}$$

(1.2)

where $G$ is the universal gravitational constant ($6.674 \times 10^{-11}$ m$^3$ kg$^{-1}$ s$^{-2}$) and $M$ is the mass of the Earth. The acceleration of the particle due to the gravitational attraction of the Earth can be calculated when Equation 1.2 is combined with Newton’s second law of motion ($F = ma$) [Kaula, 1966]. The gravitational potential of the particle, $V$, is a function of its position relative to the Earth [Jekeli, 2015].

$$V = \frac{GM}{r}$$

(1.3)

As the spatial distribution of the Earth’s mass is not uniform, the magnitude of the gravitational force varies at points on and above the Earth’s surface. The gravitational potential of the particle outside the body of the planet will be satisfied by Laplace’s equation ($\nabla^2 V = 0$) [Hoffman-Wellenhof and Moritz, 2005]. The traditional solution of Laplace’s equation for the Earth’s gravitational field is the set of spherical harmonics [Kaula, 1966, Hoffman-Wellenhof and Moritz, 2005]. Equation 1.4 shows the gravitational potential expressed as a series of spherical harmonics:

$$V(r, \theta, \phi) = \frac{GM}{r} \left\{ 1 + \sum_{l=1}^{\infty} \sum_{m=0}^{l} \left( \frac{a}{r} \right)^l \tilde{P}_{lm}(\cos \theta) \left[ \tilde{C}_{lm} \cos m\phi + \tilde{S}_{lm} \sin m\phi \right] \right\}$$

(1.4)

with $a$ being the average radius of the Earth ($\approx 6371$ km), $r$ the geocentric radial coordinate, $\theta$ the colatitude, $\phi$ the longitude, $\tilde{P}_{lm}(\cos \theta)$ the fully-normalized associated Legendre polynomials of degree $l$ and order $m$, and $\tilde{C}_{lm}$ and $\tilde{S}_{lm}$ the dimensionless spherical harmonic Stokes coefficients [Hoffman-Wellenhof and Moritz, 2005, Wahr, 2015]. The spherical harmonics are orthogonal to each other for different $l$ and $m$, which means that the dot product of any two distinct spherical harmonics is zero [Hoffman-Wellenhof and Moritz, 2005]. Fully-
Figure 1.1: (Left) Zonal Harmonic of degree 6 ($m = 0$). (Center) Sectorial Harmonic of degree and order 9 ($l = m$). (Right) Tesseral Harmonic of degree 16 and order 9. From the GFZ International Centre for Global Earth Models (ICGEM) Visualization Service.

Normalized spherical harmonics are defined such that the average square over the sphere for any degree, $l$, and order, $m$, is equal to one [Hoffman-Wellenhof and Moritz, 2005]. There are three distinct types of spherical harmonics: Zonal, where $m = 0$; Sectorial, where $m > 0$ and $l = m$; and Tesseral, where $m > 0$ and $l \neq m$. Figure 1.1 shows the different types of harmonics and demonstrates the orthogonal relationship between them.

The instantaneous shape of the Earth’s gravitational field can be described in terms of an equipotential surface, i.e. a surface of constant potential energy. An important equipotential surface for physical geodesy is the geoid, which would coincide with the sea surface if the oceans were at rest [Hoffman-Wellenhof and Moritz, 2005, Wahr et al., 1998]. The distance between the geoid and an Earth reference ellipsoid is the geoid height (or the geoidal undu-

Figure 1.2: Relationship between ellipsoid height, geoid height and topographic height. Modification of Figure 1.2 from National Research Council [2010]
lation for locations where the geoid is below the reference ellipsoid) [Hoffman-Wellenhof and Moritz, 2005]. The relationships between the Earth’s reference ellipsoid, mean sea surface height, geoid height, and surface topography are shown in Figure 1.2.

From Chao and Gross [1987], the geoid height, $N$, above the Earth’s mean spherical surface calculated from a summation of spherical harmonics is as follows:

$$N(\theta, \phi) = a \sum_{l=1}^{\infty} \sum_{m=0}^{l} \tilde{P}_{lm}(\cos \theta) \left[ \tilde{C}_{lm} \cos m\phi + \tilde{S}_{lm} \sin m\phi \right]$$  \hspace{1cm} (1.5)

For time-dependent variations in the Earth’s gravitational field, the change in the height of the geoid, $\Delta N$, is represented as changes in the Stokes coefficients, $\Delta \tilde{C}_{lm}$ and $\Delta \tilde{S}_{lm}$. Following Wahr et al. [1998], if $\Delta \rho(r, \theta, \phi)$ is the density redistribution driving the variation in gravity, then the coefficients $\Delta \tilde{C}_{lm}$ and $\Delta \tilde{S}_{lm}$ can be calculated by:

$$\begin{bmatrix} \Delta \tilde{C}_{lm} \\ \Delta \tilde{S}_{lm} \end{bmatrix} = \frac{3}{4\pi \rho_{ave} (2l + 1)} \int \Delta \rho(r, \theta, \phi) \left( \frac{r}{a} \right)^{l+2} \tilde{P}_{lm}(\cos \theta) \begin{bmatrix} \cos m\phi \\ \sin m\phi \end{bmatrix} \sin \theta \, d\theta \, d\phi \, dr \hspace{1cm} (1.6)$$

where $\rho_{ave}$ is the average density of the Earth ($\approx 5517 \text{ kg m}^{-3}$). The radial component of the density change cannot be uniquely determined using satellite geopotential observations alone [Wahr, 2015]. For fluctuations in water storage and transport (including mass variations of glaciers and ice sheets), much of the temporal variability in the Earth’s gravitation field can be assumed to be concentrated within a thin layer near the Earth’s surface. The thickness of this surface layer, $H$, is small compared to the average radius of the Earth such that $(l_{max} + 2)H/a \ll 1$, and $(r/a)^{l_{max}+2} \approx 1$ [Wahr et al., 1998]. The integral of the density change, $\Delta \rho(r, \theta, \phi)$, through the surface layer gives the change in surface mass density, $\Delta \sigma(\theta, \phi)$.

The change in surface mass density can be uniquely calculated from the total gravitational change with the surface layer assumption [Wahr et al., 1998, Wahr, 2015].

However, mass changes solely within the thin layer cannot fully explain the total gravitational
signal at a given time [Wahr et al., 1998]. Any change in the distribution of the Earth’s surface mass will induce an elastic response at depth [Wahr et al., 1998, Wahr, 2015]. The magnitude of the geopotential change from the induced solid Earth anomalies is related to the initial surface mass change by the series of degree-dependent Load Love numbers, \( k_l \) [Farrell, 1972, Wahr et al., 1998, Wahr, 2015].

\[
\begin{align*}
\begin{bmatrix}
\Delta \tilde{C}_{lm}^{\text{solid Earth}} \\
\Delta \tilde{S}_{lm}^{\text{solid Earth}}
\end{bmatrix}
&= k_l
\begin{bmatrix}
\Delta \tilde{C}_{lm}^{\text{surface mass}} \\
\Delta \tilde{S}_{lm}^{\text{surface mass}}
\end{bmatrix}
\end{align*}
\]  
(1.7)

The summation of the surface mass and solid Earth components provides an estimate of the total geopotential change caused by the redistribution of surface mass after allowing for the Earth’s elastic response [Wahr et al., 1998, Wahr, 2015].

\[
\begin{align*}
\begin{bmatrix}
\Delta \tilde{C}_{lm} \\
\Delta \tilde{S}_{lm}
\end{bmatrix}
&= \begin{bmatrix}
\Delta \tilde{C}_{lm}^{\text{solid Earth}} \\
\Delta \tilde{S}_{lm}^{\text{solid Earth}}
\end{bmatrix}
+ \begin{bmatrix}
\Delta \tilde{C}_{lm}^{\text{surface mass}} \\
\Delta \tilde{S}_{lm}^{\text{surface mass}}
\end{bmatrix}
= (k_l + 1)
\begin{bmatrix}
\Delta \tilde{C}_{lm}^{\text{surface mass}} \\
\Delta \tilde{S}_{lm}^{\text{surface mass}}
\end{bmatrix}
\end{align*}
\]  
(1.8)

Compensating for the additional elastic contribution leads to the complete equation relating the Stokes coefficients to the surface mass density change, \( \Delta \sigma(\theta, \phi) \) [Wahr et al., 1998]:

\[
\begin{align*}
\begin{bmatrix}
\Delta \tilde{C}_{lm} \\
\Delta \tilde{S}_{lm}
\end{bmatrix}
&= \frac{3}{4\pi a \rho_{\text{ave}} (2l + 1)} \int \Delta \sigma(\theta, \phi) \tilde{P}_l(\cos \theta)
\begin{bmatrix}
\cos m\phi \\
\sin m\phi
\end{bmatrix}
\sin \theta \ d\theta \ d\phi
\end{align*}
\]  
(1.9)

Following Wahr et al. [1998], the inverse equation relating a change in surface density in terms of a thin layer of water equivalent thickness at colatitude \( \theta \) and longitude \( \phi \) to the Stokes coefficients, \( \Delta \tilde{C}_{lm} \) and \( \Delta \tilde{S}_{lm} \), is given below:

\[
\Delta \sigma(\theta, \phi) = \frac{a \rho_{\text{ave}}}{3 \rho_{\text{H2O}}}
\sum_{l=0}^{\infty} \sum_{m=0}^{l} \frac{2l + 1}{1 + k_l} \tilde{P}_l(\cos \theta)[\Delta \tilde{C}_{lm} \cos m\phi + \Delta \tilde{S}_{lm} \sin m\phi]
\]  
(1.10)
1.1.1 GRACE Satellite Gravimetry

The Gravity Recovery and Climate Experiment (GRACE) was launched in March 2002 with the goal to accurately map the static and time-variable components of the Earth’s gravitational field [Tapley et al., 2004]. The GRACE mission is a joint project between the National Aeronautics and Space Administration (NASA) and the German Aerospace Center (Deutsche Forschungsanstalt für Luft und Raumfahrt, DLR). A follow-on mission (GRACE-FO), which is to continue the legacy of the original GRACE mission, is planned for launch in the fall of 2017. GRACE is comprised of two identical satellites which act together to form a long baseline gravimeter. Variations in the Earth’s gravitational field are measured by tracking changes in the distance between the two satellites with a microwave ranging instrument [Bettadpur, 2012c]. The ranging instrument records when a change in gravitational acceleration results in a velocity change of one of the satellites.

GRACE data is provided by three major processing centers: the Center for Space Research at the University of Texas (CSR), the Jet Propulsion Laboratory (JPL) and the German Research Centre for Geosciences (GeoForschungsZentrum, GFZ). The GRACE Level-1B data consists of all of the necessary measurements to compute solutions of the Earth’s gravity field, such as the satellite-to-satellite distances, non-gravitational accelerations, and the GPS positions of each satellite [Wahr, 2015]. The Level-2 GRACE data products consist of monthly spherical harmonic gravity solutions, $\tilde{C}_{lm}(t)$ and $\tilde{S}_{lm}(t)$, supplied to a set maximum degree, $l_{\text{max}}$, and order, $m_{\text{max}}$. The GRACE spherical harmonic solutions provided by the processing centers are based on the same Level-1 data, but use different background gravity models and processing strategies [Bettadpur, 2012c]. The maximum spherical harmonic degree, $l_{\text{max}}$, gives an indication of the shortest resolvable half-wavelength, $\psi_{\text{min}}$, and thereby the maximum spatial resolution of the GRACE data [Barthelmes, 2009].

$$\psi_{\text{min}}(l_{\text{max}}) = \frac{\pi \cdot a}{l_{\text{max}}} \approx \frac{20000 \text{ km}}{l_{\text{max}}} \quad (1.11)$$
Before converting into spatial maps of surface mass density change (see Equation 1.10), the GRACE spherical harmonics need to be processed to reduce the contributions from potential error sources [Wahr et al., 1998, Wahr, 2015]. Particular correlations between the GRACE harmonics are evident in the spatial domain as long north/south “stripes”. Numerous filters have been developed to remove these correlated errors from the Stokes coefficients [Swenson and Wahr, 2006]. The gravity fields are less accurate over short wavelengths due to the presence of random errors that increase rapidly with spherical harmonic degree [Swenson and Wahr, 2002]. The truncation of the spherical harmonics series to degree \( l_{\text{max}} \) also results in spurious ringing artifacts from Gibbs phenomenon. A common processing step to address these two issues is to smooth the spherical harmonic solutions using a Gaussian function [Jekeli, 1981, Wahr et al., 1998]. The amplitude of the processed data will be affected by the post-processing techniques applied [Swenson and Wahr, 2002]. Recovering an unbiased mass estimate from the GRACE data requires that the amplitude of the localized geophysical signal be restored [Velicogna and Wahr, 2006a, Landerer and Swenson, 2012].

Various processes throughout the Earth system can drive redistributions of mass resolvable by GRACE. Some of these processes include solid Earth and oceanic tides, glacial isostatic adjustment (GIA), atmospheric and oceanic transport, earthquakes and tectonic activity, sea level rise, and mass variability in the cryosphere and terrestrial hydrosphere [National Research Council, 1997]. These mechanisms can differ widely in terms of both spatial and temporal scales [National Research Council, 1997]. GRACE is sensitive to mass variations over broad spatial scales. As mass transport mechanisms can occur concurrently for a given region, the total time-dependent geopotential from GRACE can relate to several different time-varying components [Wahr, 2015]. In order to isolate the mass variation of a single component within the Earth system, all other time-variable components need to be removed from the GRACE signal [Velicogna and Wahr, 2006a, Wahr, 2015]. For ice sheet mass balance estimates from GRACE, the mass redistribution from Glacial Isostatic Adjustment is a key contaminating signal and source of uncertainty [Velicogna and Wahr, 2006a,b].
1.1.2 Glacial Isostatic Adjustment

During the last glacial period, the Greenland and Antarctic ice sheets carried much more ice volume, and additional ice sheets existed in the Northern Hemisphere. The increased ice loads induced viscoelastic deformation of the solid Earth and the flow of mantle material towards the equator [Farrell, 1972, A et al., 2013]. Areas depressed by the larger glacial period ice loads are gradually uplifting in the wake of the deglaciation as the solid Earth relaxes and mantle material shifts poleward [A et al., 2013]. This ongoing mass redistribution of the solid Earth is known as Glacial Isostatic Adjustment (GIA).

Glacial Isostatic Adjustment is evident as present-day changes in surface elevation and relative sea level, and secular trends in the Earth’s rotation and gravity field [Han and Wahr, 1997]. The magnitude of the GIA signal depends on the thickness of the Earth’s lithosphere, the viscosity structure of the Earth’s mantle, and the history of the deglaciation [Tamisiea, 2011]. For modern regional GIA models, constraints on these terms come from glacial-geological datasets, such as bedrock uplift rates from GPS, well-dated terminal morainies, and ice core accumulation records [Ivins et al., 2013]. The magnitude of the GIA correction for GRACE analyses of ice sheet mass change also depends upon the averaging area and the post-processing techniques applied [Tamisiea, 2011]. GIA uncertainty for a given deglaciation model can be estimated by varying the rheological parameters under scenarios meeting the model constraints [Ivins et al., 2013].

1.1.3 Mascon Approach for Regional Mass Balance

As GRACE measures the total mass variation over the entire planet, the mass variation of the ice sheets needs to be isolated from the global spherical harmonic solutions to assess their mass balance. One particular technique is to use mascons, or mass concentrations, to calculate if a region is in a state of mass surplus or mass deficit at a given time compared to
an initial state [Rowlands et al., 2010, Tiwari et al., 2009]. In an ideal case, the initial state for region \( k \) is a spatially discrete, uniformly distributed layer of equivalent water height at colatitudes \( \theta \) and longitudes \( \phi \) [Rowlands et al., 2010]:

\[
H_k(\theta, \phi) = \begin{cases} 
1 \text{ cm w.e.} & \text{if } (\theta, \phi) \text{ is in region } k \\
0 & \text{if } (\theta, \phi) \text{ is not in region } k
\end{cases}
\] (1.12)

GRACE measurements over the region at any given time would be a scalar multiple of this uniform layer. However, the GRACE harmonic solutions are truncated and typically smoothed, meaning that the sharp 0-to-1 transitions along regional boundaries cannot be resolved at the GRACE resolution [Wahr et al., 1998]. The mascon initial states can instead be represented as sets of truncated spherical harmonics processed in the same manner as the GRACE data [Jacob et al., 2012a]. In this case, each mascon initial state is a smoothed function with a total mass equal to the idealized case. If \( k \) is the \( k^{th} \) mascon out of \( N \) mascons, the relationship between the vectorized forms of the mascon spherical harmonics, \( C_{lm}^{\text{mascon}, k} \) and \( S_{lm}^{\text{mascon}, k} \), and the GRACE spherical harmonics at time \( t \), \( \Delta C_{lm}(t) \) and \( \Delta S_{lm}(t) \), is:

\[
\begin{bmatrix}
\Delta C_{lm}(t) \\
\Delta S_{lm}(t)
\end{bmatrix} = \begin{bmatrix}
C_{1m}^{\text{mascon}, 1} & \cdots & C_{1m}^{\text{mascon}, N} \\
S_{1m}^{\text{mascon}, 1} & \cdots & S_{1m}^{\text{mascon}, N}
\end{bmatrix} \begin{bmatrix}
\beta_1(t) \\
\vdots \\
\beta_N(t)
\end{bmatrix} + \begin{bmatrix}
\delta C_{lm}(t) \\
\delta S_{lm}(t)
\end{bmatrix}
\] (1.13)

where \( \delta C_{lm}(t) \) and \( \delta S_{lm}(t) \) represent the uncertainties in the harmonics at time \( t \) [Jacob et al., 2012a]. To solve Equation 1.13, the mascon spherical harmonics are simultaneously least squares fit to the GRACE spherical harmonic solutions [Tiwari et al., 2009]. The maximum likelihood solution for mascon \( k \) at time \( t \), \( \hat{\beta}_k(t) \), is the scale factor between the initial mascon harmonics and the GRACE solutions. The scale factors for each mascon are converted into mass terms by multiplying by the mascon areas, \( A_k \) [Jacob et al., 2012b].

\[
M_k(t)[g] = \left( \hat{\beta}_k(t) \ast 1[\text{cm w.e.}] \right) \ast A_k[\text{cm}^2] \ast \rho_{H_2O} \left[ \frac{g}{\text{cm}^3} \right]
\] (1.14)
Ideally, the final solution for the recovered mascon mass, $M_k(t)$, is equal to the true spatial average across the mascon [Jacob et al., 2012a]. Misfits in the regression or malformed initial states can lead to the leakage of GRACE signal in-between mascons or out of the system. The least squares mascon technique assumes that the GRACE signal can be well represented as scalar multiples of the mascons at any given time. As the initial mascon parameters are designed with uniform mass distributions, the GRACE anomalies over each mascon must also be uniform to limit statistical misfit [Jacob et al., 2012a]. The simplest solution to this problem is to decrease the average size of each mascon to insure that any resolvable GRACE anomaly will be uniform [Jacob et al., 2012b]. However, smaller mascons require more information at the less-determined higher degree and order harmonics. The mascon size and shape should balance the uniform distribution requirement of the technique with the resolution and error limitations of the GRACE data. The mascon sensitivity kernel, $A_k(\theta, \phi)$, is a function which describes the spatial sampling of the recovered mascon time series, $M_k(t)$ [Jacob et al., 2012a,b]. Optimizing the sensitivity kernels can help determine the best set of mascon parameters to sample a given region [Jacob et al., 2012b].

1.2 Lidar Mapping of Ice Sheet Surface Height Change

1.2.1 ICESat-1 Laser Altimetry

NASA’s Ice, Cloud and land Elevation Satellite (ICESat-1) mission was launched in January 13, 2003 into a near-circular 600 km altitude orbit [Schenk and Csatho, 2012]. The main instrument on ICESat-1 was the Geoscience Laser Altimeter System (GLAS), which consisted of three individual and identical 1064nm Nd-YAG lasers (L1, L2, and L3). After the premature failure of the L1 laser, the ICESat-1 mission was adjusted to acquire data in discrete campaigns. This decision limited the total number of surface elevation measurements compared to the initial design, but extended the overall mission lifetime to meet the
Table 1.1: ICESat-1 Laser Campaigns and Date Ranges

<table>
<thead>
<tr>
<th>Campaign</th>
<th>Start Date</th>
<th>End Date</th>
<th>Campaign</th>
<th>Start Date</th>
<th>End Date</th>
</tr>
</thead>
<tbody>
<tr>
<td>L1A</td>
<td>02/20/2003</td>
<td>03/21/2003</td>
<td>L3F</td>
<td>05/24/2006</td>
<td>06/26/2006</td>
</tr>
<tr>
<td>L2C</td>
<td>05/18/2004</td>
<td>06/21/2004</td>
<td>L3J</td>
<td>02/17/2008</td>
<td>03/21/2008</td>
</tr>
<tr>
<td>L3E</td>
<td>02/22/2006</td>
<td>03/28/2006</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

ice sheet elevation change detection objectives [Abdalati et al., 2010]. The names and date ranges of the ICESat campaigns are listed in Table 1.1. Overall, the mission fired nearly two billion shots before the failure of its final laser on October 11, 2009 [Schenk and Csatho, 2012]. The follow-on to the ICESat mission (ICESat-2) is currently planned for launch in 2018 [Abdalati et al., 2010].

The GLAS Science Computing Facility at NASA/Goddard Space Flight Center (GSFC) produces 15 distinct data products for atmospheric, ocean surface and land surface analyses. The GLA12 Antarctic and Greenland ice sheet altimetry data product provides geolocated surface elevation measurements over the ice sheets, which are calculated by fitting a series of Gaussian peaks to each lidar waveform. The Level-2 ICESat elevation data products are given in reference to the TOPEX/Poseidon reference ellipsoid. ICESat data points are negatively affected when the beam is saturated from the atmosphere or forward scattered from the presence of clouds [Pritchard et al., 2009]. In the absence of clouds, the return waveform over ice surfaces should resemble a single peak Gaussian function [Zwally et al., 2002]. The quality of every shot can be verified through a series of processing steps using the auxiliary data products provided with the GLAS data [Howat et al., 2008, Pritchard et al., 2009, Smith et al., 2009, Sørensen et al., 2011]. A table of sample culling criteria to remove data points affected by scattering or saturation from the ICESat data is listed in Table 1.2.
The ICESat elevation measurements need to be corrected for Gaussian-Centroid offset for data releases before 634, and for atmospheric saturation effects when a correction is possible [Borsa et al., 2014, Zwally et al., 2012].

### 1.2.2 Airborne Lidar Mapping

The airborne Arctic Ice Mapping (AIM) and IceBridge missions have been providing precise topographic measurements of the ice sheets since 1991 [Abdalati et al., 2002]. The first AIM campaigns with extensive coverage in Greenland were in 1993 and 1994 [Krabill et al., 2002]. NASA started airborne Antarctic campaigns in 2002 to prepare for the launch of the first ICESat mission, and started Operation IceBridge (OIB) over both Greenland and Antarctica in 2009 to bridge the gap until the second ICESat mission [Studinger et al., 2010].

The Airborne Topographic Mapper (ATM), and the Land, Vegetation, and Ice Sensor (LVIS) are the two main laser ranging instruments of Operation IceBridge [Abdalati and Krabill, 1999, Blair et al., 1999]. ATM is a conical scanning lidar developed and used by the NASA Wallops Flight Facility (WFF) in Virginia. LVIS is a large-swath scanning lidar developed at NASA’s Goddard Space Flight Center (GSFC). Both ATM and LVIS elevation datasets are given in reference to the WGS-84 reference ellipsoid. The Level-2 ATM data consists of surface elevations calculated by fitting planar surfaces to the original ATM point clouds at approximately 40m spacing along track [Krabill, 2010]. The fit procedure accounts for variations in the ice sheet surface slope when calculating the surface elevation [Krabill, 2010].
The Level-2 LVIS data consists of three surface elevation measurements computed from the Level-1B waveforms: the highest reflecting surface, the lowest reflecting surface and the centroid of the waveform [Blair et al., 1999]. The highest and lowest reflecting surfaces should be equivalent over regions with smooth terrain where the LVIS return waveforms resemble single peak Gaussian curves [Blair and Hofton, 2010]. These two elevations can diverge over regions with rough or crevassed surfaces [Blair and Hofton, 2010]. For regions of rough terrain, the elevation surface from the waveform centroid should be used as a replacement. The centroid elevation measurement represents a surface of equivalent volume to the rough terrain if it were smoothed into a single surface [Blair and Hofton, 2010].

1.2.3 Least Squares Approach for Surface Elevation Change

Due to a combination of satellite control issues and orbital maneuvers with ICESat-1, some of the repeat ground tracks are up to a few hundred meters apart [Schenk and Csatho, 2012]. As a consequence, estimating the cross-track slope and elevation change between ground tracks is difficult as the two variables are not independent [Schenk and Csatho, 2012]. One particular technique is to simultaneously solve for both elevation change and ice sheet surface shape using multivariate least squares. The first step in this technique is to spatially discretize the elevation data into a series of surface patches. Over the scale of a 1 km by 1 km patch, the ice sheet elevation surface shape can be approximated by a smooth polynomial [Schenk and Csatho, 2012]. Assuming that the surface shape of a patch does not deform in time, the spatial and temporal components of the surface elevation can be separated for each surface patch $k$ as shown in Equation 1.15 [Schenk and Csatho, 2012].

\[ S_k(x, y, t) = S_k(x, y) + S_k(t) + \sigma_k \quad (1.15) \]

Under these assumptions, each laser altimetry data point is a function of its acquisition time and its spatial coordinates within the patch. All available laser altimetry data within
the surface patch can be used in an inverse model to solve for the parameters describing the spatial and temporal variation. Different spatial, $S_k(x, y)$, and temporal models, $S_k(t)$, are tested for each patch to find the best set of parameters reducing the regression variance. Any measurements affected by drifting snow, surface crevasses or low-lying clouds are iteratively culled after each regression fit. Assuming all laser shots and all patch fits are independent, the total error for a region, $\sigma_{region}$, is the summation in quadrature of the patch elevation uncertainties, $\sigma_k$, multiplied by the region area, $A_{region}$ [Schenk and Csatho, 2012].

$$\sigma_{region} = \frac{A_{region} \sqrt{\sum_{k=1}^{N} \sigma_k^2}}{\sqrt{N}}$$

(1.16)

**1.2.4 Firn Compaction**

The density of the ice sheet surface layer varies due to the compression of low-density snow into glacier ice. New snow is compacted by thermal and wind stresses, and firn is further compacted by the weight of the overlying snow and firn. Firn becomes glacier ice once the air-filled pores between the grains of snow completely close off, which occurs approximately at a density of 830 kg/m$^3$ [van den Broeke, 2006]. The equivalent ice thickness of a column of firn and snow is calculated using the height and density of the snow layer ($\rho_{snow}$, 100 − 550 kg/m$^3$), the height and density of the firn layer ($\rho_{firn}$, 550 − 830 kg/m$^3$) and the density of glacier ice ($\rho_{ice}$, 917 kg/m$^3$) [Ligtenberg et al., 2011]. The equation to calculate the equivalent ice thickness, $h_{ice}$, of a snow/firn column from snow depth, $h_{snow}$, and firn depth, $h_{firn}$, is expressed in Equation 1.17 [Ligtenberg et al., 2011].

$$h_{ice} = h_{snow} \left( \frac{\rho_{snow}}{\rho_{ice}} \right) + h_{firn} \left( \frac{\rho_{firn}}{\rho_{ice}} \right)$$

(1.17)

In the absence of significant melting, the rate of compaction depends on the temperature, snow accumulation rate, and wind speed [Ligtenberg et al., 2011, van den Broeke, 2006, ...]
In Antarctica, firn-layer thickness variations can be the same order of magnitude as the elevation changes [Helsen et al., 2008]. For glaciers in temperate regions, densification is rapid and the firn-layer thickness is shallow (40 – 60 m). For the cold and dry interiors of ice sheets, densification is slow and the firn-layer thickness often exceeds 100 m [van den Broeke, 2006]. The rate and spatial variability of firn compaction is particularly important for calculating ice sheet mass balance from repeat altimetry measurements of surface elevation.

1.3 Research Objectives and Outline of Dissertation

The overall science goal of this research is to improve our understanding of Greenland and Antarctic ice sheet mass balance, and the processes controlling current changes. This objective is addressed using a combination of satellite and airborne datasets together with outputs from regional climate models. The technical objectives for this research are as follows: 1) improve regional GRACE ice mass balance estimates by reducing Glacial Isostatic Adjustment (GIA) uncertainties, and developing improved post-processing methodologies, 2) identify regions where most changes occur and investigate the different processes causing the observed changes, 3) produce a longterm, cross-validated regional mass balance record of the major sector of mass loss in Antarctica, 4) generate improved Greenland elevation change estimates using satellite and airborne laser altimetry, and 5) help improve our understanding of processes driving current ice sheet mass change.

In Chapter 2, we test and evaluate different GIA corrections in Greenland in an effort to help constrain ice sheet mass balance estimates from GRACE. The research utilizes differences in magnitude, spatial patterns, and temporal variations of independent Greenland ice sheet datasets within the evaluation. For this research, we develop a method for deriving regional ice sheet mass balance from GRACE using a least squares mascon approach. This work was published as:
In Chapter 3, we determine the key regions which currently dictate Greenland and Antarctic ice sheet mass change, and use the regional ice sheet mass balance technique derived in the previous study to estimate the mass change in these regions. We focus primarily on the accelerations of ice mass loss over the GRACE time period. The statistical significance of ice mass trend and acceleration spatial patterns is also determined. Within the study, the regional GRACE time series are compared with outputs from regional climate models to determine the relative contribution of surface mass balance variations to the current mass change. This work was published as:


In Chapter 4, we focus on the Amundsen Sea Embayment, a key sector of West Antarctica outlined in the previous work as the major contributor to sea level from the ice sheet, and calculate the regional mass balance with four independent techniques. Within this research, we develop a method for combining satellite and aerial laser altimetry measurements to use as one of the four techniques. We find within this work that the Operation IceBridge campaign-style measurements are sufficient to extend the ICESat-derived time series of elevation change. This work was published as:

In Chapter 5, we derive a method for calculating rates of elevation change combining satellite and airborne lidar measurements to utilize the high spatial sampling of the airborne surveys. We use this method to investigate recent changes in surface elevation in Greenland for two case studies: 1) the recent thinning of Jakobshavn Isbræ, and 2) an assessment of surface mass balance outputs from regional climate models. Chapter 6 summarizes the major findings and implications of this dissertation. Chapters 2 to 4 are lightly modified versions of previously published papers printed with permission from the publishers.
Chapter 2

Evaluating Greenland Glacial Isostatic Adjustment corrections using GRACE, altimetry and surface mass balance data

As appears in:

Abstract

Glacial isostatic adjustment (GIA) represents a source of uncertainty for ice sheet mass balance estimates from the Gravity Recovery and Climate Experiment (GRACE) time-variable gravity measurements. We evaluate Greenland GIA corrections from Simpson et al. [2009], Wu et al. [2010], and A et al. [2013] by comparing the spatial patterns of GRACE-derived ice mass trends calculated using the three corrections with volume changes from ICESat (Ice, Cloud, and land Elevation Satellite) and OIB (Operation IceBridge) altimetry missions, and surface mass balance products from the Regional Atmospheric Climate Model (RACMO).
During the period September 2003–August 2011, GRACE ice mass changes obtained using the Simpson et al. [2009] and A et al. [2013] GIA corrections yield similar spatial patterns and amplitudes, and are consistent with altimetry observations and surface mass balance data. The two GRACE estimates agree within 2% on average over the entire ice sheet, and better than 15% in four subdivisions of Greenland. The GRACE estimate corrected using the Wu et al. [2010] GIA shows similar spatial patterns, but produces an average ice mass loss for the entire ice sheet that is 64–67 Gt/yr smaller. In Northeast Greenland, the recovered ice mass change is 46 to 49 Gt/yr (245–270%) more positive than those deduced from the other two corrections. By comparing the spatial and temporal variability of the GRACE estimates with trends of volume changes from altimetry and surface mass balance from RACMO, we show that the Wu et al. [2010] correction leads to a large mass increase in the Northeast that is inconsistent with independent observations.

2.1 Introduction

Over the past 20 years, the mass balance of the Greenland Ice Sheet (GrIS) has been increasingly negative, driven by changes in both surface mass balance (SMB), and ice dynamics [Rignot et al., 2011b, Khan et al., 2010, Pritchard et al., 2009, van den Broeke et al., 2009]. Mass losses from the GrIS are significant contributors to global sea level rise and freshwater fluxes to the North Atlantic [Rignot et al., 2011b, Bamber et al., 2012].

Time-variable gravity measurements from the Gravity Recovery and Climate Experiment (GRACE) provide a powerful tool for estimating monthly ice sheet mass balance [Tapley, 2004]. GRACE ice sheet mass balance estimates need to be corrected for the mass redistribution related to glacial isostatic adjustment (GIA), the Earth’s ongoing viscoelastic response to the redistribution of ice and water masses that have occurred following the last deglaciation. GIA is usually removed using a-priori corrections [Wu et al., 2010]. In Antarctica, the
GIA correction represents the largest source of uncertainty in the GRACE ice mass balance estimates [Ivins and James, 2005, Velicogna and Wahr, 2006a, 2013]. In Greenland, the GIA correction averaged over the entire GrIS is much smaller than in Antarctica, but there are key differences between the current published corrections. The uncertainty in the GIA correction impacts both the magnitude and associated accuracy of GRACE-based ice mass estimates, particularly at regional scale, where the GIA correction may locally represent a significant portion of the GRACE signal [Simpson et al., 2009, Wu et al., 2010, A et al., 2013]. Constraining GIA corrections at the regional level will help improve the overall accuracy of the GRACE ice mass estimates.

Uncertainties in GIA models come from a lack of constraints on global ice sheet history since the Last Glacial Maximum (LGM), and the Earth’s internal rheological structure [Ivins and James, 2005]. There are two main classes of GIA models: (1) global [Peltier, 2004] and (2) regional models [Ivins et al., 2013]. The regional models are constrained by local datasets, such as relative sea level, GPS measurements of crustal uplift, and geological records [Ivins and James, 2005, Mackintosh et al., 2011, Simpson et al., 2011, Spada et al., 2012]. Wu et al. [2010] derive an alternative global GIA estimate using an inverse method. Alternative regional GIA corrections have been calculated through a combination of ice mass balance estimates from satellite altimetry and gravimetry [Riva et al., 2009, Gunter et al., 2014]. On average over the entire GrIS, the mass change corresponding to the Wu et al. [2010] correction is 15 times larger than the other published GIA corrections, which yields an ice mass balance estimate much smaller than other published estimates. Prior GIA-evaluation studies compared GIA estimates to 3D ice sheet/bedrock models [Olaizola et al., 2012], compared mass rates between GRACE and altimetry data [Gunter et al., 2009], or evaluated the input data used to constrain the GIA derivation.

Here, we present a new methodology to evaluate GIA corrections over Greenland that compares the GRACE ice mass balance estimates with observations of surface mass balance
(SMB) from the Regional Atmospheric Climate Model (RACMO) [Ettema et al., 2010], and observations of ice volume change from ICESat and Operation IceBridge (OIB) laser altimetry data [Krabill et al., 2002, Schenk and Csatho, 2012, Zwally et al., 2002]. We evaluate the different GIA corrections by determining the level of compatibility of the GRACE estimates with the amplitude and spatial pattern of these independent observations. These three datasets are fundamentally different: GRACE measures the total mass change, RACMO reconstructs the ice sheet surface mass balance (a portion of the total mass balance that does not account for changes in ice discharge), and altimetry measures the ice volume change. Ice volume may be translated into mass change if the density at which elevation changes are taking place is known. In general, this density is not well known, which introduces large uncertainties [Zwally et al., 2005]. Here, we choose not to convert the volume changes into mass, and note that elevation changes cannot be directly compared with mass changes. Surface elevation measurements have the advantage of a low sensitivity to GIA and high sensitivity to SMB.

We compare the three datasets during the period September 2003–August 2011. Within the period, we consider two sub-periods, September 2003–September 2007 and September 2007–August 2011, in order to separate the different sources of mass variability and identify uncertainties. The sub-periods are chosen based on the availability and sampling of the altimetry dataset. We use our approach to evaluate three GIA corrections: Simpson et al. [2009], A et al. [2013], and Wu et al. [2010]. We investigate the temporal variability of each GIA-corrected GRACE time series at four regions in Greenland using a least squares mascon approach. Within this approach, we evaluate different kernel designs to best estimate the regional time series. In the following sections we discuss how the comparison is implemented, how we use the results to evaluate different GIA corrections, and conclude on the robustness of the corrections.
2.2 Data and Methods

We use 94 monthly GRACE Release-5 (RL05) gravity solutions provided by the Center for Space Research (CSR) at the University of Texas [Tapley et al., 2004, Bettadpur, 2012b] for the period September 2003 to August 2011. This is the longest period for which all the examined datasets are available. GRACE RL05 data use improved dealiasing products of the Earth’s non-tidal atmospheric and oceanic variations, updated background gravity models, and improved processing methods [Bettadpur, 2012b]. The dealiasing products remove non-tidal atmospheric and oceanic mass variability from the monthly GRACE products using outputs from the European Centre for Medium-Range Weather Forecasts (ECMWF) and the baroclinic Ocean Model for Circulation and Tides (OMCT) [Bettadpur, 2012b]. A key improvement with the new release (RL05) data products is the marked reduction of “striping” errors and spherical harmonic noise compared to the fourth release (RL04) [Bettadpur, 2012a]. Each CSR gravity field solution consists of spherical harmonic Stokes coefficients, $C_{lm}$ and $S_{lm}$ up to degree, $l$, and order, $m$, 60. $C_{20}$ coefficients derived from GRACE show anomalously large variability due to excessive noise and high sensitivity to tidal aliasing errors [Chen and Wilson, 2010]. We replace the GRACE-derived $C_{20}$ coefficients with monthly estimates from Satellite Laser Ranging (SLR) [Cheng and Tapley, 2004]. GRACE does not recover degree-1 coefficients, which are related to motion of the Earth’s geocenter [Swenson et al., 2008]. The omission of degree-1 can significantly degrade estimates of ice mass variability by excluding long-wavelength components of the Earth’s mass change, and by leaking far-field signals into the regional estimates [Velicogna, 2009]. We account for the variation of degree-1 using coefficients calculated from a combination of GRACE coefficients and ocean model outputs [Swenson et al., 2008].

We apply different GIA corrections: (1) Simpson et al. [2009], (2) A et al. [2013], and (3) Wu et al. [2010], to the GRACE data and obtain three different ice mass balance estimates.
For the remainder of the paper, we refer to the three corrections as SM09-GIA, AW13-GIA, and Wu10-GIA, respectively, and to the associated ice mass estimates (GRACE-GIA) as SM09, AW13, and Wu10. The SM09-GIA correction is a regional GIA model using a thermomechanical ice sheet model calibrated with relative sea level data and geological observations of ice sheet extent [Simpson et al., 2009]. The AW13-GIA correction is an update to the 2007 Paulson global GIA model [Paulson et al., 2007] using the ICE-5G deglaciation history, compressibility parameters from the Preliminary Reference Earth Model (PREM), and a layered approximation of the Peltier VM2 mantle viscosity profile [A et al., 2013]. The Wu10-GIA correction is generated using a global inversion method including terrestrial and space geodetic data (GPS, Satellite Laser Ranging, and Very Long Baseline Interferometry), and ocean bottom pressure output from the JPL ECCO model [Wu et al., 2010].

Greenland ice mass anomalies are calculated relative to the period September 2003 to August 2011. We account for the elastic deformation of the solid Earth induced by variations in mass loading using load Love numbers of gravitational potential, $k_l$, calculated by Wahr et al. [1995]. We simultaneously fit annual and semiannual signals, a linear trend and a constant to the Stokes coefficient time series. To reduce the random spherical harmonic error component, which increases as a function of decreasing wavelength, we smooth the GRACE data using a normalized version of Jekeli’s Gaussian averaging function with a 250 km radius [Jekeli, 1981, Wahr et al., 1998]. Finally, we generate evenly spaced latitude-longitude grids for the three GRACE-derived ice mass changes (Equation 1.10, Figure 2.1).

We use these maps to compare spatial patterns in the SM09, AW13, and Wu10 ice sheet mass changes. To evaluate how the three GIA corrections impact the ice mass balance estimates at a regional scale, we divide the ice sheet in four regions: Northwest (NW), Northeast (NE), Southwest (SW) and Southeast (SE) as shown in Figure 2.1 and Figure 2.2. For each region, we calculate the average time series using a least squares mascon approach [Tiwari et al., 2009, Jacob et al., 2012b]. Each of the four regions is composed of several small
mascons. Each mascon is a 3-degree diameter equal-area spherical cap with a mass equal to a uniformly distributed centimeter of water [Farrell, 1972]. For each mascon, we calculate a set of Stokes coefficients, which we smooth with a 250 km Gaussian function and convert into mass [Jacob et al., 2012b, Wahr et al., 1998]. We calculate the mass associated to each mascon by simultaneously fitting the mascon Stokes coefficients to the GRACE monthly coefficients corrected for the GIA correction (Equation 1.13) [Jacob et al., 2012b].

The error of the regional ice mass estimates is due to the leakage error, GRACE measurement error, GIA error and the statistical uncertainty of the fit. We evaluate these contributions as described below. For each region, we calculate the corresponding sensitivity kernel to evaluate how mass at a given point within the region contributes to total time series [Tiwari et al., 2009]. We find that if the GRACE mass anomalies are distributed uniformly over each mascon, the fit results will recover the total variability for each region. To reduce the leakage from the glaciers and ice caps of the Canadian Archipelago, we distribute additional mascons over this region as shown in Figure 2.2. We evaluate the error introduced by assuming a uniform mass distribution within each mascon by estimating the leakage error for a field of simulated, realistic ice mass change rates across the Greenland ice sheet based on mass balance estimates obtained using the mass budget method [Rignot et al., 2011b]. We include an additional “statistical” leakage term in our regional error budgets calculated by first removing the retrieved mascon estimates from the GRACE spherical harmonic fields, and then refitting our mascon fields [Tiwari et al., 2009]. We estimate the effects of measurement errors in the individual GRACE monthly fields by convolving the sensitivity kernel for each mascon with uncertainty estimates for the GRACE Stokes coefficients [Wahr et al., 2006]. Over the ice sheet, the leakage and GRACE error components are approximately 11 and 14 Gt/yr respectively. GIA error is calculated considering the different rheological parameters, such as lithospheric thickness and mantle viscosity provided with SM09 and AW13, and by using the estimate of the inversion uncertainty for Wu10. The summation of the resultant errors is shown in Table 2.1.
Repeat laser altimetry measures the change in ice sheet elevation at the scale of individual glaciers [Shepherd et al., 2012, Zwally et al., 2002]. We use rates of surface elevation change from the University of Buffalo’s Surface Elevation Reconstruction And Change detection (SERAC) project [Schenk and Csatho, 2012, Rezvanbehbahani, 2012]. SERAC determines surface elevation changes by reconstructing the temporal variation of polynomial surfaces, which are fit to the altimetry data from Pre-IceBridge ATM (Airborne Topographic Mapper), ICESat, and IceBridge ATM and LVIS (Land, Vegetation and Ice Sensor) datasets. Aerial laser altimetry data are integrated into the SERAC solutions in order to increase the spatial and temporal coverage of elevation measurements in the south of Greenland and at key outlet glaciers [Krabill et al., 2002, Schenk and Csatho, 2012, Rezvanbehbahani, 2012]. The ICESat data is from Release 531 of the GLAS/ICESat Antarctic and Greenland Ice Sheet Altimetry Data product (GLA12) computed by the GLAS Science Computing Facility at NASA/GSFC [Zwally et al., 2002]. Aerial altimetry datasets were acquired from the National Snow and Ice Data Center (NSIDC) and include the Pre-IceBridge and IceBridge ATM Level-2 Icessn Elevation, Slope, and Roughness products, and the IceBridge LVIS L2 Geolocated Surface Elevation product [Krabill and Thomas, 2010, Krabill, 2010, Blair and Hofton, 2010]. In our altimetry dataset, yearly rates of elevation change are calculated by differencing the surface elevation between Greenland balance years (September 1–August 31), which begin near the start of the accumulation season and finish near the end of the ablation season.

We apply an ICESat intermission bias correction (IBC) based on ocean elevation measurements to the ICESat elevation data [Urban and Schutz, 2005, Urban et al., 2012]. This correction eliminates the major error (trend) caused by the Gaussian-Centroid (G-C) processing issue [Borsa et al., 2014]. In addition, the procedure of least squares fitting analytical functions that we use to determine the local annual elevation changes from altimetry time series reduces the random component of the G-C error [Schenk and Csatho, 2012]. Thus, the impact of the G-C correction bias on the ICESat elevation data used in our analysis is negligible. Elevation changes are calculated in reference to the WGS-84 ellipsoid, and corrected
for saturation effects with the GLA12 correction product. The effects of crustal deformation from ocean tides and solid Earth tides on the elevation measurements are corrected using the GOT99.2 global tide model [Ray, 1999]. Elevation measurements are corrected for GIA-induced crustal uplift using a relation between the crustal uplift rates and GIA Stokes coefficients [Wahr et al., 2000]. Altimetry measurements are not very sensitive to GIA as the rate of crustal uplift from GIA is much smaller than the ice elevation change at most locations. In our comparison, we correct the ice sheet elevation measurements with each of the three GIA corrections to compare with the corresponding GRACE estimates.

We use monthly mean components of surface mass balance (SMB) calculated from a 1960 to 2012 climate simulation of the Regional Atmospheric Climate Model (RACMO2.1) from Utrecht University [Ettema et al., 2010]. Surface mass balance is the sum of mass accumulation (snow, rain and deposition) minus surface ablation (sublimation, runoff, and the erosion and sublimation of windblown snow) [Ettema et al., 2009]. In Greenland, surface mass balance represents approximately 50% of the total ice mass loss signal [van den Broeke et al., 2009]. RACMO is a high-resolution regional climate model (∼11 km) forced at the lateral boundaries and the sea surface by reanalysis datasets from the European Centre for Medium-Range Weather Forecasts (ERA-40 and ERA-Interim). RACMO employs a physical snow/ice surface model to calculate the time-variable surface albedo as a function of ice sheet properties, and to better represent processes affecting SMB, such as meltwater penetration and refreezing [Bougamont et al., 2005, Ettema et al., 2009, 2010]. The estimated uncertainty in the accumulation component of RACMO, evaluated from a comparison between model predictions and observations from ice cores and research stations, is approximately 9% [Ettema et al., 2009]. When combined with the uncertainty in runoff, the total uncertainty in ice sheet SMB increases to 17% [Howat et al., 2011].

Cumulative anomalies in SMB are calculated in reference to a period of assumed net balance, 1961 to 1990. During this 30-year reference period, the total ice mass of the ice sheet has
been assumed to be largely in balance [Rignot et al., 2008b]. Over the entire record, there is no indication of a longterm change in accumulation of snowfall [Howat and Eddy, 2011, van den Broeke et al., 2009].

Measurements by laser altimetry and RACMO outputs contain higher-spatial resolution information than the GRACE measurements. We process the altimetry and SMB fields in the same manner as the GRACE data by an expansion into spherical harmonics truncated to degree $l_{\text{max}}=60$, convolution with a 250km radius Gaussian smoothing kernel, and conversion back into spatial domain (Equations 1.9 and 1.10) [Sneeuw, 1994, Wahr et al., 1998]. To perform a comparison of the regional mass change through time, we apply the same mascon fit technique as for the GRACE data to the RACMO output products.

2.3 Results and Discussion

The GIA correction affects both the total magnitude and spatial variability of ice mass changes (Figures 2.1,2.4). However, the GIA signal is constant over the analyzed period. This means that errors in GIA will have the same impact on ice mass changes for the entire period and for the sub-periods. For the analyzed period, the ice mass balance of Greenland and the corresponding GIA correction are, respectively, $-256 \pm 21 \text{ Gt/yr}$ and $-3 \pm 12 \text{ Gt/yr}$ (1%) for SM09, $-253 \pm 23 \text{ Gt/yr}$ and $-6 \pm 5 \text{ Gt/yr}$ (2%) for AW13, and $-189 \pm 27 \text{ Gt/yr}$ and $-69 \pm 19 \text{ Gt/yr}$ (36%) for Wu10 (Table 2.1). At the regional scale, the ice mass estimates are more dependent on the GIA correction, especially in NE Greenland where the Wu10-GIA correction is the largest portion of the signal measured by GRACE (Table 2.1).

Over the entire analyzed period, SM09 and AW13 mass changes show consistent spatial patterns, with most of the mass loss concentrated in the SE and NW. When averaged over the entire ice sheet, the two estimates agree within 2% (3 Gt/yr), which is at the 95% confidence interval (Table 2.1). At the regional scale, AW13 and SM09 agree within the
associated errors (3–15%) for all regions and Wu10 agrees within the error budget for the SE, SW and NW regions. The spatial pattern of the Wu10 ice mass change is markedly different with a large mass increase in the NE and smaller coastal losses. When averaged over the entire ice sheet, the Wu10 ice mass change is approximately 25% smaller (64–67 Gt/yr) than the AW13 and SM09 estimates. In the NW, SE and SW, the Wu10 ice mass changes are 2 to 12 Gt/yr (4–17%) less negative that the AW13 ones and 2 to 7 Gt/yr (4–10%) less negative than the SM09 values. In the NE, the Wu10 values are 46 to 49 Gt/yr (245–270%) more positive than the AW13 and SM09 values respectively (Table 2.1).

Figure 2.3 shows spatial patterns of SM09 ice mass changes (Figure 2.3a–c), altimetry-derived elevation changes (Figure 2.3d–f) and SMB (Figure 2.3g–i). In 2003–2011, SM09 and altimetry indicate ice mass loss and thinning, respectively, concentrated in the SE and NW. In 2003–2007, the mass loss and thinning are stronger in the SE, and they spread to the NW in 2007–2011. The difference in amplitude cannot be full analyzed since the density at which elevation changes take place is not known.

As GRACE senses the total mass change, the GRACE signal contains information about both SMB and ice discharge. Inconsistencies between the GRACE and SMB spatial patterns and time series can be attributed to ice mass losses by ice discharge, errors in the GIA correction, errors in the SMB model and errors in the GRACE data. In the SE, GRACE (Figure 2.3a–c) displays consistently larger mass losses than the SMB estimate (Figure 2.3g–i), and GRACE trends are 58–66 Gt/yr (132–150%) larger than the SMB trend (Table 2.1). We attribute the difference in the SE to a strong ice discharge component, which is noted in other studies [Rignot and Kanagaratnam, 2006, van den Broeke et al., 2009]. Similarly, in the NW the GRACE estimates agree within error bars, but show much larger losses than SMB. The difference in the NW between SMB and GRACE for the time period 2003–2011 is 23 Gt/yr (50%), 28 Gt/yr (61%) and 16 Gt/yr (34%) for SM09, AW13 and Wu10 respectively. We attribute the difference to glacier velocity increases over the 2007–2011 period [Moon et al.,
Conversely in the SW, the SMB signal dominates the total ice mass change, and the different GRACE estimates and SMB agree within error bounds. This finding is consistent with the fact that most of the glaciers are land terminating, and glacial discharge has not changed significantly [Rignot and Kanagaratnam, 2006, Moon et al., 2012].

In the NE, we find the largest differences between the three GRACE estimates. This region is most sensitive to errors in GIA due to a lower ice mass change to GIA ratio. AW13 and SM09 show negative ice mass trends of -17 Gt/yr and -20 Gt/yr, respectively, whereas Wu10 shows a mass increase of +29 Gt/yr. The Wu10 differs from SMB by +54 Gt/yr, which is outside the SMB error bounds, while SM09 and AW13 are within 5 to 8 Gt/yr, which is within error bounds. If the GIA corrections are accurate, the agreement between SM09, AW13 and SMB suggests that the ice mass variability in this region could be largely explained by SMB. In fact, observations of ice discharge and ice flux in the NE have reported little ice dynamic change [van den Broeke et al., 2009, Moon et al., 2012]. Moon et al. [2012] report sub-threshold (i.e. low velocity or with erratic behavior) glacier velocity changes for the period 2000–2010, with the only exception of Adolf Hoel Gletscher that switched from lower velocity during 2000–2005 to higher velocity over 2005–2010. Sasgen et al. [2012] estimated a small mass gain (less than 5 Gt/yr) in the regional ice discharge component over 2002–2011. In order to explain the ice mass increase observed in Wu10, we would need either a larger change in ice discharge than previously observed with a significant decrease in ice velocities and in associated fluxes, or a positive anomaly in SMB much larger than the error bounds on SMB. The error in SMB is 17% for the entire ice sheet, however, the regional uncertainties may be larger [Ettema et al., 2009, van den Broeke et al., 2009, Howat et al., 2011, Vernon et al., 2013]. A recent study however estimated that the mean standard deviation between SMB models in the north is 17% [Vernon et al., 2013]. We conclude that it is unlikely that the mass gain estimated by Wu10 could be attained by SMB errors.

Measurements of surface elevation change provide an independent check of both GRACE
and SMB estimates. In Greenland, errors in the GIA correction have a lower impact on ice elevation changes compared to the impact on the GRACE estimates because the rate of GIA crustal uplift is much smaller than the ice elevation changes at most locations. While SMB and GRACE both estimate mass change, altimetry measures volume change. The spatial pattern of the changes observed by GRACE and altimetry should be similar; however, the magnitude of the GRACE and altimetry signals will differ depending on the density at which the change in surface elevation occurs, i.e. $0.3 \pm 0.2 \, \text{g/cm}^3$ for snow versus $0.917 \, \text{g/cm}^3$ for pure glacier ice. For a mass change involving accumulation, the surface elevation change will be highly magnified compared to the corresponding change in water height measured by GRACE ($2\text{–}18\times$). For a mass change involving the entire column of ice, such as from changes in ice dynamics, the two signals will be similar in magnitude or within 10%. In the NE, we see an increase in surface elevation in altimetry consistent in spatial pattern with SMB and SM09, but larger in amplitude than the SM09 GRACE mass change. This would suggest that the change is related to changes in accumulation rather than changes in the regional ice dynamics. The observed change in surface elevation is 50% smaller than the amplitude of the Wu10 GRACE signal. The Wu10 GRACE signal is inconsistent with both an accumulation signal and an ice dynamic signal based on this discussion.

The Wu10-GIA is not obtained using a standard GIA model. The authors use a kinematic approach to the simultaneous estimation problem of GIA and mass balance, which is very different than the methods used by SM09-GIA and AW13-GIA. We cannot explain why the Wu10 result is inconsistent with other observations. It may be the result of a sparse GPS network, issues with the original set of equations and assumptions used in the inversion, or reasons that would require further study to be fully clarified.

We examine a wide variety of mascon designs in the development of this method. Each design is tested using synthetic mass fields with realistic mass distributions constructed using surface mass balance outputs [Ettema et al., 2010] and the mass budget method [Rignot
et al., 2011b]. The spherical caps method for calculating regional time series with GRACE is a significant improvement over the near equal-area blocks method of Jacob et al. [2012b], and a moderate improvement over an unpublished method using four large mascons with scaling factors (Figure 2.5). The blocks method relies more on the noisier higher degree and order harmonics to resolve the block edges, and has greater levels of spatial ringing in the resultant kernels. Statistical misfit limits the effectiveness of the large mascon method in regions with lower signal-to-noise ratios, such as in NE Greenland, a focal point of this study. The three methods produce similar results for the entirety of Greenland, but considerably different results for the four regions (Figure 2.5). When analyzing subregions, the spherical caps method outperforms the other two methods in terms of both the resultant noise levels and signal recovery of the synthetic mass fields. When integrating entire regions or analyzing isolated signals, the methods are largely in agreement.

2.4 Conclusion

In this study, we compare three independent techniques for monitoring the Greenland ice sheet over the period September 2003–August 2011. We derive a spherical cap method for GRACE time-variable gravity data that is a significant improvement from existing least squares mascon methods for analyzing subregions of the Greenland ice sheet. We use the results of the comparison to evaluate different GIA corrections by taking advantage of the difference in magnitude, spatial pattern, and time series of the observed signal and associated uncertainties, and the different sensitivities of each method to GIA errors. For Greenland, we conclude that the Wu10-GIA correction is not compatible with observations of ice elevation changes and reconstructions of SMB, whereas the SM09-GIA and AW13-GIA are compatible with these other observations. The same methodology could be applied in Antarctica to evaluate the regional GIA corrections.
Figure 2.1: Rates of ice sheet mass change for the period September 2003–August 2011 calculated using GRACE monthly solutions corrected using (a) SM09-GIA, (b) AW13-GIA, and (c) Wu10-GIA. Red line is the zero contour of mass change (0 cm/yr). Red asterisk denotes the location of Adolf Hoel Gletscher in NE Greenland. White lines define four regions: 1) Northwest (NW), 2) Northeast (NE), 3) Southwest (SW) and 4) Southeast (SE) for which we derive the regional trends in Table 2.1 and regional time series in Figure 2.4.
Table 2.1: Ice sheet-wide and regional mass changes for the time period September 2003 to August 2011

<table>
<thead>
<tr>
<th>Region</th>
<th>Greenland Total</th>
<th>Northeast (NE)</th>
<th>Northwest (NW)</th>
<th>Southeast (SE)</th>
<th>Southwest (SW)</th>
</tr>
</thead>
<tbody>
<tr>
<td>AW13</td>
<td>$-253 \pm 21 \mathrm{Gt/yr}$</td>
<td>$-17 \pm 6 \mathrm{Gt/yr}$</td>
<td>$-74 \pm 7 \mathrm{Gt/yr}$</td>
<td>$-107 \pm 3 \mathrm{Gt/yr}$</td>
<td>$-55 \pm 5 \mathrm{Gt/yr}$</td>
</tr>
<tr>
<td>SM09</td>
<td>$-256 \pm 23 \mathrm{Gt/yr}$</td>
<td>$-20 \pm 6 \mathrm{Gt/yr}$</td>
<td>$-69 \pm 10 \mathrm{Gt/yr}$</td>
<td>$-111 \pm 3 \mathrm{Gt/yr}$</td>
<td>$-55 \pm 5 \mathrm{Gt/yr}$</td>
</tr>
<tr>
<td>Wu10</td>
<td>$-189 \pm 27 \mathrm{Gt/yr}$</td>
<td>$29 \pm 11 \mathrm{Gt/yr}$</td>
<td>$-62 \pm 10 \mathrm{Gt/yr}$</td>
<td>$-103 \pm 3 \mathrm{Gt/yr}$</td>
<td>$-53 \pm 4 \mathrm{Gt/yr}$</td>
</tr>
<tr>
<td>RACMO</td>
<td>$-173 \pm 7 \mathrm{Gt/yr}$</td>
<td>$-25 \pm 2 \mathrm{Gt/yr}$</td>
<td>$-46 \pm 3 \mathrm{Gt/yr}$</td>
<td>$-45 \pm 2 \mathrm{Gt/yr}$</td>
<td>$-57 \pm 5 \mathrm{Gt/yr}$</td>
</tr>
</tbody>
</table>

The top four rows are the three GRACE estimates of mass change and the trends in RACMO SMB values. Rows five to seven are the relative scales of the three GRACE estimates. The three bottom rows are the relative scales of the GRACE estimates versus SMB. Errors denote the 95% confidence level. SM09, AW13, and Wu10 denote the GRACE ice mass estimates obtained using, respectively, the Simpson et al (2010), A et al. (2013) and Wu et al. (2010) GIA corrections.

Figure 2.2: Mascons used for the Greenland Ice Sheet. Each disc represents a single mascon. In Greenland, mascons of the same color define each of the four regions for which we derive the regional trends in Table 2.1 and regional time series in Figure 2.4: NW (red), NE (blue), SW (purple) and SE (green). Mascons used to calculate the signal from the Canadian Glaciers and Ice Caps (GIC) are shown in orange.
Figure 2.3: Greenland ice mass and elevation changes. (a-c) GRACE ice mass changes corrected using SM09-GIA in cm/yr of water; (d-f) altimetry measurements of ice volume change corrected using SM09-GIA in cm/yr of surface elevation; (g-i) RACMO SMB changes in cm/yr of water for the time periods (a,d,g) September 2003–August 2011, (b,e,h) September 2003–September 2007, (c,f,i) September 2007–August 2011. Red line denotes the contour at 0 cm/yr. White lines define the four regions, NW, NE, SW and SE.
Figure 2.4: Time series of ice mass in Gigatonnes over the period September 2003 – August 2011 for the four regions (a) NW, (b) NE, (c) SW and (d) SE Greenland shown in Figure 2.2. GRACE ice mass estimates calculated using AW13-GIA (blue), SM09-GIA (red), Wu10-GIA (green) are shown alongside estimates of RACMO SMB (orange).
Figure 2.5: Time series of ice mass in Gigatonnes over the period September 2003 – August 2011 for the whole of Greenland (a), and the four regions (b) NW, (c) NE, (d) SW and (e) SE Greenland shown in Figure 2.2. GRACE ice mass estimates calculated using mascon coefficients from Jacob et al. [2012b] corrected with SM09-GIA are shown in blue, scaled GRACE ice mass estimates calculated using large mascons corrected with SM09-GIA are shown in green, GRACE ice mass estimates from this study corrected with SM09-GIA are shown in red, and RACMO SMB estimates from this study are shown in orange.
Chapter 3

Regional acceleration in ice mass loss from Greenland and Antarctica using GRACE time-variable gravity data.

As appears in:

Abstract

We use GRACE monthly gravity fields to determine the regional acceleration in ice mass loss in Greenland and Antarctica for 2003–2013. The mass loss for both ice sheets is largely controlled by only a few regions. In Greenland, the southeast and northwest sectors generate 70% of the loss (280 ± 58 Gt/yr) largely from ice dynamics. The total acceleration in loss is dominated by the southwest sector that accounts for 54% of (25.4 ± 1.2 Gt/yr²) from a decrease in surface mass balance (SMB), followed by the northwest sector (34%). We find no significant acceleration in the northeast sector. In Antarctica, the Amundsen Sea sector and the Peninsula account for 64% and 17%, respectively, of the total loss (180 ± 10 Gt/yr)
mainly from ice dynamics. The Amundsen Sea sector contributes most of the acceleration in loss (11 ± 4 Gt/yr²) and Queen Maud Land in East Antarctica is the only sector with a significant mass gain due to a local increase in SMB (63 ± 5 Gt/yr).

3.1 Introduction

At present, the Greenland and Antarctic Ice Sheets are undergoing significant changes in mass [Shepherd et al., 2012]. In Greenland, the contribution to sea level has increased from 0.09 mm/yr over 1992–2001 to 0.59 mm/yr over 2002–2011 [Vaughan et al., 2013]. The Greenland mass loss signal is due to a combination of increases in ice flow and increases in snow/ice melt [Rignot et al., 2011b]. In Antarctica, the contribution to sea level has increased from 0.08 mm/yr over 1992–2001 to 0.4 mm/yr over 2002–2011, which is largely due to increases in ice flow in West Antarctica and the Antarctic Peninsula. Time-variable gravity data from the Gravity Recovery and Climate Experiment (GRACE) mission have been used to estimate the mean ice sheet mass balance over certain time periods, $dM/dt$, where $M(t)$ is the ice sheet mass at time $t$. More recently, attempts have been made to estimate the changes in mass balance with time or acceleration, $d^2M/dt^2$, in Greenland and Antarctica, at the continental scale [Velicogna, 2009, Wouters et al., 2013, Svendsen et al., 2013, Williams et al., 2014]. Few attempts have been made to determine the acceleration of ice mass change at the regional scale. One major difficulty in determining any accelerations from GRACE time-variable gravity data is that the time series begins in 2002, which is short in comparison to natural climate variability. Analyses of synoptic changes in surface mass balance indicate that several decades of observation may be necessary to separate the short-term variability in surface mass balance from the long-term trends in total mass balance [Wouters et al., 2013]. In addition, the inherent GRACE resolution limits the minimum size of the region for which mass changes can be studied.
Here, we use monthly measurements of time-variable gravity from GRACE for the period January 2003 to December 2013 to estimate the spatial patterns of the average ice mass balance rate ($dM/dt$), and the change in mass balance rate ($d^2M/dt^2$) for both Greenland and Antarctica. We study the temporal variability at both regional and continental scales using a least squares mascon approach [Sutterley et al., 2014, Chapter 2]. Our goal is to evaluate whether the ice mass loss is increasing with time, constant, or not statistically significant. We compare our GRACE results with monthly time series of cumulative surface mass balance (SMB) from the Regional Atmospheric Climate Model (RACMO2.1) to determine the fraction of the GRACE signal that is explained by changes in SMB. Based on our results, we conclude on the nature and statistical significance of changes in ice mass balance for key regions of Greenland and Antarctica.

3.2 Data and Methods

We use RL05 GRACE gravity solutions in the form of spherical harmonic coefficients truncated to maximum degree 60, computed by the Center for Space Research at the University of Texas for January 2003 to December 2013 [Bettadpur, 2012a]. We replace the GRACE $C_{20}$ coefficients with monthly values from satellite laser ranging [Cheng and Tapley, 2004]. We include geocenter coefficients calculated as described in Swenson et al. [2008]. Leakage effects from outside the ice sheet, including the eustatic ocean, are calculated as described in Velicogna and Wahr [2013]. The signal associated with glacial isostatic adjustment (GIA), the viscoelastic response of the solid Earth to glacial unloading over the past several thousand years, is subtracted from the GRACE data. In Antarctica, we use coefficients from the Ivins et al. [2013] regional ice deglaciation model. In Greenland, we use coefficients from the Simpson et al. [2009] regional ice deglaciation model validated in prior work [Sutterley et al., 2014, Chapter 2]. The uncertainty in the GIA correction is calculated by considering a range of rheological parameters for the selected deglaciation model. Over the time span of
the mission, the GIA signal is present as a linear trend in \( M(t) \), or a constant in the rate of mass change, \( dM/dt \). The secular nature of the GIA signal means that it does not contribute to the acceleration in mass, \( d^2 M/dt^2 \), or the acceleration uncertainty.

We smooth the Stokes coefficients using a Gaussian averaging function with a 250 km radius [Wahr et al., 1998], and we calculate regular latitude-longitude grids of monthly ice mass anomaly (Equation 1.10). Before we calculate ice mass grids for Greenland, we remove the contamination from the nearby Canadian, Icelandic and Svalbard Glaciers and Ice Caps (GIC) from the GRACE data. We do this by applying the least squares mascon approach described in Sutterley et al. [2014] over the GIC regions and the ice sheet. Each mascon is a 3° diameter equal-area spherical cap with a mass equal to a uniformly distributed centimeter of water [Farrell, 1972, Chapter 2]. For each mascon, we calculate a set of Stokes coefficients, which we smooth with a 250 km Gaussian function and convert into mass. We simultaneously fit the mascon coefficients to the monthly GIA-corrected GRACE coefficients in order to obtain estimates of the monthly mass variability for each mascon. We obtain an estimate of the GIC signal by combining the signal of the corresponding GIC mascons. We remove this signal from the GRACE monthly coefficients and calculate a set of ice mass grids for the Greenland ice sheet.

We use the Greenland and Antarctic grids to form maps of the acceleration in ice mass loss by simultaneously fitting annual and semiannual signals, a quadratic trend, a linear trend and a constant at each grid point. To be conservative in our error estimate when fitting the regression model, we do not make any assumption about the size of the error affecting the GRACE monthly estimates. We only consider points for which the acceleration is significant at the 2-\( \sigma \) level. To determine if the quadratic model is statistically more significant than the linear one, we also calculate the best fitting linear trend and the associated error. We select the points for which the trend is significant at the 2-\( \sigma \) level and we use a variant of the Akaike Information Criterion for use with small-sample sized datasets (\( AIC_c \)) to compare
the quadratic and linear models to identify the model that best fits the signal [Burnham and Anderson, 2002]. This will determine if the mass balance rate is increasing with time rather than constant at each grid point.

The results in Figure 3.1 are a smoothed and damped representation of the mass signal. To examine quantitatively the temporal evolution of ice mass changes at the regional scale, we calculate time series for selected regions using the least squares mascon approach. For this study, we improve the original spherical cap methodology used in [Sutterley et al., 2014, Chapter 2] that uses a distribution along latitude lines. We create new spherical cap configurations for Greenland and Antarctica tiling the spherical caps hexagonally, which reduces and optimizes the gaps between neighboring caps. Our new configuration reduces the overall noise levels in the resultant time series, and improves the signal recovery of the synthetic mass fields.

The error of each regional ice mass estimate is due to signal leakage, GRACE measurement errors, GIA uncertainty and statistical uncertainty of the model fit. We evaluate each of these contributors as described below. For each region, we calculate the corresponding sensitivity kernel to evaluate how mass at a given point within the region contributes to the total time series [Sutterley et al., 2014, Chapter 2]. If the GRACE mass anomalies are distributed uniformly over each mascon, the fit results will recover the total variability for each region. We evaluate the error introduced by assuming a uniform mass distribution within each mascon by estimating the leakage error within each region using a field of simulated, realistic ice mass change based on results from the mass budget method [Rignot et al., 2011b]. We evaluate the leakage within each region from the outside signal by convolving each sensitivity kernel with the simulated signal for all regions except the one selected. We include this leakage error in our regional error budget. The measurement errors in individual GRACE monthly fields are calculated by convolving each sensitivity kernel with uncertainty estimates for the GRACE Stokes coefficients following Wahr et al. [2006]. The “statistical” leakage term
is calculated by first removing the retrieved mascon estimates from the GRACE spherical harmonic fields, and then refitting the mascon fields to the harmonics [Tiwari et al., 2009].

In Greenland, we distinguish five regions based on their ice dynamics and surface mass balance (SMB) characteristics [van den Broeke et al., 2009]: 1) the northwest (NW) with high accumulation rates, many small, fast-moving glaciers and the major glacier Jakobshavn Isbræ, 2) the north (N) with few, large, slow-moving ice streams and low snowfall accumulation rates, 3) the northeast (NE) similar to the N but with glaciers flowing through high mountain ranges, 4) the southeast (SE) with high accumulation rates and many fast, small outlet glaciers, and 5) the southwest (SW) with many land terminating glaciers and a very large ablation area [van den Broeke et al., 2009].

In Antarctica, we distinguish five regions identified based changes highlighted in prior studies [Rignot et al., 2008b, Chen et al., 2009, Pritchard et al., 2009, Shepherd et al., 2012]: 1) the Amundsen Sea (AS) sector of West Antarctica, 2) the Antarctic peninsula (AP), 3) Queen Maud Land (QML), 4) the Totten/Moscow/Frost sector (TMF), and 5) Victoria/Wilkes land (VW) in East Antarctica. In Antarctica, the sum of all the regions is not equal to the entire ice sheet (AIS). All regions are selected to be compatible in terms of GRACE errors and resolution. Figures 3.4–3.5 show the configuration of the mascons and the regional boundaries for both ice sheets.

We compare the GRACE-derived time series of ice mass change with time series of cumulative SMB anomalies from the Regional Atmospheric Climate Model (RACMO2.1) [van den Broeke et al., 2006]. RACMO2 is forced at the lateral boundary and at the sea surface by the latest reanalysis of the European Centre for Medium-Range Weather Forecasts (ERA-40 and ERA-Interim) [Simmons et al., 2006/2007]. The most recent version of RACMO2 does not assimilate the field data of van de Berg et al. [2006] for calibration, but uses the data to estimate its absolute precision. In the Antarctic, the uncertainty (1σ) in SMB for the grounded ice sheet averages 7% or 144 Gt/yr [Lenaerts et al., 2012]. The Antarctic RACMO2 data
is currently available through July 2012. In Greenland, the uncertainty in SMB averages 17% or 77 Gt/yr [Ettema et al., 2009]. The Greenland RACMO2 data is currently available through December 2012. We calculate cumulative SMB anomalies relative to longterm average climatologies. For Greenland, we use the reference period 1961–1990, a period when the ice sheet was close to balance [van den Broeke et al., 2009]. For Antarctica, we use the reference period of 1979–2008 [Shepherd et al., 2012]. Error in monthly cumulative SMB anomalies are calculated as described by [van den Broeke et al., 2009].

The goals of the comparison of GRACE with cumulative SMB anomalies is to determine how well the two time series agree in different regions, and to determine which fraction of the GRACE signal is explained by SMB variability. To perform this comparison, the SMB data are processed in the same manner as the GRACE data by converting into the spectral domain, truncating to degree 60, and averaging spatially with a Gaussian filter (Equations 1.9 and 1.10). To compare temporally at the regional level, we apply the same mascon fit technique to the SMB data as described for the GRACE data.

### 3.3 Results

In Greenland, the ice mass loss has spread across the entire ice sheet. In total, we find a linear trend in mass loss of $280 \pm 58$ Gt/yr and an acceleration in loss of $25.4 \pm 1.2$ Gt/yr$^2$ (Table 3.1, GIS). The uncertainties include contributions added in quadrature from the GIA correction, leakage of signal from outside the ice sheet including the eustatic ocean, and the statistical uncertainty of the fit. The signal from the Canadian GIC, which was removed from the Greenland signal, corresponds to a mass loss of $74 \pm 7$ Gt/yr with an acceleration in loss of $10 \pm 2$ Gt/yr$^2$. Our mass loss assessment for Canadian GIC agrees with some recent estimates [Lenaerts et al., 2013, Gardner et al., 2013].

The mass loss rate in Greenland is increasing with time for several regions. Over most of
Greenland Ice Sheet, this acceleration is statistically significant at the 2-σ level (Figure 3.1c). While the largest losses are recorded in SE and NW (Figure 3.1a), the largest acceleration in mass loss is from the SW. In the NW, the mass loss rate is also increasing at a statistically significant level, but its magnitude is low, and the signal is confined along the coast. In the N, the acceleration in ice loss is significant. In the SE and portions of NE, we detect significant losses during the analyzed period, but no statistically significant increase in the mass loss rate with time.

In Antarctica, we detect an overall mass loss rate of $67 \pm 44 \text{ Gt/yr}$ and an acceleration in loss of $11 \pm 4 \text{ Gt/yr}^2$ during 2003–2013 (Table 3.1, AIS). The mass loss affects a portion of the ice sheet (Figure 3.1b). The largest losses are in AS and in the northern tip of AP. The mass loss in AS extends beyond its drainage into the surrounding basins. Much smaller losses are detected in East Antarctica in the VW (near Cook Ice Shelf) and TMF (near Totten Glacier) regions. The GRACE signal for the VW and TMF regions is concentrated along the coast. Some areas are experiencing a mass gain, most notably in QML.

In contrast with Greenland, most Antarctic regions do not experience an acceleration in mass change at a statistically significant level (Figure 3.1d). We detect an acceleration in mass loss in the AS sector and on the western and southern tip of AP, and a statistically significant acceleration in mass gain in QML, East Antarctica. Elsewhere in East Antarctica, acceleration values are low and generally not significant. Error maps for both ice sheets for the linear trend and acceleration are shown in Figure 3.6.

To examine changes in ice mass quantitatively at the regional scale, we calculate time series of ice mass change for the selected five regions of each ice sheet (Figure 3.2-3.3). Results for each region are summarized in (Table 3.2). For Greenland, the mass loss in the SE is $110 \pm 21 \text{ Gt/yr}$, with no statistically significant acceleration (Figure 3.2f). The time series reveals that the region experienced large losses in 2003–2007, followed by a slow down in 2008–2009, and a small increase in mass loss since 2010. The overall SE mass change is
best described as a constant mass loss. In the NW, the mass loss is $87 \pm 22$ Gt/yr with a statistically significant acceleration in loss of $8.7 \pm 2.1$ Gt/yr$^2$. Combined together, the SE and NW regions generate 70% of the mass loss of the entire ice sheet. In the SW and N, we find statistically significant accelerations in loss of $13.7 \pm 0.8$ Gt/yr$^2$ and $3.5 \pm 0.5$ Gt/yr$^2$, respectively. These regions are responsible for 12%, or $34 \pm 12$ Gt/yr and 11.8%, or $33 \pm 11$ Gt/yr of the total loss. The SW and NW regions are the largest contributors to the total acceleration in loss, with 54% from the SW and 34% from the NW. In the NE, the mass loss of $15 \pm 9$ Gt/yr has a statistically significant acceleration of only $1.5 \pm 1.1$ Gt/yr$^2$.

When comparing the GRACE results with the cumulative SMB anomaly time series during the common time period of January 2003–December 2012, we find most of the GRACE signal in the N, NE and SW regions is explained by the variability in SMB (Figure 3.2). In the SW, the acceleration in loss from GRACE and SMB are $15.2 \pm 0.9$ Gt/yr$^2$ and $13.4 \pm 2.6$ Gt/yr$^2$, respectively. In the N, acceleration in loss from GRACE and SMB are $5.2 \pm 0.6$ Gt/yr$^2$ and $4.1 \pm 1.0$ Gt/yr$^2$, respectively. In contrast, in the SE and NW, the cumulative SMB anomaly only accounts for a small portion of the total mass change (Table 3.3, GIS). In the SE, GRACE mass change is $-108 \pm 21$ Gt/yr and SMB is $-46 \pm 9$ Gt/yr. In the NW, SMB change is $-33 \pm 6$ Gt/yr or 39% of the GRACE mass change ($-85 \pm 22$ Gt/yr), but it accounts for most of the acceleration in loss, with $7.9 \pm 1.7$ Gt/yr$^2$ of the $10.8 \pm 3.1$ Gt/yr$^2$s detected by GRACE.

In Antarctica (Figure 3.3), the AP experiences a mass loss of $31 \pm 4$ Gt/yr with an acceleration in loss of $3.2 \pm 0.6$ Gt/yr$^2$. The significant portion of the AP mass loss signal is in northern tip of the peninsula, averaging $18 \pm 3$ Gt/yr, however, the acceleration here is positive, averaging $2.5 \pm 0.6$ Gt/yr$^2$. We detect a net increase in mass loss for the entire AP after 2006, but the loss decreases in time along its northern tip. In the AS sector, the mass loss is $116 \pm 6$ Gt/yr, with an acceleration in loss of $13 \pm 2$ Gt/yr$^2$. This region account for 94% of the West Antarctic mass loss and dominates the total ice sheet mass loss. In the TMF sector, the mass
loss is $17 \pm 4 \text{ Gt/yr}$, with an acceleration in loss of $4 \pm 0.7 \text{ Gt/yr}^2$. We note a large variability in cumulative SMB anomaly during the analyzed period. In QML, the mass balance and acceleration for the entire period are $63 \pm 6 \text{ Gt/yr}$ and $15 \pm 1 \text{ Gt/yr}^2$, respectively. In VW, the mass change is $-16 \pm 5 \text{ Gt/yr}$, with most changes occurring between 2007–2010.

When comparing GRACE with cumulative SMB anomaly in Antarctica during the common period, January 2003 to July 2012, we find an excellent agreement in VW ($-19 \pm 5 \text{ Gt/yr}$ for GRACE versus $-24 \pm 4 \text{ Gt/yr}$ for SMB), and good agreement in QML ($16.8 \pm 1.2 \text{ Gt/yr}^2$ for GRACE versus $18.2 \pm 2.8 \text{ Gt/yr}^2$ for SMB) and TMF regions ($-15 \pm 4 \text{ Gt/yr}$ for GRACE versus $-17 \pm 2 \text{ Gt/yr}$ for SMB). In the AS and AP regions, SMB only explains a small fraction of the total change in mass observed by GRACE. In AS, mass change from GRACE is $-110 \pm 6 \text{ Gt/yr}$ and $6 \pm 3 \text{ Gt/yr}$ from SMB. In AP, the mass loss is $31 \pm 4 \text{ Gt/yr}$ from GRACE and $14 \pm 4 \text{ Gt/yr}$ from SMB. Over the entire ice sheet, SMB only explains a small part of the evolution of the change in mass balance (Table 3.3, AIS).

### 3.4 Discussion

The agreement between GRACE and cumulative SMB anomaly in N, NE and SW Greenland indicates that SMB dominates the variability of the GRACE data in these regions. This implies that changes in ice dynamics are not a significant portion of the GRACE mass signal in these regions. This is confirmed by observations of steady flow in these sectors, especially N and SW during the analyzed period [Moon et al., 2012]. The modest acceleration of Zachariae Isstrøm in the NE after 2004 as observed in [Rignot et al., 2008b] does not produce a detectable increase in mass loss in the GRACE data. In the SW, we find that the mass loss is mainly driven by SMB, which is in agreement with the results of van den Broeke et al. [2009] showing that the loss is driven by enhanced runoff. The agreement between GRACE and cumulative SMB anomalies indirectly provides additional confidence in both the GRACE
analysis and the retrieval of SMB from RACMO2. In the SW, where ice dynamics is not a major factor, the agreement between GRACE and cumulative SMB anomaly spans from the monthly to the decadal time scales, which increases confidence in both the GRACE analysis and the retrieval of SMB from RACMO2.

In SE and NW, we find that ice dynamics plays significant role, which is consistent with recent studies [Howat et al., 2007, Moon et al., 2012]. In the SE, where the mass loss is the largest in Greenland and the contribution of ice dynamics is significant, the mass loss has remained more or less constant in time despite glacier acceleration in 2000–2005 followed by a deceleration after 2005 [Howat et al., 2007]. This suggests that changes in SMB have coincidentally counteracted changes in glacier discharge. Overall, in Greenland, SMB has contributed 68% of the GRACE-derived mass loss ($-180 \pm 33$ Gt/yr versus a total loss of $-265 \pm 59$ Gt/yr), and 79% of the observed acceleration ($23.3 \pm 4.7$ Gt/yr$^2$ versus a total acceleration of $29.7 \pm 1.3$ Gt/yr$^2$) during 2003–2012.

In the NE, our mass loss and acceleration estimates do not agree with the recent analysis by Khan et al. [2014]. When we select a region covering the northeast ice stream sector used by the same authors (Figure 3.7b), we find a mass balance rate of $-10 \pm 4$ Gt/yr for the entire period, an acceleration of $-1.0 \pm 0.7$ Gt/yr$^2$ (Figure 3.7a). The agreement between GRACE and cumulative SMB anomaly for the period January 2003–December 2012 suggests that the mass loss is predominantly controlled mostly by a temporal variability in SMB. When we subtract the anomaly in discharge from Table 2 in Khan et al. [2014] from the cumulative SMB anomaly time series, the difference between the resulting time series of mass change and GRACE exceeds the error bars (Figure 3.7a). The hypothesis that this sector is undergoing a major instability in discharge and that the glaciers sped up after 2006 is not confirmed by our data. Resolving these differences will require further study.

In Antarctica, GRACE and cumulative SMB anomalies agree in many regions, particularly TMF and VW (Table 3.3). This provides strong support for the RACMO2 SMB reconstruc-
tion. Conversely, the differences between cumulative SMB anomalies and GRACE in the AS sector indicates that changes in glacier flow are dominant. Changes in the AS regional ice dynamics are well known [Mouginot et al., 2014], and SMB accounts for less than 10% of the total mass balance during January 2003–July 2012. In AP, SMB explains $45 \pm 13\%$ of the observed mass changes during the common period and displays no significant acceleration.

Prior work by Williams et al. [2014] compared the GRACE time series at a single location inland of Totten Glacier (representing an area of 250,000 km$^2$) with the time series of RACMO cumulative SMB anomaly for the entire drainage basin of Totten/Moscow/Frost (area 538,000 km$^2$). The authors found a consistent behavior between GRACE and SMB only for 2008–2010. Here, our comparison of the datasets for the TWF region indicates a good agreement for the entire period January 2003–July 2012. Cumulative SMB anomalies account for most of the total mass balance during this period, however, the increase in SMB only explains 40% of the GRACE acceleration signal for the period. Ice dynamics may play a role in this sector as suggested by recent studies [Flament and Rémy, 2012, Khazendar et al., 2013]. Our GRACE time series is too short to be conclusive.

Similarly, over the VW region, we find that GRACE closely follows the fluctuations in cumulative SMB anomaly, which suggests that ice dynamics is not contributing significantly to the GRACE mass trend. There are no reports of large fluctuations in the velocities of these glaciers [Shepherd et al., 2012]. Our results also do not confirm earlier conclusions for the time period 2002–2008 that a significant loss of ice is taking place in the Cook Ice Shelf region [Chen et al., 2009]. We attribute these earlier changes to a temporal variability in SMB. In VW, our analysis complements the work of Sasgen et al. [2013] by showing that the agreement between GRACE and cumulative SMB anomaly extends over the entire time period on a monthly basis.

We note that quadratic trends in SMB in QML, VW and TMF likely do not represent a longterm trend in SMB. Rather, the increase in SMB in QML is an isolated event, while
the trends in VW and TMF are well within interannual variations in SMB for those sectors [Rignot et al., 2011b].

Our results expand upon Wouters et al. [2013] who examined the whole ice sheet and concluded that most observed changes are controlled by SMB, and that is is not possible to detect an acceleration in loss in Antarctica with the existing GRACE time series. While their conclusion is correct for the entire ice sheet, the results are different at the regional scale. For example, in the AS sector, the uncertainty in acceleration is only 15% of the signal. Similarly, in Greenland, some regions exhibit a strong acceleration in loss, such as the $13.7 \pm 0.8 \text{ Gt/yr}^2$ acceleration in the SW caused by an increase in runoff since the 1980s [van den Broeke et al., 2009]. By examining data at the regional scale, our ability to observe acceleration in loss improves rather than degrades as the mass changes are not uniform. The signal-to-noise ratio of the mass change at the regional scale may be larger than the signal-to-noise ratio of the entire ice sheet.

The difference in fit skill between the quadratic and linear models is generally less than 5% over the analyzed regions, except in regions where the acceleration is large, such as SW Greenland and the AS, QML and TMF regions of Antarctica (Table 3.2). The improvement in fitting of the GRACE data with a quadratic model is always significant at a very high confidence level (98–99%), except for VW in Antarctica (95%) and NE in Greenland (96%). SE Greenland is the only region where a linear fit of the data is currently the best solution.

3.5 Conclusion

We evaluate the significance of linear and quadratic trends in Greenland and Antarctica at the continental and regional scales. The regions with acceleration signals appear clearly over the subcontinent and continent. The mascon analysis provides quantitative estimates of the mass loss and acceleration at the regional level. We find that only a few regions drive the
mass loss and acceleration for both ice sheets. In Greenland, the mass loss is controlled by the SE (40%) and NW (30%), and the acceleration by the SW (54%) and NW (34%). The largest acceleration in loss is caused by a decrease in cumulative SMB anomalies in the SW, whereas fluctuations in ice dynamics in the SE and NW do not result in a significant acceleration over the study period. In the NE, changes in mass are small, largely driven by SMB changes and not suggestive of dynamic instability as reported by Khan et al. [2014]. During 2003–2012, SMB changes accounts for 68% of the mass loss of the entire ice sheet and 79% of its acceleration.

In Antarctica, most of the mass loss is from the AS and AP regions, and most of the acceleration is from the AS region. In QML, SMB has increased to yield a positive mass balance for the years after 2008. Longer time series of GRACE data are required to determine the significance of changes in mass loss observed in other sectors of Antarctica.

**Table 3.1:** GRACE regional ice mass balance and acceleration

<table>
<thead>
<tr>
<th>Region</th>
<th>Trend (Gt/yr)</th>
<th>Acceleration (Gt/yr²)</th>
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<tr>
<td>GIS</td>
<td>−280 ± 58</td>
<td>−25.4 ± 1.2</td>
</tr>
<tr>
<td>SE</td>
<td>−110 ± 21</td>
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<tr>
<td>NW</td>
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<td>N</td>
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<td>SW</td>
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<tr>
<td>TMF</td>
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</tr>
<tr>
<td>VW</td>
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<td>2.1 ± 0.8</td>
</tr>
</tbody>
</table>

Figure 3.1: Greenland and Antarctica linear trends in ice mass balance, $dM/dt$, in cm of water (a,b), and accelerations in ice mass balance, $d^2M/dt^2$, in cm/yr$^2$ (c, d) determined from GRACE data for January 2003–December 2013. Contour interval is 2 cm/yr for the linear trend (a,b) and 0.5 cm/yr$^2$ for the acceleration. White lines define regions for which time series are calculated in Figures 3.2 and 3.3 and Tables 3.1, 3.2, and 3.3.
Figure 3.2: Time series of ice mass, $M(t)$, in Gigatonnes for (a) the entire Greenland Ice Sheet (GIS), (b) North (N), (c) Northeast (NE), (d) Northwest (NW), (e) Southwest (SW) and (f) Southeast (SE) Greenland as shown in Figure 1. GRACE time series for January 2003–December 2013 are shown in blue, the best-fitting trend is shown in green. Monthly errors are shown as light blue band. Time series of cumulative SMB anomaly for January 2003–December 2012 are shown in red; monthly errors are shown as light red band. Included are GRACE linear and quadratic trend estimates.
Figure 3.3: Time series of ice mass, $M(t)$, in Gigatonnes for (a) the entire Antarctic Ice Sheet (AIS), (b) the Antarctic Peninsula (AP), (c) the Amundsen Sea (AS) sector (d) Queen Maud Land (QML), (e) the Totten/Moscow/Frost sector (TMF), (f) Victoria/Wilkes land (VW) as shown in Figure 1. GRACE time series for January 2003–December 2013 are shown in blue, the best-fitting trend is shown in green. Light blue band are monthly errors. Time series of cumulative SMB anomaly for January 2003–July 2012 are shown in red; light red band are monthly errors. Included are GRACE linear and quadratic trend estimates.
Table 3.2: Longterm mass balance and mass acceleration for all the analyzed regions calculated simultaneously using a least squares regression for the period January 2003–December 2013 relative to the mean date of the period, which is April 2008.

<table>
<thead>
<tr>
<th>Region</th>
<th>Trend (Gt/yr)</th>
<th>$R^2_{adj}$</th>
<th>$AIC_c$</th>
<th>Acceleration (Gt/yr²)</th>
<th>$R^2_{adj}$</th>
<th>$AIC_c$</th>
</tr>
</thead>
<tbody>
<tr>
<td>GIS</td>
<td>−280 ± 58</td>
<td>97</td>
<td>1601</td>
<td>−25.4 ± 2.2</td>
<td>99</td>
<td>1483</td>
</tr>
<tr>
<td>SE</td>
<td>−110 ± 21</td>
<td>97</td>
<td>1386</td>
<td>+1.9 ± 2.9</td>
<td>97</td>
<td>1386</td>
</tr>
<tr>
<td>NW</td>
<td>−87 ± 22</td>
<td>96</td>
<td>1364</td>
<td>−8.7 ± 2.3</td>
<td>98</td>
<td>1285</td>
</tr>
<tr>
<td>N</td>
<td>−33 ± 11</td>
<td>92</td>
<td>1216</td>
<td>−3.5 ± 0.8</td>
<td>94</td>
<td>1179</td>
</tr>
<tr>
<td>SW</td>
<td>−34 ± 12</td>
<td>68</td>
<td>1450</td>
<td>−13.7 ± 1.3</td>
<td>87</td>
<td>1339</td>
</tr>
<tr>
<td>NE</td>
<td>−15 ± 10</td>
<td>68</td>
<td>1257</td>
<td>−1.5 ± 1.4</td>
<td>68</td>
<td>1255</td>
</tr>
<tr>
<td>AIS</td>
<td>−67 ± 42</td>
<td>60</td>
<td>1675</td>
<td>−10.6 ± 4.3</td>
<td>62</td>
<td>1669</td>
</tr>
<tr>
<td>QML</td>
<td>+63 ± 6</td>
<td>84</td>
<td>1475</td>
<td>+15.1 ± 1.5</td>
<td>94</td>
<td>1361</td>
</tr>
<tr>
<td>AP</td>
<td>−31 ± 4</td>
<td>87</td>
<td>1278</td>
<td>−3.2 ± 1.0</td>
<td>89</td>
<td>1261</td>
</tr>
<tr>
<td>AS</td>
<td>−116 ± 6</td>
<td>96</td>
<td>1421</td>
<td>−12.7 ± 1.9</td>
<td>99</td>
<td>1299</td>
</tr>
<tr>
<td>TMF</td>
<td>−17 ± 4</td>
<td>67</td>
<td>1160</td>
<td>−4.0 ± 1.0</td>
<td>74</td>
<td>1160</td>
</tr>
<tr>
<td>VW</td>
<td>−16 ± 5</td>
<td>43</td>
<td>1374</td>
<td>2.1 ± 1.6</td>
<td>44</td>
<td>1373</td>
</tr>
<tr>
<td>EAIS_{other}</td>
<td>+57 ± 36</td>
<td>86</td>
<td>1417</td>
<td>+4.9 ± 5.5</td>
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<td>1408</td>
</tr>
<tr>
<td>WAIS_{other}</td>
<td>−7 ± 13</td>
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<td>1484</td>
<td>−12.7 ± 2.4</td>
<td>48</td>
<td>1422</td>
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</tbody>
</table>

Longterm mass balance and mass acceleration calculated simultaneously using a least squares regression for the period January 2003–December 2013, relative to the mean date of the period (April 2008), for different regions of the Greenland Ice Sheet (Northwest, NW; North, N; Northeast, NE; Southeast, SE; Southwest, SW) and the entire ice sheet (GIS) and for different regions of the Antarctic Ice Sheet (Amundsen Sea sector, AS; Antarctic peninsula, AP; Queen Maud Land, QML; Totten/Moscow/Frost sector, TMF; Victoria/Wilkes land, VW) and the entire ice sheet (AIS). 2-sigma errors are included. $AIC_c$ indicates the model that best fits the signal based on the Akaike Information Criterion: L indicates linear model and Q the quadratic model. $R^2_{adj}(L)$ and $R^2_{adj}(Q)$ are R-square adjusted values for linear (L) and quadratic (Q) model respectively, F-test indicates at which statistical level the improvement in the quadratic model fit is significant. In Antarctica, the sum of all the regions is not equal to the entire ice sheet.
Table 3.3: GRACE and cumulative surface mass balance (SMB) longterm mass balance and mass acceleration for the Greenland and Antarctica Ice Sheets, and analyzed subregions. Trends are calculated for the common period to the two datasets: January 2003–December 2012 for Greenland and January 2003–July 2012 for Antarctica. 2-sigma error are included, only statistically significant trends are shown. $R^2_{adj}(L)$ and $R^2_{adj}(Q)$ are R-square adjusted values for linear trend (L) and acceleration (Q) respectively. In Antarctica, the sum of all the regions is not equal to the entire ice sheet.

<table>
<thead>
<tr>
<th>Region</th>
<th>Linear Trend ($Gt/yr$)</th>
<th>$R^2_{adj}$ Linear</th>
<th>Acceleration ($Gt/yr^2$)</th>
<th>$R^2_{adj}$ Quadratic</th>
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</thead>
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<tr>
<td></td>
<td>GRACE</td>
<td>SMB</td>
<td>GRACE</td>
<td>SMB</td>
</tr>
<tr>
<td>GIS</td>
<td>$-265 \pm 59$</td>
<td>$-178 \pm 4$</td>
<td>97</td>
<td>97</td>
</tr>
<tr>
<td>SE</td>
<td>$-108 \pm 21$</td>
<td>$-51 \pm 2$</td>
<td>96</td>
<td>95</td>
</tr>
<tr>
<td>NW</td>
<td>$-85 \pm 22$</td>
<td>$-33 \pm 2$</td>
<td>95</td>
<td>88</td>
</tr>
<tr>
<td>N</td>
<td>$-31 \pm 11$</td>
<td>$-40 \pm 1$</td>
<td>91</td>
<td>97</td>
</tr>
<tr>
<td>SW</td>
<td>$-27 \pm 12$</td>
<td>$-45 \pm 2$</td>
<td>63</td>
<td>86</td>
</tr>
<tr>
<td>NE</td>
<td>$-13 \pm 10$</td>
<td>$-10 \pm 1$</td>
<td>65</td>
<td>90</td>
</tr>
<tr>
<td>AIS</td>
<td>$-59 \pm 47$</td>
<td>$-35 \pm 5$</td>
<td>54</td>
<td>59</td>
</tr>
<tr>
<td>QML</td>
<td>$+55 \pm 6$</td>
<td>$+22 \pm 1$</td>
<td>78</td>
<td>44</td>
</tr>
<tr>
<td>AP</td>
<td>$-31 \pm 5$</td>
<td>$-14 \pm 2$</td>
<td>82</td>
<td>75</td>
</tr>
<tr>
<td>AS</td>
<td>$-110 \pm 6$</td>
<td>$+7 \pm 2$</td>
<td>95</td>
<td>19</td>
</tr>
<tr>
<td>TMF</td>
<td>$-15 \pm 4$</td>
<td>$-17 \pm 2$</td>
<td>56</td>
<td>72</td>
</tr>
<tr>
<td>VW</td>
<td>$-19 \pm 5$</td>
<td>$-23 \pm 1$</td>
<td>49</td>
<td>78</td>
</tr>
</tbody>
</table>
Figure 3.4: Mascons used for the Greenland Ice Sheet. Each disc represents a single mascon. In Greenland, mascons of the same color define each of the five regions for which time series are calculated. Mascons used to calculate the signal from the Canadian Glaciers and Ice Caps (GIC) are shown in dark blue and maroon. Black lines define region boundaries.

Figure 3.5: Mascons used for the Antarctic Ice Sheet. Each disc represents a single mascon. Mascons of the same color define each of the five regions for which time series are calculated. Black lines define region boundaries.
Figure 3.6: Error map for the Greenland and Antarctica Ice Sheets linear trends in ice mass balance, in cm/yr of water (a, b), and the accelerations in ice mass balance, in cm/yr$^2$ (c, d) determined from GRACE data for the time period January 2003–December 2013. Contour interval is 0.1 cm/yr for the linear trend (a,b) and 0.1 cm/yr$^2$ for the acceleration. White lines define regions for which time series are calculated.

Figure 3.7: a) Time series of changes in ice mass balance in Gigatonnes for a region in northeast Greenland that best approximates the NEGIS region used in Khan et al. [2014] for the period January 2003–December 2012. GRACE monthly time series are shown in blue, SMB monthly time series are shown in red, and SMB minus the ice discharge (D) component of Khan et al. [2014] is shown in green based on Table 1 of their paper, row entitled “Dynamic NEGIS basin”. b) Map of Greenland indicating the mascon discs (red) used to extract the GRACE and SMB signal over the NEGIS basin. The boundaries of the NEGIS basin used by Khan et al. [2014] are shown in green.
Chapter 4

Mass loss of the Amundsen Sea Embayment of West Antarctica from four independent techniques

As appears in:

Abstract

We compare four independent estimates of the mass balance of the Amundsen Sea Embayment of West Antarctica, an area experiencing rapid retreat and mass loss to the sea. We use ICESat and Operation IceBridge laser altimetry, Envisat radar altimetry, GRACE time-variable gravity, RACMO2.3 surface mass balance, ice velocity from imaging radars and ice thickness from radar sounders. The four methods agree in terms of mass loss and acceleration in loss at the regional scale. Over 1992–2013, the mass loss is \( 83 \pm 5 \text{ Gt/yr} \) with an acceleration of \( 6.1 \pm 0.7 \text{ Gt/yr}^2 \). During the common period 2003–2009, the mass loss is
84 ± 10 Gt/yr with an acceleration of 16.3 ± 5.6 Gt/yr², nearly three times the acceleration over 1992–2013. Over 2003–2011, the mass loss is 102 ± 10 Gt/yr with an acceleration of 15.7 ± 4.0 Gt/yr². The results reconcile independent mass balance estimates in a setting dominated by change in ice dynamics with significant variability in surface mass balance.

4.1 Introduction

The glaciers flowing into West Antarctica’s Amundsen Sea Embayment (ASE) are a focal point of glaciological studies due to their rapid acceleration, large negative mass balance and unstable bed configuration [Hughes, 1973, Rignot, 1998, 2001]. The ASE glaciers flow with some of the highest surface velocities in the continent while draining a catchment that receives high rates of snowfall [Rignot et al., 2011a, Mouginot et al., 2014, van den Broeke et al., 2006, Lenaerts et al., 2012, Medley et al., 2014]. Observations from satellite radar interferometry have shown significant surface velocity increases on the Pine Island (PIG) and Thwaites (THW) glaciers since the 1990’s in conjunction with significant retreats of their grounding line positions [Rignot, 1998, 2001, Rignot et al., 2002, 2014]. Increased mass fluxes from the smaller regional glaciers of Smith (SMI), Kohler (KOH), Pope (POP), and Haynes (HAY) have also contributed significantly to the overall acceleration in ice mass discharge into the embayment [Thomas et al., 2004, Rignot et al., 2008a, Mouginot et al., 2014]. Mouginot et al. [2014] report a 77% increase in total ice discharge of the ASE (145 ± 22 Gt/yr increase) between 1973 and 2013, with 50% of the discharge increase occurring between 2003 and 2009. Elevation measurements of PIG have shown strong and accelerated dynamic thinning over areas of fast flow, extending from the calving front to the upper tributaries [Park et al., 2013, Pritchard et al., 2009, Flament and Rémy, 2012]. This sector has been identified as the largest contributor from Antarctica to present-day global sea level rise using gravity data and the mass budget method [Rignot et al., 2011b, Velicogna et al., 2014, Chapter 3]. Projections from ice sheet numerical models suggest that the region will continue to be a considerable
source of global sea level rise over the next century [Seroussi et al., 2014, Joughin et al., 2014, Favier et al., 2014]. Significant differences remain among published mass balance estimates for the ASE [Shepherd et al., 2002, Zwally et al., 2005, Rignot et al., 2008a, Sasgen et al., 2013, Medley et al., 2014]. In particular radar altimeter estimates [Shepherd et al., 2002, Zwally et al., 2005] are typically lower than estimates from the mass budget method [Rignot et al., 2008a]. These discrepancies can be partially resolved by comparing data over the same time period and the same region, but prior to this research had not yet been done.

Here, we examine the mass balance of the ASE using four independent methods during the overlapping periods: 1) satellite time-variable gravity, 2) mass budget method (MBM), 3) satellite radar altimetry, and 4) satellite and airborne laser altimetry. We use 12 years of time variable gravity measurements from the NASA/DLR GRACE (Gravity Recovery and Climate Experiment) satellite mission, a 22 year span of ice discharge from synthetic aperture radar (SAR) data combined with ice thickness data derived from Operation IceBridge (OIB) radio echo sounding, 22 years of surface mass balance (SMB) output products from the Regional Atmospheric Climate Model (RACMO2.3) [van Wessem et al., 2014], 9 years of radar altimetry data from the European Space Agency Environmental Satellite (Envisat) mission, 7 years of laser altimetry data from ICESat-1 and 3 years from OIB. We determine the differences between the different methods in terms of mass balance, $dM/dt$, and acceleration in mass balance, $d^2M/dt^2$, and conclude with a reconciled and comprehensive estimate of the ASE contribution to sea level in 1992–2014 evaluated using multiple techniques.

4.2 Data and Methods

We use 135 monthly GRACE Release-5 (RL05) gravity solutions provided by Center for Space Research (CSR) for the period April 2002 to May 2014 [Bettadpur, 2012b]. Each CSR solution consists of fully-normalized spherical harmonic coefficients $(C_{lm}, S_{lm})$ up to degree, $l$,
and order, $m$, 60. We substitute the GRACE-derived $C_{20}$ coefficients with monthly estimates from satellite laser ranging (SLR) [Cheng et al., 2013], and we account for the variation of the Earth’s geocenter using degree-1 coefficients provided by Swenson et al. [2008]. Leakage effects from outside the ice sheet are calculated as described in Velicogna and Wahr [2013]. We correct the GRACE mass changes for glacial isostatic adjustment (GIA), the Earth’s viscoelastic response to the glacial unloading over the past several thousand years using GIA coefficients from Ivins et al. [2013] regional ice deglaciation model. We smooth the corrected GRACE spherical harmonics using a 250 km radius Gaussian averaging function [Jekeli, 1981], and we generate regular latitude-longitude monthly ice mass grids (Equation 1.10). We use the grids to calculate the linear trend in a least squares regression simultaneously fitting annual and semiannual signals [Velicogna and Wahr, 2006a, Wahr et al., 1998] to obtain a digital map of longterm ice mass balance for the ASE (Figure 4.1a).

We generate time series of ice mass balance for ASE by applying the least squares mass concentration (mascon) approach described in Velicogna et al. [2014] to the Antarctic Ice Sheet [Chapter 3]. To do this, we cover the entire ice sheet with a set of equal-area mascons (Figure 4.3). Each mascon is a 3° diameter spherical cap with a mass equal to a uniformly-distributed centimeter of water [Farrell, 1972, Sutterley et al., 2014, Chapter 2]. For each mascon, we calculate a set of Stokes coefficients, which we smooth with a 250 km Gaussian function and convert into mass. We simultaneously fit the mascon Stokes coefficients to the monthly GIA-corrected GRACE coefficients to obtain estimates of the monthly mass variability for each mascon. This procedure retrieves scaled estimates of regional ice mass variation at each time step. We calculate the mass anomaly time series, $M(t)$, for the ASE through summation of the regional mascons (Figure 4.3). To calculate $dM/dt$, we first smooth the mass anomaly time series to remove annual variations, and then calculate the derivative over 13-month windows using a Savitzky-Golay filter [Velicogna, 2009, Savitzky and Golay, 1964]. Uncertainty in the GRACE estimates of ice mass changes are a combination of GRACE measurement error, leakage error, GIA uncertainty, and statistical uncertainty. Errors are calculated as
described in Velicogna et al. [2014] and Chapter 3.

Ice mass balance from the mass budget method (MBM) is calculated combining estimates of ice discharge (D) with surface mass balance (SMB) for each drainage basin as in [Rignot et al., 2011b]. For ice discharge, we use the measurements provided in Mouginot et al. [2014] combining ice motion measurements from synthetic aperture radar (SAR) and ice thickness from the Rignot et al. [2014]. For SMB, we use the monthly products calculated from a 1979–2013 climate simulation of RACMO2.3 [Ligtenberg et al., 2013, van Wessem et al., 2014]. Field data have been used to estimate the RACMO absolute precision [van de Berg et al., 2006]. In the Antarctic, the uncertainty (1σ) in SMB over grounded ice averages 7% or 144 Gt/yr [Lenaerts et al., 2012]. In ASE, the uncertainty in SMB increases to 14.8% or 28 Gt/yr [Rignot et al., 2008a]. RACMO2.3 products are available through December 2013. To compare the results with GRACE, the rates of ice discharge are linearly interpolated into a set of monthly fluxes assuming that seasonal variations in regional ice velocity are minimal, which has been verified over short time periods. We calculate the monthly $dM/dt$ time series by subtracting $D$ from SMB. We generate a time series of monthly mass anomalies, $M(t)$, from the MBM by subtracting monthly rates of ice discharge from monthly rates of SMB, and then calculating the cumulative sum of the time series.

We use along-track repeat Envisat radar altimetry measurements from the Laboratoire d’Etudes en Géophysique et Océanographie Spatiales (LEGOS) at the Centre National de la Recherche Scientifique (CNRS) [Flament and Rémy, 2012]. The along-track altimetry technique increases the number of processed data points on the Antarctic ice sheet compared to the traditional crossover analysis. We use 83 cycles of 35-day repeat orbits retrieved over the period September 2002 to October 2010. Relative surface elevations are calculated for 500 m radius disks using a least squares algorithm which simultaneously solves for radar waveform properties, along-track slope, cross-track slope, regional surface curvature assuming a quadratic shape, and the elevation time series [Flament and Rémy, 2012, Rémy and Parouty,
The waveform properties are computed using the ice sheet-optimized ICE-2 retracking algorithm, which solves for leading edge amplitude, leading edge width, trailing edge slope, waveform backscatter coefficient and the corrected range [Legrésy et al., 2005]. Additional corrections are included to account for the varying electromagnetic properties of the ice sheet surface [Flament and Rémy, 2012, Rémy and Parouty, 2009]. Seasonal variations in radar penetration due to snowpack properties may still account for part of the seasonal signal in ice sheet elevation. To compare with GRACE, we use the Envisat along-track elevation time series to build monthly 25 km² grids for the same dates used by the GRACE fields when the Envisat data are available on the 35-day repeat orbit. Error estimates for each grid point are calculated as described in [Flament and Rémy, 2012]. Figure 4.1c shows the map of surface elevation change from Envisat for the period September 2002 to October 2010. We use the two-step smoothing and Savitzky-Golay differentiation procedure previously described in the GRACE analysis to calculate the $dV/dt$ time series.

We also use elevation measurements from ICESat-1, Operation IceBridge (OIB) Airborne Topographic Mapper (ATM) and Land, Vegetation and Ice Sensor (LVIS) to quantify the surface elevation change. Our ICESat measurements are Release-33 of the GLA12 Antarctic and Greenland Ice Sheet Altimetry data provided by the National Snow & Ice Data Center (NSIDC) [Zwally et al., 2012]. We remove cloud-affected data points following the methods described in Howat et al. [2008], Pritchard et al. [2009], Smith et al. [2009], Sørensen et al. [2011], and Chapter 1. Elevation changes are calculated in reference to the WGS-84 ellipsoid, corrected for saturation effects with the GLA12 correction product [Zwally et al., 2012], and for Gaussian-Centroid (G-C) offset [Borsa et al., 2014]. OIB ATM, and LVIS data products are used as additional constraints to the surface shape and elevation time series [Krabill, 2010, Blair and Hofton, 2010]. We use a least squares approach to simultaneously solve for the elevation time series and surface shape (e.g., along-track and cross-track slope) of 1 km surface patches [Schenk and Csatho, 2012, Chapter 1]. This is the first published use of a method similar to Schenk and Csatho [2012] for combining satellite and airborne
lidar measurements to calculate surface elevation change in Antarctica. The OIB aerial laser altimetry datasets greatly increase the total number and spatial coverage of elevation data points within each surface patch [Schenk and Csatho, 2012, Rezvanbehbahani, 2012]. Assuming that the surface shape of each patch does not change in time, the airborne altimetry data helps separate the spatial and temporal components of elevation change within each patch [Schenk and Csatho, 2012, Chapter 1]. For the temporal component, a low-order polynomial is chosen to reduce the impact of annual variations, which may not be captured in the two to three campaign acquisitions per year. Errors for each time-step are calculated propagating the regression fit error as described in Schenk and Csatho [2012], and the GIA uplift uncertainty. From our reconstructed centroid time series, we calculate interpolated maps of relative surface elevation using inverse multiquadric radial basis functions and calculate elevation change maps by differentiating sets of interpolated elevation maps [Hardy, 1971]. Figure 4.1d shows the map of surface elevation change from ICESat/OIB for the period 2003–2009.

In order to convert the surface elevation measurements from Envisat and ICESat/OIB into ice mass, we apply a simple density conversion assuming that the surface changes in areas of fast flow (speed greater than about 50 m/yr) are entirely due to ice dynamics, i.e., are taking place at a density of 900 ± 20 kg/m³ [Shepherd et al., 2012]. This assumption is justified by the fact that changes in surface elevation (Figure 4.1c-d) are strongly correlated with the changes in speed (Figure 4.1b), not with changes in SMB. In slower-moving regions, changes in surface elevation are assumed to be dominated by changes in SMB rather than ice dynamics (the latter is also not observable over the entire domain). Over the slow-moving interior, we employ a density of 550 ± 250 kg/m³ (a 45% uncertainty). Our final error combines the errors from both regions.
4.3 Results

Figure 4.1 shows the map of GRACE ice mass trend for January 2003 to May 2014, the change in ice flux density (product of ice velocity by ice thickness) between 1996 and 2008 combining changes in ice motion from SAR with ice thickness from BEDMAP2 [Fretwell et al., 2013], Envisat radar altimetry $dH/dt$ for 2002–2010 and ICESat, OIB and LVIS $dH/dt$ for 2003–2009. GRACE trend (Figure 4.1a) shows a significant mass loss in the region, with a loss per unit area exceeding 20 cm water equivalent per year. The limited spatial resolution of the GRACE data (∼350 km) compared to the size of the glaciers limits the interpretation of the spatial pattern of ice mass change.

The map of flux density change (Figure 4.1b) highlights the speed up of all glaciers in the region: PIG, THW, SMI, KOH, POP and HAY [Mouginot et al., 2014]. The maps of surface elevation change from Envisat (Figure 4.1c) and ICESat/OIB (Figure 4.1d) indicate that ASE is dominated by ice thinning. The rate of thinning is higher in regions of fast flow, as denoted by the velocity magnitude contours from Rignot et al. [2011a], and areas of larger change in flow speed in Figure 4.1b. We find thinning of PIG propagating upstream, broad thinning of THW, and significant thinning of the smaller HAY, SMI, POP and KOH. The rates of surface thinning exceed several meters per year in the areas of fast flow, and the spatial pattern of thinning is consistent with the pattern of ice velocity change. This suggests that the pattern of thinning is due to changes in ice dynamics instead of changes in SMB.

The inherent differences in spatial resolution and temporal coverage between datasets makes it difficult to compare the spatial patterns of the mass balance results. We subsequently focus on the basin-scale assessments of mass balance. During the entire time period (1992–2013), variations in SMB modulate the $dM/dt$ time series significantly (Figure 4.2a). The average SMB for this time period is $185 \pm 26 \text{ Gt/yr}$, yet interannual variations in SMB up to 150 Gt are not uncommon in the GRACE, Envisat and MBM time series (Figure 4.2b).
During the common period, 2003–2009, GRACE, MBM, and Envisat mass balance time series, $dM/dt$, are in good agreement in terms of total magnitude and timing of the cyclic oscillations (Figure 4.2b). The three $dM/dt$ time series agree within ±13%. Over the same period the GRACE, Envisat and MBM time series of cumulative mass anomaly, $M(t)$, agree within ±5% (Figure 4.2c).

The ICESat/OIB time series does not capture interannual variations because of its low temporal sampling (two to three measurements per year). Over the period common to all four techniques, 2003–2009, however, we find an excellent agreement with all the other techniques in terms of average mass balance and acceleration in mass balance. We find a total mass loss of 81 ± 16 Gt/yr for ICESat/OIB, 90 ± 8 Gt/yr for GRACE, 89 ± 7 Gt/yr for MBM, and 74 ± 8 Gt/yr for Envisat. During the same period, the acceleration in loss are, respectively, 13.8 ± 9.6 Gt/yr² for ICESat/OIB, 16.4 ± 2.6 Gt/yr² for GRACE, 19.6 ± 3.4 Gt/yr² for MBM, and 15.5 ± 3.6 Gt/yr² for Envisat (Table 4.1).

We calculate a reconciled mass balance for the ASE during 2003–2009 as a linear average of the individual estimates from GRACE, MBM, Envisat and ICESat/OIB and the associated error as the sum in quadrature of each technique error. We find a rate of mass loss of 84 ± 10 Gt/yr with an average acceleration of 16.3 ± 5.6 Gt/yr². For comparison, over the entire period of 1992–2013, the mass loss of ASE as determined by the MBM averages 83 ± 5 Gt/yr with an acceleration of 6.1 ± 0.7 Gt/yr², or almost 3 times less than in the more recent period.

Over the time period 2003–2011, we have coincident datasets from ICESat/OIB, GRACE and MBM. We find a mass loss of 95 ± 14 Gt/yr for ICESat/OIB, 104 ± 7 Gt/yr for GRACE, and 105 ± 6 Gt/yr for MBM. During the same time period, the acceleration in mass loss is 13.8 ± 6.3 Gt/yr² for ICESat/OIB, 15.5 ± 1.7 Gt/yr² for GRACE, and 18 ± 2.3 Gt/yr² for MBM (Table 4.1). This period includes the ICESat period (2003–2009) and the period of yearly OIB campaigns (2009–2011) when ICESat is no longer available. During 2003–2011,
our reconciled estimate from GRACE, MBM and ICESat/OIB is an average mass loss of
102 ± 10 Gt/yr with an acceleration of 15.7 ± 4.0 Gt/yr², which is not significantly different

4.4 Discussion

The excellent agreement and high correlation between independent time series from GRACE,
MBM, Envisat, and ICESat/OIB during their common period significantly increases con-
fidence in the various analyzed techniques. The coincidence in magnitude and temporal
oscillations of the time series provides a significant cross-validation of the techniques at the
regional scale in a glaciological setting where mass changes are significant. The agreement
within the confidence intervals confirms that the error estimates for the different techniques
are realistic. Beyond the end of the MBM, Envisat, and ICESat/OIB record (mid-2012),
the ongoing GRACE time series of $M(t)$ measurements (Figure 4.2c) indicate that the mass
loss of the ASE is continuing at the same rate after 2012 until the middle of year 2014 which
is the end of our current GRACE record. As new SMB estimates are produced and longer
time series of OIB laser data are acquired, we will extend the duration and quality of the ice
sheet mass balance record in the region.

Our mass balance numbers are within the error estimates of the recent CryoSat-2 estimates
from [McMillan et al., 2014] who report a mass loss of 120 ± 18 Gt/yr for the time period
2010–2013 for basins 21 and 22, the equivalent of ASE in this study. For comparison, we
calculate an average mass loss of 144 ± 7 Gt/yr from the linear average of GRACE and MBM
in 2010–2013. The lower number from CryoSat-2 is likely due to the variability in firn depth
and snowfall affecting the short-term (3-year) Cryosat-2 time series. Overall, however, the
mass balance estimates agree within confidence intervals.

Although the ICESat/OIB mass balance time series do not capture interannual variations in
ice elevation, our results suggest that it correctly captures the total change in mass balance of the ASE. Hence, campaign-style measurements by OIB, combined with the longterm reference from ICESat, are sufficient to extend the time series of laser altimetry data in time and maintain a consistent record of ice mass balance in the region. In our estimate, we only use the OIB data within 1 km of the ICESat tracks. The elevation change results could be improved by including additional ATM and LVIS tracks acquired in the region since 2002, but the statistical analysis would become significantly more complex. With our approach, we confirm that the resulting numbers are already consistent with those obtained with more comprehensive, complementary, independent MBM and GRACE techniques.

In the ASE, the choice of the GIA correction only minimally impacts the GRACE mass balance estimates. Here we use the Ivins et al. [2013] regional GIA model. Using any of the other available GIA changes the mass balance numbers by 8% or 9 Gt/yr for the Whitehouse et al. [2012] model and 2% or 2 Gt/yr for the A et al. [2013] global model based on ICE-5G ice history [Peltier, 2004]. These errors are within the uncertainty bounds of the reconciled estimates. Results using GIA coefficients from the Whitehouse et al. [2012] deglaciation model are shown in Figure 4.4.

The Envisat and ICESat/OIB results fall within the error estimates of the GRACE and MBM time series when changes in surface elevation in areas of fast flow (speed greater than about 50 m/yr) are assumed to be due to ice dynamics and taking place at the density of ice, here $900 \pm 20 \text{ kg/m}^3$, and changes in slower-moving regions are assumed to be dominated by changes in SMB (rather than ice dynamics) and to occur at $550 \pm 250 \text{ kg/m}^3$. Overall most changes in mass in the ASE occur at low elevation (97% of the loss is contained below 1,300 m elevation) and at high speed (87% of the loss for areas flowing above 50 m/yr), where changes are very likely to take place at the density of ice. We note that Envisat measurements may miss some of the coastal region due to loss of signal along the edges of the ice sheet typical of satellite radar altimeters.
The ASE receives high rates of snowfall compared to the average in Antarctica [Lenaerts et al., 2012, Medley et al., 2014]. In 1992–2013, the RACMO2.3 SMB averages $185 \pm 26$ Gt/yr for the ASE. SMB varies significantly over short time scales ($\sigma = 27$ Gt/yr in 2002–2013). Over the entire 22 year period, however, changes in SMB are negligible, $-0.2 \pm 0.3$ Gt/yr$^2$ (Figure 4.2a). Our study confirms that multidecadal periods of observation are needed to determine the longterm trend in ice mass balance and its acceleration and to minimize the impact of firn compaction on altimetry results [Rignot et al., 2011b, Shepherd et al., 2012, Wouters et al., 2013]. Similarly, it is difficult to evaluate the exact partitioning between SMB and ice dynamics over short periods. For example, the partitioning in mass balance over 2003–2009 does not reflect the partitioning over 1992–2013.

The longer MBM record and its comparison with independent techniques provides evidence that the increase in regional mass loss is caused almost entirely by changes in ice velocity. The longterm (1992–2013) change in SMB ($-0.2 \pm 0.3$ Gt/yr$^2$) is small compared to the change in ice discharge ($+5.7 \pm 0.4$ Gt/yr$^2$) (Figure 4.2a). The SMB fluctuations modulate the yearly mass balance, yet never mask out the trend in dynamic loss in the region. Most of the ice mass loss took place in the past decade. The cumulative loss of $1160 \pm 30$ Gt during the 2002–2013 GRACE period is 71% of the total loss of $1630 \pm 30$ Gt for the 1992–2013 period. The losses correspond to equivalent rises of global sea level of $3.2 \pm 0.1$ mm and $4.5 \pm 0.1$ mm for the 2002–2013 and 1992–2013 periods respectively (Figure 4.2c).

4.5 Conclusion

In this study, we provide the first quantitative comparison of the ice sheet mass balance of the Amundsen Sea Embayment (ASE) between four independent geodetic techniques. We find an excellent agreement in mass loss and acceleration in mass loss from these independent techniques during common periods at the regional scale in a sector that dominates the mass
balance of the continent. In this study, we develop a new method for combining satellite and 
airborne lidar measurements to calculate surface elevation change in Antarctica based on 
Schenk and Csatho [2012]. This methodology expands the spatial and temporal coverage by 
synergistically using all available laser altimetry observations. We show that OIB campaign 
style measurements are sufficient to extend the time series of mass balance estimates using 
ICESat laser altimetry data in time and maintain a record of ice mass balance in the region. 
We also show that the significant fluctuations in SMB observed over short periods average 
out after a couple of decades. Our comprehensive record, evaluated from multiple techniques, 
of mass loss in West Antarctica, shows a tripling in mass loss in recent years with respect 
to the entire analyzed period 1992–2013. The rapid rate of convergence of the independent 
techniques examined indicates that the measurements have now reached maturity and may 
be used with increased confidence for glaciological interpretation and inclusion in ice sheet 
numerical models with data assimilation methods.

<table>
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<tr>
<th>Dataset</th>
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<th>Mass Balance [Gt/yr]</th>
<th>Change in Mass Balance [Gt/yr²]</th>
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<td>−90 ± 8</td>
<td>−16.4 ± 2.6</td>
</tr>
<tr>
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<td>−89 ± 7</td>
<td>−19.6 ± 3.5</td>
</tr>
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<td>−15.5 ± 3.6</td>
</tr>
<tr>
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<td>−81 ± 16</td>
<td>−13.8 ± 9.6</td>
</tr>
<tr>
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<td>−15.5 ± 1.7</td>
</tr>
<tr>
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<tr>
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<td>1992 – 2013</td>
<td>−83 ± 5</td>
<td>−6.1 ± 0.7</td>
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</table>

Mean mass balance and change in mass balance calculated for each period simultaneously using a weighted least squares regression from the mass balance time series, $M(t)$ and $dM(t)/dt$. 

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Figure 4.1: (a) Ice mass trend estimated using GRACE time-variable gravity in centimeters of water equivalent. The red contour delineates 0 cm yr$^{-1}$. (b) Change in ice flux density between 1996 and 2008 combining velocity changes from Mouginot et al. [2014] and ice thickness from Rignot et al. [2014]. (c) Elevation change estimated using repeat Envisat radar altimetry from Flament and Rémy [2012]. The dashed line represents the polar extent of the Envisat orbit (81.5$^\circ$ S). (d) Elevation change from laser altimetry combining ICESat-1 (GLAS) with Operation IceBridge (ATM and LVIS). Contours on (c) and (d) denote surface ice speeds of 50 (gray), 125 (red), 250 (blue), and 500 (green) m/yr from Rignot et al. [2011a]. Plots are overlaid on a 2003–2004 MODIS mosaic of Antarctica [Haran et al., 2013].
Figure 4.2: (a) Rates of RACMO surface mass balance, SMB (blue) and ice discharge, D, from Mouginot et al. [2014] (black). (b) Mass balance estimates, $dM(t)/dt$, and (c) cumulative mass anomalies, $M(t)$, for the Amundsen Sea Embayment (ASE) of Antarctica from the Mass Budget Method, MBM, (black), GRACE time-variable gravity (red), Envisat radar altimetry (green) and ICESat/IceBridge laser altimetry (orange).
**Figure 4.3:** Mascons used for the Antarctic Ice Sheet. Each disc represents a single mascon. Red disks define the Amundsen Sea Embayment mascons for which the GRACE time series is calculated. Thick black line denotes the basin boundary.

**Figure 4.4:** Cumulative mass anomalies for the ASE from the Mass Budget Method (black) between 2002–2013, and GRACE time-variable gravity between 2002–2014 corrected with GIA coefficients from Ivins et al. [2013] (red) and Whitehouse et al. [2012] (blue).
Chapter 5

Improved Greenland surface elevation measurements from satellite and airborne laser altimeters

Abstract

We present improved estimates of Greenland Ice Sheet (GrIS) surface elevation change from a novel combination of satellite and airborne laser altimetry measurements. Our method combines measurements from the Airborne Topographic Mapper (ATM), the Land, Vegetation and Ice Sensor (LVIS) and ICESat-1 to generate elevation change rates at high spatial resolution. The combination method extends the records of each instrument, increases the overall spatial coverage compared to a single instrument, and produces high-quality, coherent maps of surface elevation change. We validate the combination of ATM and LVIS for periods with near term overlap of the two instruments. With our observations, we extend the record of thinning rates of Jakobshavn Isbræ, the fastest flowing and most active glacier in Greenland. We find that the glacier has thinned approximately 90m near the ice front since 2002, and 120m since 1993. We assess the accuracy of the surface mass balance (SMB) outputs and the contribution of SMB to regional thinning by comparing our elevation results with SMB results from two regional climate models, the Regional Atmospheric Climate Model
(RACMO) and the Modèle Atmosphérique Régional (MAR). At one study site in northwest Greenland, we find that the MAR model has better correspondence with the elevation data over the RACMO model primarily due to differences in the characterization of meltwater refreezing.

5.1 Introduction

Airborne laser altimetry surveys of the Greenland Ice Sheet (GrIS) have provided evidence of regional surface elevation change since they were first conducted in the early 1990’s [Krabill et al., 2004]. Here, we combine measurements from satellite and airborne laser altimeters to quantify the surface elevation change of Greenland at high spatial resolution. We use 19 years of elevation measurements from the Airborne Topographic Mapper (ATM), 4 years of elevation measurements from the Land, Vegetation and Ice Sensor (LVIS) and 6 years of elevation measurements from ICESat-1. Our elevation measurements are co-registered on a shot-by-shot basis using triangulated irregular networks (TIN’s), which account for variations in the slope of the ice sheet surface [Pritchard et al., 2009]. For validation purposes, we compare our lidar combination technique with the standard NSIDC idhdt4 elevation product calculated with solely ATM measurements [Krabill, 2010].

We use the results from our method to explore two separate case studies. The first study is an investigation of the recent thinning of Jakobshavn Isbræ, a major West Greenland glacier losing an average of 30 Gt/yr to the sea [Thomas et al., 2003, Howat et al., 2011]. The second study is a comparison with surface mass balance outputs from two regional climate models at locations near weather stations on the ice sheet [Noël et al., 2015, Fettweis, 2007]. In the following sections we discuss how the method is implemented, how we use the results to remark on the changes of Jakobshavn Isbræ and assess surface mass balance outputs, and conclude on the significance of our two case studies.
5.2 Data and Methods

Our ICESat-1 measurements are Release-33 of the GLA12 Antarctic and Greenland Ice Sheet Altimetry data provided by the National Snow & Ice Data Center (NSIDC) [Zwally et al., 2012]. As the accuracy of the ICESat elevation measurements is degraded by the presence of clouds, we remove cloud-affected data points using a set of culling criteria following Howat et al. [2008], Pritchard et al. [2009], Smith et al. [2009], and Sørensen et al. [2011]. The ICESat elevation measurements are converted to be in reference to the WGS-84 ellipsoid, corrected for saturation effects with the GLA12 correction product [Zwally et al., 2012], and corrected for Gaussian-Centroid (G-C) offset [Borsa et al., 2014]. Our airborne lidar measurements are Level-2 Airborne Topographic Mapper (ATM) and Land, Vegetation and Ice Sensor (LVIS) datasets provided by NSIDC [Krabill and Thomas, 2010, Krabill, 2010, Blair and Hofton, 2010]. ATM is a conically scanning lidar which has flown extensively in Greenland since 1993, and was developed at the NASA Wallops Flight Facility (WFF) [Krabill et al., 2000]. LVIS is a large-swath scanning lidar which flew in Greenland from 2007 to 2013, and was developed at NASA’s Goddard Space Flight Center (GSFC) [Blair et al., 1999, Hofton et al., 2008]. The Level-2 LVIS data supplies 3 different elevation surfaces calculated from the Level-1B full waveform product: the lowest returning surface, the highest returning surface, and the centroid surface [Blair and Hofton, 2010]. Here, we use the lowest returning surface when the waveform resembles a single-peak gaussian, and the centroid surface when the waveform is multi-peak.

Our technique derives rates of elevation change by comparing a set of measured elevation values with a set of interpolated elevations from a different time period. The method follows the altimetry analysis of Pritchard et al. [2009] and Rignot et al. [2013], but with the addition of the higher resolution airborne altimetry datasets. The set of interpolated values are constructed using triangulated irregular networks (TIN’s) based on Delaunay triangulation.
[Pritchard et al., 2009, 2012]. For each data point in a flight line, a set of Delaunay triangles is constructed from a separate flight line using all data points within 300 meters from the original point. If the original point lies within the confines of the Delaunay triangulation convex hull, the triangular facet housing the original point is determined using a winding number algorithm. The new elevation value is calculated using barycentric interpolation with the elevation measurements at the three triangle vertices. The elevation at each vertex point is weighted in the interpolation by the area of the triangle created by the enclosed point and the two remaining vertices (Figure 5.1). Assuming that the ice sheet surface is not curved over the scale of the individual triangular facet, interpolating to the original coordinates will compensate for slopes in the ice sheet surface.

Any crevassed terrain, snow drifts or low-lying clouds will contaminate the lidar elevation values for the interpolation. In order to limit the effect of contaminated points, elevation change rates are filtered using an interquartile range algorithm [Pritchard et al., 2009]. The method produces elevation change rates along flight lines with very high spatial sampling (∼10–50 m), much higher than rasterization methods (e.g. 500 m grids in Kjeldsen et al. [2013]). We validate the combination of ATM and LVIS for periods with near term overlap of the two instruments. Maps of elevation change are constructed by combining flight lines for common year ranges. Elevations are compared using only the spring campaigns to limit the effect of seasonal fluctuations in elevation.

We compare with the Pre-Icebridge and Operation IceBridge ATM Level-4 Surface Elevation Rate of Change product provided by NSIDC [Krabill, 2010]. This elevation change product computes rates directly from the Level-1B Qfit elevation product. To compare the two methods, we rasterize the estimates of elevation change rates from both techniques into 5km by 5km polar stereographic images. We then compare the spatial patterns and average elevation differences between the two techniques.

We use monthly mean components of surface mass balance (SMB) calculated from 1958–2014
climate simulations by the Regional Atmospheric Climate Model (RACMO2.3) and Modèle Atmosphérique Régional (MAR 3.5.2) [Noël et al., 2015, Fettweis, 2007]. Surface mass balance is the sum of mass accumulation from snow and rain minus surface ablation from meltwater runoff, sublimation, and wind scour [Ettema et al., 2009]. Cumulative anomalies in SMB and the individual SMB components are calculated in reference to a period of near total net balance, 1961 to 1990 [van den Broeke et al., 2009, Rignot et al., 2008b]. To compare with the altimetry estimates, we interpolate the climate model outputs to locations with multiple years of lidar measurements near 3 surface observations sites: NE Promice (79.9°N, 24.6°W), the K-transect (67.1°N, 50.0°W) and Camp Century (77.2°N, 61.3°W). Each resultant SMB time series is smoothed using a 13-month lowess filter to remove the effects of seasonal variability [Velicogna, 2009].

5.3 Results and Discussion

Figure 5.2 shows a comparison of the triangulated data with the standard NSIDC (idhdt) product for the Figure 5.2a-b) high melt years covering Spring 2011 to Spring 2013 and Figure 5.2c-d) high accumulation season covering Spring 2013 to Spring 2014. The results demonstrate that the triangulation technique produces coherent measurements of surface elevation change. The addition of LVIS data increases the spatial coverage of elevation measurements. This is most evident in southwest Greenland for the 2011–2013 period map. Agreement between ATM and LVIS for near concurrent acquisitions (less than 10 days apart) is 24 cm, which suggests that the two lidars are compatible and complementary. The RMS difference between the idhdt product and the triangulated elevations is 68 cm, which is less than the combined average error of both datasets (109 cm). For most regions across the ice sheet, the triangulated elevation change rates are smoother across flight lines when compared to the idhdt product. The comparison gives confidence in the value of combining laser altimetry datasets and the overall quality of the technique.
5.3.1 Jakobshavn Isbræ

Figure 5.3 shows rates of elevation change covering periods from 1993 onward for the center west (CW) basin of Rignot and Mouginot [2012]. Between 1993 and 1998 (Figure 5.3a), the surface elevation at Jakobshavn Isbræ increased close to the ice front while interior regions showed patterns of both marginal thinning and thickening as described in Thomas et al. [2003]. Between 1998 and 2009 (Figure 5.3b), the glacier developed into a state of overall thinning with sporadic points in the interior showing slight thickening. Total spatial coverage increased greatly after the onset of Operation IceBridge in 2009. For the periods after the ICESat-1 mission (Figure 5.3c-h), Jakobshavn has experienced years of very strong thinning (2010 and 2012) and a year of more moderate thinning (2013). As a consequence of the record 2012 melt season, thinning between 2012 and 2013 reached deep into the interior of the ice sheet as shown in Figure 5.3f. Coastal areas peripheral to Jakobshavn Isbræ thickened between the 2013 and 2014 spring campaigns. From 2014 to 2015 (Figure 5.3h), the glacier thinned while interior regions showed patterns of concurrent thinning and thickening.

The triangulation technique co-registers all near overlap data to single flight lines, which allows for the comparison of relative elevations through time without any further interpolation. The relative elevation profiles for a 22 year period along the main trunk of Jakobshavn Isbræ are shown in Figure 5.4. This extends the Jakobshavn Isbræ relative elevation plot of Thomas et al. [2003] by 14 years. The elevations are shown relative to a 2002 flight line and with respect to the glacier’s 2000–2001 terminus position from Moon et al. [2015]. Data from early 1990’s flight lines are highly clustered due to the weak elevation change during the period [Thomas et al., 2003]. For periods after the onset of elevation change, the thinning is greatest close to the ice front and propagates far inland. The major 2012 melt season is particularly evident due to the larger than average elevation difference between the 2012 and 2013 lines compared to other years. The 2013 and 2014 melt seasons had weaker rates of thinning compared to the 2012 season as shown by the decreased differences in relative
elevation lines. Overall, elevation loss since the publication of Thomas et al. [2003] has continued unabated with the glacier thinning approximately an additional 90m since 2002. Overall, the glacier has thinned approximately 120m near the ice front since 1993.

5.3.2 Surface Mass Balance

Relative elevation measurements through time can also be compared at specific points on the ice sheet. Here, we compare between elevation measurements and outputs from surface mass balance models at three locations: near the northeast Promice (Program for Monitoring the Greenland Ice Sheet) weather station (Figure 5.5a), near Site-5 of the University of Utrecht K-Transect in west Greenland from van de Wal et al. [2012] (Figure 5.5b), and near Camp Century in northwest Greenland (Figure 5.5c). The exact locations are chosen to maximize the total number of elevation measurements in time. For all three locations, the elevation change should be governed by changes in surface mass balance. The Promice and K-transect locations are also deep in their respective ablation zones. For the three sites, the elevation measurements are converted to mass using a simple density conversion ($\rho_{\text{ice}}$: 917 kg/m$^3$). The three primary surface mass balance components of precipitation, runoff and sublimation for both models are shown in Figure 5.6. Precipitation is shown in the top row (a-c), runoff is shown in the middle row (d-f) and sublimation is shown in the bottom row (g-i). The two runoff subcomponents (snowmelt and meltwater refreeze) are shown in Figure 5.7.

We find that the SMB estimates correspond well with the altimetry measurements over long periods, but some short term oscillations are missed. Deficiencies over short time periods could be due to SMB errors, the spatial resolution of the SMB models compared to point measurements, or errors in the density conversion from elevation change to mass. At the NE Promice site (Figure 5.5a), altimetry measurements correspond well with SMB outputs from the MAR model for all time periods, and with the RACMO model after 1999. At this site, the longterm surface mass balance change is primarily dictated by changes in meltwater
runoff (Figure 5.6). The primary difference between the RACMO and MAR models at this site over both annual and interannual scales is the meltwater refreezing subcomponent of runoff as shown in Figure 5.7. At the K-transect location (Figure 5.5c), the altimetry data shows a period of relative stability in the 1990’s not shown in the SMB datasets. The increase in thinning after 2002 is well captured by both models. At this site, meltwater runoff is the primary surface mass balance component dictating the longterm change in SMB (Figure 5.6). At the Camp Century site (Figure 5.5c), the altimetry shows a spiked increase in mass occurring two years before the SMB data for both models. Both models correspond well with the altimetry estimate before the mass increase. At this site, the longterm surface mass balance change is primarily dictated by changes in precipitation as all snowmelt is balanced by internal refreezing (Figures 5.6 and 5.7).

5.4 Conclusion

In this study, we develop a new method to based on Pritchard et al. [2009] and Rignot et al. [2013] to combine multiple laser altimetry datasets. This is the first effort to combine the laser altimetry datasets using the triangular irregular networks method. We use this method to quantify the surface elevation change of Greenland at high spatial resolution. The combination of multiple laser altimetry datasets increases the total spatial coverage of elevation measurements compared to a single instrument. The triangulation technique produces highly coherent estimates of surface elevation change at unprecedented spatial resolution for multiple instruments. We validate our combination method by comparing with the standard NSIDC elevation change product calculated using overlapping Level-1B ATM data. Elevation measurements from each instrument are tested for periods of nearly concurrent acquisitions to ensure that the lidars are compatible.

We presented on two case studies: the thinning of Jakobshavn Isbræ, and a novel comparison
with surface mass balance outputs. Our results show large changes in elevation of Jakobshavn Isbræ and a propagation of thinning into the interior since the late 1990’s. Jakobshavn Isbræ thinning rates during the major melt season of 2012 were the strongest of any year since surveys began. The surface mass balance outputs from RACMO and MAR regional climate models show good correspondence with mass changes derived from surface elevation changes over long periods. The MAR model has better correspondence with the altimetry estimate at the NE Promice site for the entire period. The differences at this location are primarily due to the characterization of how snowmelt water refreezes within the ice sheet.

Our combination method takes advantage of the fact that scanning airborne lidars can be triangulated for each day of acquisition, which enables fast and efficient processing of new datasets as they become available. As new Operation IceBridge and eventually ICESat-2 laser data are acquired, we will extend the duration of the multi-instrument thinning rate maps for the ice sheet. The method can be used to investigate seasonal elevation changes combining data from spring and fall IceBridge campaigns. The approach can also be used as an independent metric in the Calibration and Validation stage of the ICESat-2 mission.
Figure 5.1: Triangulated mesh formulated around a 1994 ATM flight line point using points from a separate ATM flight line. P1, P2 and P3 represent the three vertices of the triangle housing the original ATM point (denoted by the red star). Elevation values at each vertex point are weighted in the interpolation by their respective areas, A1, A2 and A3.
Figure 5.2: a-b) Rasterized 5km polar stereographic images of (a) idhdt4 and (b) triangulated elevation change rates for high melt years covering Spring 2011 to Spring 2013. c-d) Rasterized 5km polar stereographic images of (c) idhdt4 and (d) triangulated elevation change rates for high accumulation season covering Spring 2013 to Spring 2014. Plots overlaid on an image mosaic from the Greenland Ice Mapping Project [Howat et al., 2014].
Figure 5.3: a) Rates of elevation change for the early ATM period (1993–1998). b) Rates of elevation change from the 1998 ATM campaign to the final year of ICESat-1 (1998–2009). c-h) Rates of elevation change for successive years from Operation IceBridge. Plots overlaid on an image mosaic from the Greenland Ice Mapping Project [Howat et al., 2014].
Figure 5.4: Relative elevation profiles of Jakobshavn Isbræ compared to a 2002 ATM survey. Position of the 2000–2001 glacier terminus provided by Moon et al. [2015]. The inset map shows in red the location of the segment used from the 2002 ATM flight line.

Figure 5.5: Comparisons between Operation IceBridge laser altimetry (black), RACMO2.3 surface mass balance (blue), and MAR3.5.2 surface mass balance (red) near three sites: a) NE Promice weather station, b) Site-5 of the K-Transect, and c) Camp Century. Thick red and blue lines denote 13-month smoothed versions of the SMB time series [Velicogna, 2009]. Inset maps denote the locations of the sites with a red star.
**Figure 5.6:** Comparisons between RACMO2.3 climate model output products (blue), and MAR3.5.2 climate model output (red) near three sites: a,d,g) NE Promice weather station, b,e,h) Site-5 of the K-Transect, and c,f,i) Camp Century. a-c) shows the precipitation component, d-f) shows the meltwater runoff component and g-i) shows the sublimation component. Inset maps denote the locations of the sites with a red star.
Figure 5.7: Comparisons between RACMO2.3 runoff subcomponents (blue), and MAR3.5.2 runoff subcomponents (red) near three sites: a,d) NE Promice weather station, b,e) Site-5 of the K-Transect, and c,f) Camp Century. a-c) shows the snowmelt subcomponent and d-f) shows the refreezing and meltwater retention subcomponent. Inset maps denote the locations of the sites with a red star.
Chapter 6

Conclusions

This research work is an effort to improve our understanding of regional ice sheet change using a combination of airborne, satellite and climate model outputs. The first two studies we present as part of this dissertation use a method of regional ice sheet mass balance from the Gravity Recovery and Climate Experiment (GRACE) to examine uncertainties, and to contribute towards a better understanding of the processes dominating regional ice sheet change. The third study is a regional study comparing the GRACE mass balance method with three independent techniques at a key sector of Antarctic mass change. The final study focuses on surface elevation changes in Greenland using a novel combination of satellite and airborne laser altimeters. This work, in summary, provides new approaches for observing regional ice sheet change using multiple independent datasets, and investigates the mechanisms driving the modern-day change.

6.1 Summary of Results

The first study within this dissertation (Chapter 2) evaluates glacial isostatic adjustment (GIA) rates in Greenland by comparing GIA-corrected GRACE mass trends with independent altimetry and surface mass balance (SMB) datasets. GIA is a major contaminating signal in GRACE-based observations of ice sheet change. Determining which GIA correc-
tions are valid should reduce the overall spread of ice sheet mass balance estimates from GRACE. We evaluate glacial isostatic adjustment rates from A et al. [2013], Simpson et al. [2009] and Wu et al. [2010] for the period September 2003–August 2011. The spatial trends from GRACE, laser altimetry and SMB outputs are compared over multiple time periods to take advantage of the longterm, secular nature of the GIA signal. Correcting GRACE for GIA with Wu et al. [2010] leads to a large positive mass anomaly in Northeast Greenland. This mass anomaly was not evident in GRACE corrected with the other GIA models or in the independent datasets. To examine the temporal variability, we divide the Greenland ice sheet into fourth major sectors and use a least squares mascon approach to calculate each time series. In the Northeast, the recovered ice mass change from Wu et al. [2010] is 46 to 49 Gt/yr larger than with the other two corrections, and 54 Gt/yr larger than the surface mass balance trend. Overall, we find that GRACE corrected with Wu et al. [2010] results in much lower average ice sheet losses than with the other two GIA corrections. Within this research, we conclude that the GRACE ice mass changes calculated using the GIA correction from Wu et al. [2010] is not compatible with observations of ice elevation changes and regional climate model outputs of SMB. GRACE ice mass changes calculated using the Simpson et al. [2009] and A et al. [2013] GIA corrections are found to be consistent with the altimetry observations and SMB data. For this study, I processed the GRACE data, developed the GRACE regional mass balance method using spherical caps, converted the altimetry and SMB datasets into GRACE-like estimates, and calculated the GRACE and SMB time series.

Chapter 3 is the second study presented as part of this dissertation. This study investigates the regional acceleration in ice mass loss in Greenland and Antarctica using GRACE time-variable gravity over 2003–2013. The statistical significance of linear and quadratic model spatial patterns are determined using both regression fit errors and a form of the Akaike information criterion (AIC). The temporal variability of regions with significant trend and acceleration signals are evaluated using the least squares mascon method we derive in the
previous study. For both ice sheets, we find that a few key regions are the major drivers of current ice sheet mass change. In Greenland, the mass loss over the study period is predominantly controlled by the SE (40%) and NW (30%) sectors, while the acceleration in loss is dominated by the SW (54%) and NW (34%) sectors. The largest acceleration in Greenland mass loss is determined to be caused by a decrease in cumulative SMB anomalies within the SW sector. Surface mass balance changes account for 68% of Greenland mass loss of the entire ice sheet and 79% of the mass loss acceleration during 2003–2012. In Antarctica, most of the mass loss is from the Amundsen Sea Embayment and the Antarctic Peninsula. The mass loss in these two sectors is predominantly due to changes in the regional velocity structures. The Amundsen Sea Embayment is also the dominant sector of accelerating ice mass loss. Queen Maud Land in East Antarctica is found to experience an overall mass gain for the study period, which largely occurs in two pulses after 2008. For both ice sheets, longer time series of GRACE data are required in order to determine the significance of the accelerating mass loss signals with respect to natural fluctuations in surface mass balance. Within this research, I processed the GRACE data, developed the new configurations of spherical caps for Greenland and Antarctica, converted the SMB datasets into GRACE-like estimates, and calculated the GRACE and SMB time series.

The third study within this dissertation (Chapter 4) focuses on the Amundsen Sea Embayment, the region we determine to be the primary contributor to Antarctic mass loss in the previous study. Here, we quantify the ice sheet mass balance using four independent techniques, including the least squares mascon approach we develop in the prior studies. We find excellent agreement in terms of mass loss and acceleration in mass loss from these independent techniques during common periods. We show that the airborne lidar measurements from Operation IceBridge are sufficient to extend the time series of ice sheet mass balance from ICESat laser altimetry. We find that the significant fluctuations in surface mass balance average out after a couple of decades of observation within the region. Our mass balance record shows a large increase in ice sheet mass loss rates in recent years in comparison to
the early 1990’s and the entire observation period. We find that most of the cumulative ice mass loss occurs over the last ten years of observation. Within this study, I developed the laser altimetry method following Schenk and Csatho [2012], processed the GRACE and laser altimetry datasets, and calculated the time series from each method. Data from this work has been incorporated into additional studies of the regional mass and surface elevation change.

The final study within this dissertation (Chapter 5) uses a combination of satellite and airborne lidar measurements to evaluate surface elevation changes in Greenland. The method we use within this work derives rates of elevation change at very high spatial resolution by associating elevation measurements by triangulated irregular networks (TIN’s). The combination of multiple laser altimetry datasets increases the overall spatial coverage of elevation measurements for several years. Results from the technique are used to investigate regional thinning at Jakobshavn Isbræ, the fastest and most active glacier in Greenland, and to assess the accuracy of surface mass balance outputs from regional climate models. We find that rates of thinning over Jakobshavn Isbræ during the major melt season of 2012 were the strongest compared to any other period. Surface mass balance outputs from the regional climate models are found to correspond well with the altimetry estimates over long periods. The technique can be used for other regions of surface elevation change, and can incorporate additional altimetry datasets. The 2015 IceBridge Arctic fall campaign provides the opportunity to study seasonal ice sheet elevation change to compare with regional climate model outputs. This work is intended for publication following further investigation of the results.
6.2 Implications and Future Directions

The observations we provide as part of this dissertation work may help inform policy makers on the current state of the ice sheets and the implications of future ice sheet loss. Assuming that the current GRACE trend and acceleration values (358±72 Gt/yr and 31±3 Gt/yr\(^2\) over 2002–2015 respectively) continue into the future, losses from the Greenland and Antarctic ice sheets will raise global sea levels by 450 ± 50 mm from the start of the GRACE mission to the year 2100 [Update from Velicogna et al., 2014, Chapter 3]. However, the duration of the GRACE mission is short compared to the periods of natural variability in surface mass balance, which limits the partition between the current ice mass loss acceleration and natural variability. In addition, how the magnitude of the ice sheet mass loss acceleration will change into the future is highly uncertain. More observations are needed to help understand the role of the processes contributing to current ice sheet change, such as increases in meltwater runoff and interactions with warming ocean waters. There are current and future land, sea, air and space missions that may help provide some of these observations. The methods outlined in this dissertation will work well with several upcoming NASA satellite missions towards investigating regional ice sheet change across spatial and temporal scales.

The second ICESat laser altimetry mission is due for launch in 2018. The primary instrument of ICESat-2 is the Advanced Topographic Laser Altimeter System (ATLAS), a multi-beam, photon-counting laser altimeter. The 6-beam setup will provide measurements with greater spatial resolution and accuracy than the original mission. The cross-track acquisitions will help determine the surface slopes at each nadir point. ICESat-2 will acquire data continuously throughout the year, which will enable the quantification of seasonal surface elevation changes ice-sheet-wide, and investigations of rapid thinning at key outlet glaciers and surface ablation zones. The high-precision instrument will enable better characterization of the processes governing current ice sheet change. Two testbed airborne altimeters,
MABEL (Multiple Altimeter Beam Experimental Lidar) and SIMPL (Slope Imaging Multi-Polarization Photon Counting LiDAR), have flown to examine the capabilities of the future ICESat-2 lidar instrument. ICESat-2 laser data can be used in conjunction with new Operation IceBridge data to extend the surface elevation time series of both Chapters 4 and 5 once the data becomes available.

The GRACE follow-on (GRACE-FO) mission is a collaboration between NASA and the German Research Centre for Geosciences (GeoForschungsZentrum, GFZ) slated for launch in 2017. The mission is to continue the legacy of the original GRACE mission of providing observations of the Earth’s gravitational field. GRACE-FO will have much of the same instrumentation as the original GRACE and lunar GRAIL missions with the exception of an additional ranging instrument. The high-precision laser interferometer will be able to detect smaller gravity differences than the onboard microwave ranging instrument. New GRACE-FO data, once available, will help extend the regional GRACE time series from Chapters 2 to 4 into the future. Our GRACE regional ice sheet method can possibly be improved by tailoring the mascon kernels for different regions of interest. One potential method is to use optimized centroidal Voronoi tessellations (CVT) to define the center points of the spherical caps with refinement on an area of interest. In addition to the upcoming ICESat-2 and GRACE-FO data, our Amundsen Sea Embayment time series can be supplemented further by incorporating new ice discharge data, new surface mass balance outputs from multiple RCM’s, and CryoSat-II radar altimetry data. Longer time series are critical for the separation of climate signals from accelerating losses, and for the detection of new and rapid ice sheet change.

The Glacial Isostatic Adjustment (GIA) correction in Antarctica is the largest source of uncertainty in ice sheet mass balance estimates from GRACE time-variable gravity. New sets of GIA constraints would help modelers of Glacial Isostatic Adjustment, and help improve measurements and projections of ice sheet mass loss. A methodology similar to the GIA
evaluation study used in Greenland (Chapter 2) can be applied in Antarctica to investigate GIA on a regional scale. However, calculating and constraining GIA in East Antarctica is a very difficult problem due to the region’s vast size, lack of grounding points for GPS, and low signal-to-noise ratios for several measurements. The roughly 15 cm single-shot accuracy of ICESat-1 makes the discrimination of ice sheet surface elevation changes from laser altimetry in the slow-changing East Antarctic interior a difficult problem. Dataset accuracy and lack of publicly available firn models currently limits the capability of gravimetry and laser altimetry combinations to study GIA (as described in Wahr et al. [2000]) in East Antarctica. ICESat-2 and GRACE-FO can contribute towards solving this problem by providing better, longer, and more complete observations.

Lack of observations on the individual surface mass balance (SMB) components, such as internal refreezing, makes it hard to verify current models. The GRACE method outlined in Chapter 3 can be used as an integrated evaluation of surface mass balance estimates for regions where changes in ice sheet dynamics is minimal or well known. Our independent GRACE measurements can help evaluate outputs from the current regional climate models (RCM’s) Regional Atmospheric Climate Model (RACMO) and Modèle Atmosphérique Régional (MAR), and coupled global climate models (GCM’s). However, the limited spatial resolution of GRACE and uncertainties in the reference period for calculating cumulative surface mass balance anomalies [Colgan et al., 2015] restricts the effectiveness of this method. One potential technique is to compare the short term variations, such as the timing and magnitude of each summer melt season, between GRACE and the surface mass balance outputs. The triangulated irregular network (TIN’s) method developed in Chapter 5 can provide another independent measurement for evaluating RCM outputs in the ablation zones of Greenland at locations where dynamic thinning is minimal and the surface elevation changes occur over glacial ice. Uncertainties in firn compaction and the conversion from volume to mass limits the effectiveness of this method. However, this method is useful in certain cases, and can provide a multi-decadal metric from 1993 onwards. Improvements
to current RCM and GCM outputs will in turn help constrain ice sheet accumulation and runoff estimates under future climate change scenarios.

Finally, ice sheet retreat and collapse are major deficiencies in predicting and reconstructing sea levels. The coherent set of observations of the rapidly changing Amundsen Sea Embayment in Chapter 4 can be used for glaciological interpretation, and for inclusion within data assimilating ice sheet numerical models. Our improved observations of surface elevation change from Chapter 5 can be used for the surveyed zones of Greenland and Antarctica to investigate the dynamic thinning of key outlet glaciers. Providing sets of independent, baseline observations to ice sheet modeling community may help improve the ability of models to the reproduce current changes. Improvements to ice sheet models and coupled ocean-ice dynamical models will help improve our predictive capability of future sea level rise and climate change due to changes in the Earth’s ice sheets.
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