Long-term and inter-annual mass changes in the Iceland ice cap determined from GRACE gravity

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Abstract. The Gravity Recovery and Climate Experiment (GRACE) satellites have measured anomalies in the Earth's time-variable gravity field since 2002, allowing for the measurement of the melting of glaciers due to climate change. Many techniques used with GRACE data have difficulty constraining mass change in small regions such as Iceland, often requiring broad averaging functions in order to capture trends. These techniques also capture data from nearby regions, causing signal leakage. Alternatively, Slepian functions may solve this problem by optimally concentrating data both in the spatial domain (e.g., Iceland) and spectral domain (i.e., the bandwidth of the data). In this project, we use synthetic experiments to show that Slepian functions can capture trends over Iceland without meaningful leakage and influence from ice changes in Greenland. We estimate a mass change over Iceland from GRACE data of approximately -9.68 $\pm$ 0.99 Gt/yr between January 2002 and November 2016, with an acceleration of 1.07 $\pm$ 0.50 Gt/yr^2.

1 Introduction

Earth's mountain glaciers and ice caps have been losing mass in response to climate change (Stocker et al., 2013), and in 2014 this mass loss accounted for approximately 25% of current observed global mean sea level rise (Chen et al., 2017). About 11% of Iceland is covered by approximately 3,500 Gt of ice (Björnsson and Pálsson, 2008; Pfeffer et al., 2014). Between 1995 and 2013, Iceland experienced -9.5 ± 1.5 Gt/yr of average ice mass change, with negative acceleration (meaning an increasing rate of mass loss) (Björnsson et al., 2013). In Iceland, ice mass changes and volcanic activity are linked, as unloading caused by ice melt causes decompression melting. This melting allows the mantle to upwell (Carolina and Freysteinn, 2008) and increases volcanic activity (Compton et al., 2017). Detailed and accurate measurements of glacier mass balance are therefore important to understand not only sea level rise, but also volcanic activity in Iceland and potential international consequences. For example, the 2010 Eyjafjallajòkull eruption stranded over 8.5 million passengers of 108,000 canceled flights, costing airlines 1.7 billion USD (Alexander, 2013).

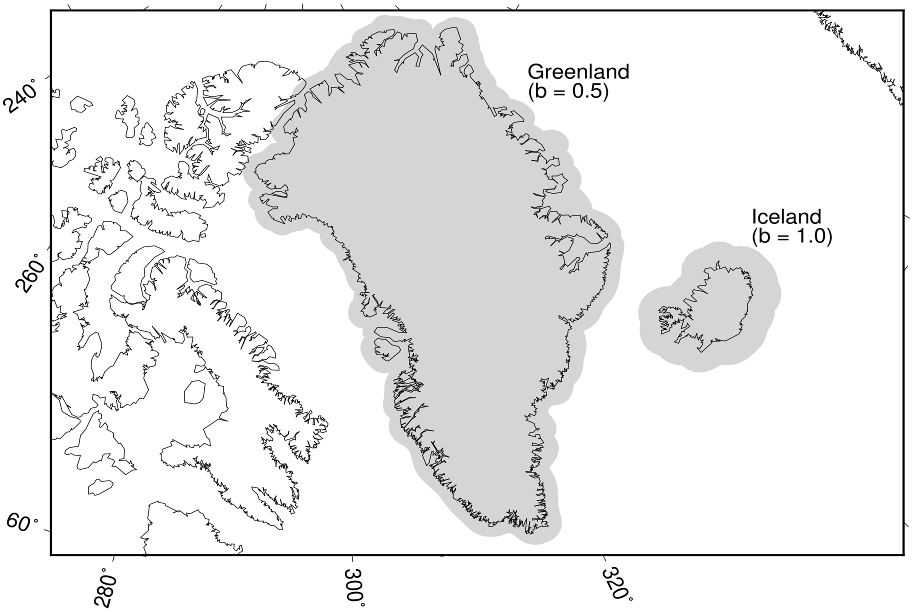

Figure 1. A map of the localization regions used for Greenland and Iceland. The gray areas represent the region of Greenland buffered by 0.5° and the region of Iceland buffered by 1.0°, respectively. We use an oblique mercator projection centered in Greenland.

The Gravity Recovery and Climate Experiment (GRACE) consisted of twin satellites which orbited the Earth about 15 times per day from 2002 until 2017. Monthly global gravity field data products were derived from the recorded GPS positions of the satellites and changing distance between them (Tapley et al., 2004). GRACE data are sensitive to mass changes near Earth's surface, and have been used to investigate time variable changes in the cryosphere (Harig and Simons, 2012), hydrosphere (Longuevergne et al., 2010), atmosphere (Syed et al., 2005), and the solid Earth (Davis et al., 2004).

In this paper we derive ice mass loss trends over Iceland from GRACE data. In order to examine signals in Iceland separately from other mass changes we use a method of spatio-spectral localization on the sphere in which we transform the data onto a basis of spherical Slepian functions (Simons et al., 2006; Slepian, 1983). This technique has previously been applied to various spatial domains including Greenland (Harig and Simons, 2012, 2016), Antarctica (Harig and Simons, 2015), and the High Plains Aquifer in the USA (Longuevergne et al., 2010). We perform a series of synthetic tests to examine the method's ability to resolve smaller magnitude signals (i.e. Iceland) in the presence of nearby larger magnitude signals (i.e. Greenland). Finally, we discuss recent ice mass changes in the context of recent Icelandic volcanism.

2 Methods

We use GRACE Release 5 Level 2 monthly degree-60 gravity field data from the Center for Space Research (CSR), University of Texas Austin, spanning April 2002 to November 2016 (inclusive). These data are distributed in the form of Stokes coefficients, which describe the data in a spherical harmonic basis. To analyze the data, we use the method of Harig and Simons

(2012) and transform GRACE data from the spherical harmonic basis into a Slepian basis. The code to perform Slepian analysis on the sphere is freely available online (Harig et al., 2015). Per the recommendations of the CSR and Jet Propulsion Laboratory (JPL) data centers (Center for Space Research, 2013), we replace the degree 2 order 0 coefficients with measurements from satellite laser ranging (Cheng et al., 2013). GRACE does not directly provide degree 1 coefficients, which are related to the relative positions of the center of mass and center of figure of the Earth's outer surface. We therefore use degree 1 coefficients from Swenson et al. (2008). Finally, we convert the geopotential data into surface mass density using the method of Wahr et al. (1998).

The spherical harmonic basis is orthonormal over the unit sphere but is not orthogonal over arbitrary regions of the sphere, which makes it ill-suited for the representation of GRACE data within spatially localized regions such as Iceland. We therefore project our data onto a basis of spherical Slepian functions, which are orthogonal over arbitrary (and non-convex) regions of the sphere (Bates et al., 2017). Slepian functions, also known as prolate spheroidal wave functions, optimally concentrate energy (and consequentially sensitivity to the data) both in the spatial domain (e.g., Iceland) and spectral domain (in this case the bandwidth of the data) (Slepian, 1983).

The Shannon Number, $N = (L+1)^2 A/(4\pi)$, is a measurement of spatiospectral optimization, and gives the number of optimal Slepian functions concentrated within a spectral domain and a spatial region of area $A/(4\pi)$ on the unit sphere (Kennedy and Sadeghi, 2013). We project the GRACE data from the spherical harmonic basis onto the N most optimal functions of the Slepian basis concentrated within the region of interest and bandwidth of the data. We use synthetic experiments to determine whether these functions fully capture the trend over the region. If the Slepian functions are insufficiently spatially concentrated, then we are able to enlarge the region (therefore increasing N) by buffering the coastlines of the region by a specific distance, e.g. $1.0°$. Our synthetic experiments inform the choice of a spatial buffer around the region within which the functions are better contained, as illustrated for Iceland and Greenland in Fig. 1. These are described in detail in Section 3.

GIA model	Mass Trend Adjustment (Gt/yr)
Paulson et al. (2007), adjusted by A et al. (2013)	-0.39
Wang et al. (2008)	0.58

Table 1. GIA adjustments for Iceland when the models are processed in the same manner as GRACE data.

We correct the GRACE data for viscous deformation of the solid Earth due to glacial isostatic adjustment (GIA). These models are processed in the Slepian basis similarly to GRACE data. We consider the model of Paulson et al. (2007) which uses a radially symmetric Earth structure, and the model of Wang et al. (2008) which considers laterally varying viscosity in the upper mantle. The viscosity of the upper mantle beneath Iceland is estimated to be lower than average, implying that modern mass change rates estimated by using a globally averaged Earth structure may not be accurate (Sigmundsson, 2006).

In order to choose between the adjusted Paulson et al. (2007) model and the Wang et al. (2008) model for GIA correction, we compare the the two corrections when applied over Iceland for our time period (Table 1). The first model suggests a mass gain over Iceland due to GIA, while the second suggests a mass loss. Schmidt et al. (2013) uses a 3-D model of GIA in Iceland since

1980, and improves on prior studies by considering all of Iceland's ice caps and mantle, rather than only modeling Iceland's largest ice cap and the section of mantle beneath it. Sørensen et al. (2017) improves on prior models of GIA in Iceland by incorporating both GRACE and relative sea level (RSL) data, and their model predicts a pattern consistent with that predicted for PGR using the ICE-5G history. Both studies conclude that GIA over Iceland causes a net mass gain, meaning that the GIA correction over Iceland should increase the magnitude of the adjusted mass loss trend. This supports the use of the adjusted Paulson et al. (2007) model over the Wang et al. (2008) model. We therefore use the adjusted Paulson et al. (2007) model.

Following correction for GIA, we construct third-degree polynomial functions with annual and semi-annual periodicity to fit the time series of each Slepian coefficient. We do this for the N most optimally concentrated Slepian functions within the bandwidth of our data and buffered region of interest. The sum of integrals of these polynomials over time produce the total time-variable mass loss trend for the region. We use a 95% confidence interval for our uncertainty. We consider the uncertainty of the GIA correction ill-posed, because too few GIA models exist for meaningful comparison. We therefore do not incorporate the uncertainty of the GIA correction into our overall uncertainty.

For a given region R of the unit sphere Ω we choose an optimal buffer through a synthetic experiment, as in Harig and Simons (2012). We use a "boxcap" function to construct a synthetic signal with 200 Gt/yr of mass loss uniformly distributed over R and 0 Gt/yr over $\Omega \setminus R$. Next, we add random gaussian noise to the synthetic data, with covariance matching that of the actual GRACE data. This introduces synthetic uncertainty which reflects the actual uncertainty in the data. From these synthetic data, we then use our Slepian based analysis to recover the trend over R at each combination of buffer from 0.0° to 2.0° and bandwidth from 0 to 60, and plot the resulting percent-recovered contours. We select the (buffer, bandwidth) tuple of bandwidth closest to 60 which best aligns with the 100% recovery contour of the original signal.

A large buffer for Iceland introduces the problem of signal leakage since Iceland neighbors another area with large mass change signals, Greenland. For non-overlapping regions the total mass change will be conserved, but there is the possibility that signals in Greenland might be observed in Iceland. We address this with a synthetic experiment which measures leakage from Greenland to Iceland and from Iceland to Greenland.

3 Results

Our first synthetic experiment informs our choice of a 1.0° buffer around Iceland. In Fig. 2 we show the results of our recovery experiment for Iceland with a synthetic boxcap signal. For several combinations of spectral bandwidth and buffer extent the synthetic data trend can be completely recovered (panel a). In panel b we see that the green contour indicating 100% synthetic trend recovery corresponds to several different magnitudes of recovered trends from GRACE data. As buffer extent starts increasing above roughly 2.0°, the Iceland region begins to intersect with the Greenland region resulting in the larger negative trend values at large buffers. Overall, we choose the experimental case where $L = 60$, corresponding to a buffer size of 1.0°, because it maximizes the available bandwidth of the GRACE data and reduces the opportunity for signal leakage with the nearby region of Greenland (see below). When we repeat this synthetic experiment for Greenland, we confirm the 0.5° buffer choice of Harig and Simons (2012).

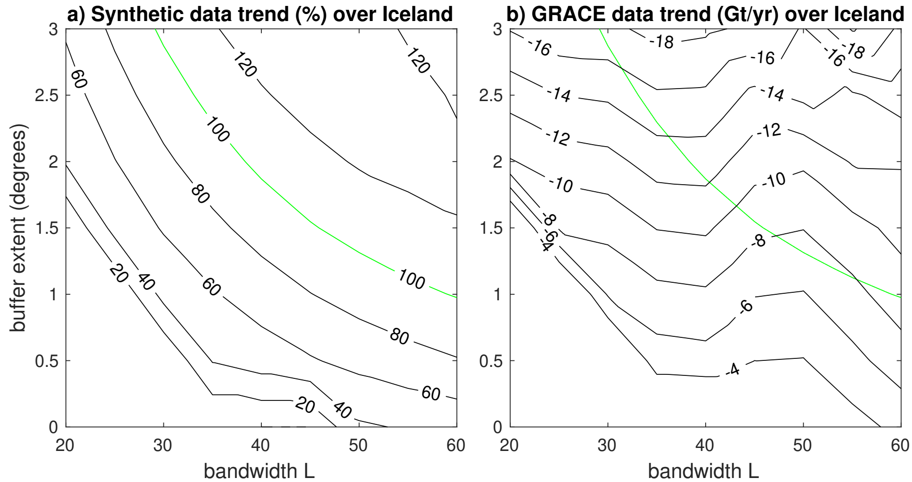

Figure 2. a) Synthetic trend recovery over Iceland. The contours represent the percent signal recovered from the synthetic data for each choice of buffer and bandwidth. b) Real trend recovery over Iceland. The contours represent the actual trend estimates calculated from GRACE data for each choice of buffer and bandwidth. The 100% recovery contour from a) is shown in green. For the bandwidth of the data ($L = 60$), the 100% recovery contour closely aligns with the choice of a 1.0° buffer around Iceland.

Using a buffer of 0.5° around Greenland and 1.0° around Iceland, we measure -238.20 ± 6.29 Gt/yr of mass change over Greenland and -9.68 ± 0.99 Gt/yr of mass change over Iceland between January 2002 and November 2016. These results are discussed further below. Here we use these preliminary trends in a second synthetic experiment to estimate leakage between Iceland and Greenland. We use a "boxcap" function to create synthetic signals over the unbuffered regions as detailed in Table 2 below. Then for each region we compare the trends recovered with and without the other region in the synthetic data. Synthetic noise matching the covariance of the actual GRACE data is added to each region before recovery, in order to apply the uncertainty in the GRACE data to the synthetic data.

Signal Applied (Gt/yr)		Signal Recovered (Gt/yr)	
Greenland	Iceland	Greenland	Iceland
238.20	9.68	235.54	9.13
238.20	0.00	235.79	
0.00	9.68		9.49

Table 2. Synthetic trends recovered over each region with and without leakage from the other region. We recover the trend over Greenland buffered by 0.5° and over Iceland buffered by 1.0°.

The absolute value of the difference in trends recovered over each region A with and without the other region B in the synthetic signal serves as an approximation of the leakage from B into A. Leakage from Greenland appears to cause ± 0.36 Gt/yr of uncertainty in the trend over Iceland, and leakage from Iceland appears to cause ± 0.25 Gt/yr of uncertainty in the trend over Greenland. These uncertainties are respectively smaller than our $\pm 2\sigma$ uncertainties of $\pm$ 0.99 Gt/yr over Iceland and $\pm$ 6.29 Gt/yr over Greenland, meaning that uncertainty caused by leakage between Greenland and Iceland does not meaningfully influence the outcome of our measurement over either region. We therefore do not adjust our results for leakage.

In Fig. 3 we show using a linear estimate that Iceland has lost -9.68 $\pm$ 0.99 Gt/yr from 2002 through 2016. During and after 2010 the data diverge from the prior linear trend. In 2010 there is an abrupt decrease in mass, and from 2010 to 2016 the secular trend in mass is relatively flat. The abrupt change in 2010 occurs near in time to the Eyjafjallajökull volcanic eruption, which is highlighted on the graph in red.

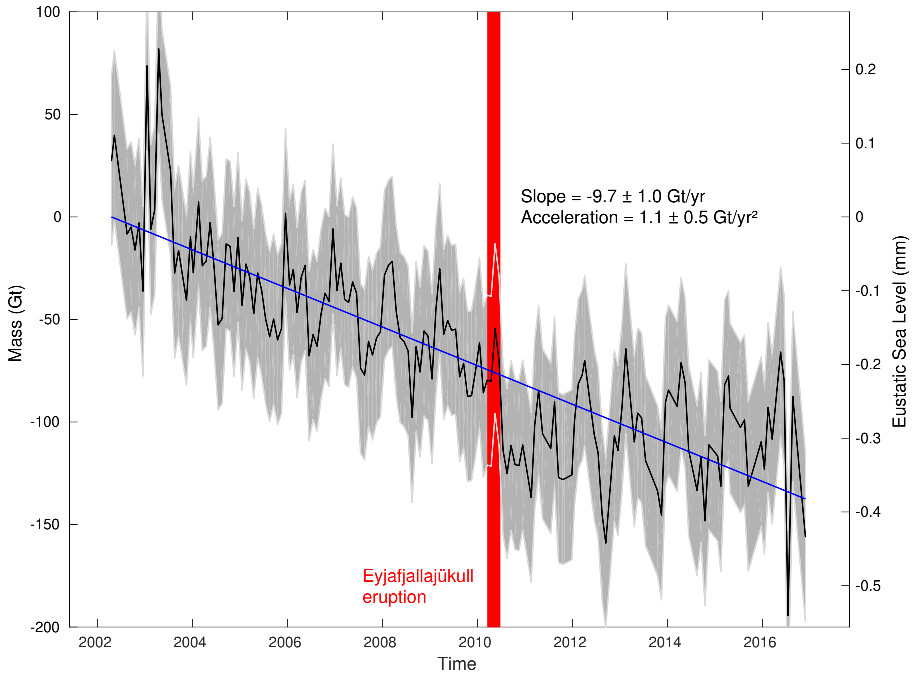

Figure 3. GRACE-derived mass loss trend over Iceland. The black line indicates the total integrated monthly mass change since 2002 (gigatons left axis and equivalent eustatic sea level change, right axis), and the gray shaded region is its $\pm 2\sigma$ uncertainty interval. The blue line is the best-fit linear equation over the entire timespan. The red strip highlights the time period of the Eyjafjallajökull volcanic eruption.

We next represent the total integrated mass change over Greenland (Fig. 4) and Iceland (Fig. 5) in map form. The map for Greenland has the same qualitative spatial distribution as the equivalent map by Harig and Simons (2012), but differs in magnitude because the date ranges are different. As expected, mass loss is mostly concentrated in the southeast and northwest coastal areas of Greenland. In Iceland, mass loss is centered over the eastern half of the island, which contains the larger volume of ice stored in glaciers and ice caps (Björnsson and Pálsson, 2008).

The map for Iceland is simpler because there are only two Slepian functions optimally concentrated within the buffered region and bandwidth, in contrast to 20 such functions for Greenland. The reason for this is the bandwidth of the data; the spatial resolution of the GRACE data at $L = 60$ is ~330 km, and Iceland is roughly 500 km wide, so most of Iceland is described by a single independent point in the GRACE data. With a Shannon number of $N = 2$ the combined basis closely resembles an axisymmetric function. In contrast the $N = 20$ functions of Greenland mean that, in Greenland, the five largest functions by integral mass contribute 85% of the total mass change while the remaining functions primarily contribute changes to the spatial pattern (Harig and Simons, 2012).

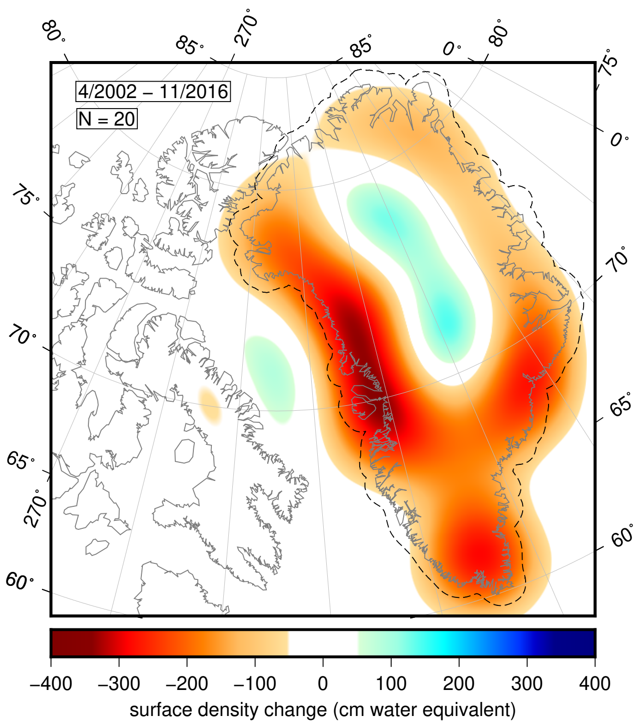

Figure 4. Geographic pattern of mass change over Greenland, shown in centimeters water equivalent. The dashed line around Greenland is its $0.5°$ buffer. The figure represents the total mass change over the date range specified in the top left. Below the date range is the Shannon number $N = 20$ for the buffered region and bandwidth of 60, indicating that the figure represents the sum of 20 optimal Slepian functions.

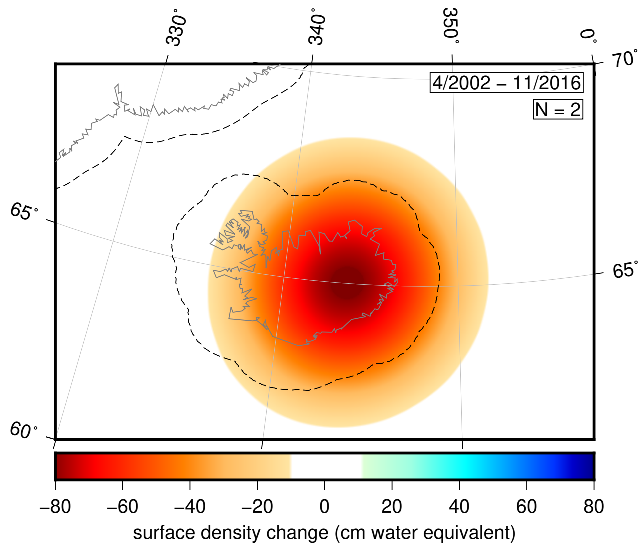

Figure 5. Geographic pattern of mass change over Iceland, shown in centimeters water equivalent. The dashed border in the northwest is the 0.5° buffer around Greenland, and the similar dashed border around Iceland is its 1.0° buffer. The figure represents the total mass change over the date range specified in the top right. Below the date range is the Shannon number $N = 2$ for the buffered region and bandwidth of 60, indicating that the figure represents the sum of 2 optimal Slepian functions.

4 Discussion

The April 2010 Eyjafjallajòkull volcano eruption rapidly melted 0.2 Gt of ice through magma heat transfer and deposited an almost equal quantity of tephra (half of the total tephra emitted) within a 5 km radius of the summit caldera (Gudmundsson et al., 2012; Tarasewicz et al., 2012). Tephra distributed by the eruption across Iceland and Europe decreased exponentially in deposition thickness relative to distance from the summit caldera (Bonadonna et al., 2011), so that most of the distribution of tephra outside of this 5 km radius within Iceland ranged between 2 and 0.01 cm (Arnalds et al., 2013).

Tephra both increases melting by decreasing albedo and decreases melting by insulating ice. If a tephra deposition is sufficiently thick, its insulating effect can overpower melting caused by decreased albedo, leading to a net decrease in melting (Reznichenko et al., 2010; Reid and Brock, 2010; Östrem, 1959). The critical thickness of tephra emitted by Iceland's major volcanoes at which the tephra changes from generally increasing melting to generally decreasing melting is estimated at between 0.85 and 1 cm (Möller et al., 2016, 2018). Tephra coverage below this critical thickness across the south and southeast of Iceland produced a short-term decrease in Iceland's average albedo and resultant 15 Gt of additional ice melt over 2010 (Björnsson et al., 2013). Tephra grain sizes generally decreased with distance from the caldera (Gudmundsson et al., 2012), and smaller grain sizes are more transient (Arnalds et al., 2013). Iceland experienced extreme wind erosion between June and October 2010 (Arnalds et al., 2013), as well as increased snow and ice melt over the following winter, all of which likely redistributed tephra from thin layers to more concentrated areas, partially explaining the short lifetime of this increase in melting (Björnsson et al., 2013).

Following the eruption, tephra deposition thickness within a 5 km radius of the Eyjafjallajòkull caldera ranged from 5 cm to 200 m. This exceeds the critical thickness and will probably meaningfully delay the total deglaciation of Eyjafjallajòkull (Gudmundsson et al., 2011). While the short-term result of tephra distribution was a decrease in albedo and increase in melting, the thick and course tephra over the Eyjafjallajòkull ice cap appears to dominate the long-term trend, causing a net decrease in melting across Iceland (Compton et al., 2017).

Date Range	Slope (Gt/yr)	Acceleration (Gt/yr^2)
17 April 2002 - 26 November 2016	-9.7 $\pm$ 1.0	1.1 $\pm$ 0.5
17 April 2002 - 20 March 2010	-10.6 $\pm$ 2.0	1.9 $\pm$ 2.0
23 June 2010 - 26 November 2016	-0.3 $\pm$ 3.5	-1.7 $\pm$ 4.2

Table 3. Ice mass trends and accelerations over Iceland before and after the 2010 Eyjafjallajòkull eruption. The Eyjafjallajòkull volcano erupted from 20 March to 23 June 2010. Our data spans 17 April 2002 to 26 November 2016.

We observe a linear mass loss trend over Iceland of approximately -10.6 Gt/yr prior to the 2010 eruption (Table 3), which aligns well within the uncertainty of $\pm$ 2.0 Gt/yr with prior estimates of -11.4 Gt/yr (B. et al., 2008; Sørensen et al., 2017) and -11.0 Gt/yr (Björnsson et al., 2013) over similar time periods. The observed positive acceleration prior to the 2010 eruption compares favorably with the 3.1 $\pm$ 5.1 Gt/yr^2 estimated over March 2003 - February 2010 by Schrama and Wouters (2011), but contrasts with the vertical acceleration of the Icelandic crust (Compton et al., 2015) and horizontal acceleration in the ice-fronts of Icelandic glaciers (Chandler et al., 2016). This disagreement raises the question of how much data are required to accurately measure either slope or acceleration over Iceland with a 1.0° buffer.

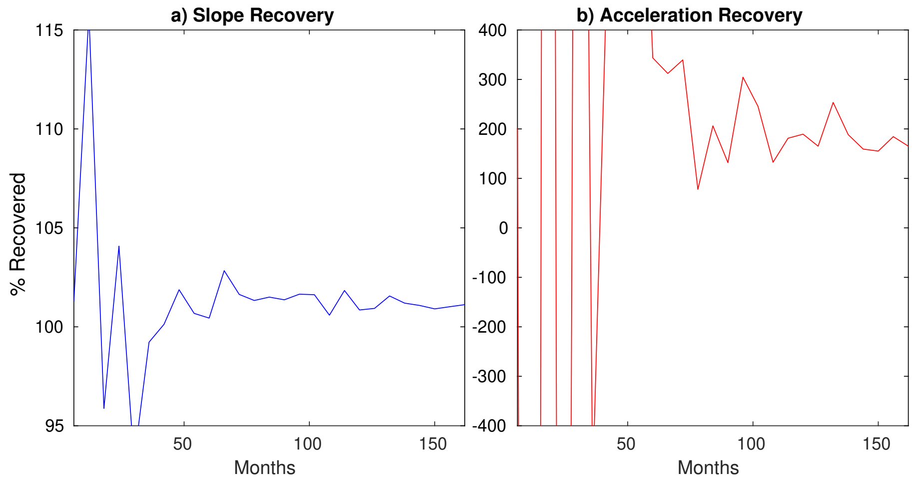

Figure 6. Results of our synthetic experiment for Iceland to test recovery of a) slope and b) acceleration parameters. For varying amounts of months of consistent synthetic data over Iceland we use our Slepian method to measure the slope and acceleration of the mass loss trend. The x-axis denotes the number of cumulative months of synthetic data processed, and the y-axis denotes the $100 * \frac{R}{A}$, where R is the recovered slope or acceleration and A is the applied synthetic slope or acceleration.

We perform a final synthetic experiment to test our ability to recover the slope and acceleration parameters of our fit for Icelandic mass change. In this synthetic experiment we apply -200 Gt/yr uniformly distributed over Iceland buffered by 1.0° and then recover the first and second-order trends over cumulative 6-month intervals. In Fig. 6a we see that after about four years the slope recovered is close to 100% of the slope applied, indicating that a minimum of four years of data are needed to accurately estimate the slope of a linear trend. To test acceleration recovery we apply a trend with an initial slope of -200 Gt/yr and acceleration of 1.0 Gt/yr^2 uniformly distributed over Iceland buffered by 1.0° and then recover the acceleration over cumulative 6-month intervals (Fig. 6b). We see that the recovered acceleration only begins to converge upon 100% of the applied acceleration after 12 years.

In the context of this last experiment our slope estimates over the time periods preceding and following the 2010 eruption are likely accurate, but our acceleration estimates over each are too uncertain to be meaningful. Our average acceleration estimated over the 14 year timespan likely exhibits an uncertainty of < 50%, in rough agreement with the uncertainty yielded by the 95% confidence interval. We and Schrama and Wouters (2011) both estimate acceleration values with > 100% uncertainties prior to the 2010 eruption, and our synthetic experiments suggest that any error in the acceleration is probably positive, so the negative vector acceleration suggested by Compton et al. (2015) and Chandler et al. (2016) is most likely correct for the time period preceding the eruption. Insufficient data exist to describe the acceleration after the eruption. (Wouters et al., 2013) find that at least 10 years of data for Antarctica or 20 for Greenland are required to resolve the long-term acceleration trend over each within an accuracy of $\pm$ 10 Gt/yr^2. Since Iceland is much smaller than Greenland or Antarctica and exhibits orders of magnitude less mass change, its long-term acceleration may require even more time to distinguish within the context of short-term ice sheet variations. As with the larger Greenland and Antarctic ice sheets, stochastic variability is likely the main reason that short term and long term trends do not necessarily agree.

5 Conclusions

We find that mass loss trends over Iceland can be accurately measured in GRACE data using Slepian functions without explicitly accounting for leakage from nearby Greenland. We measure an average mass loss trend over Iceland of -9.68 $\pm$ 0.99 Gt/yr between April 2002 and November 2016, which agrees well with the previous work of Björnsson et al. (2013) who found an average trend of -9.5 $\pm$ 1.5 Gt/yr since 1995. Prior to the eruption, we observe a linear trend of -10.6 $\pm$ 2.0 Gt/yr, which aligns closely with existing estimates over that time period.

The feedback loops of ice unloading, seasonal weather patterns, and volcanic activity in Iceland are poorly understood, complicating prediction of long-term mass loss after the eruption (Compton et al., 2017). We consider less than six years of data after the 2010 eruption, and less than 10 years of data before it. Wouters et al. (2013) as well as our synthetic slope and acceleration recovery experiments suggest that both date ranges are insufficient to accurately measure acceleration in the GRACE data.

If Iceland returns to a mass loss trend between -9.7 and -10.6 Gt/yr, it could completely deglaciate within 3 centuries (Pfeffer et al., 2014). However, feedback loops of climate change, post-glacial rebound, ice mass unloading, and volcanic activity will likely increase the variability of annual mass loss trends in Iceland (Compton et al., 2017; Björnsson et al., 2013). More GRACE data are required to determine long-term trends (Wouters et al., 2013). The GRACE Follow-On (GRACE-FO) mission launched on May 22, 2018. By incorporating GRACE-FO data with vertical GPS measurements, satellite imaging of tephra deposition, and *in situ* measurements, future research may better illuminate the post-eruption mass-loss trend over Iceland.

Code availability. The code used in this work is available freely online (Harig et al., 2015). Installation instructions for the various Slepian code repositories can be found at http://github.com/Slepian/Slepian.

Data availability. GRACE data are freely available from the NASA Physical Oceanography Distributed Active Archive Center: ftp://podaac.jpl.nasa.gov/a

Competing interests. No competing interests are present.

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