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Mass change of Arctic ice caps and glaciers: implications of regionalizing elevation changes

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Abstract

Recent studies have determined mass changes of Arctic ice caps and glaciers from satellite altimetry. Determining regional mass balance of ice caps and glaciers using this technique is inherently difficult due to their size and geometry. Furthermore these studies have mostly relied on one method or the same types of methods to determine the regional mass balance, by extrapolating elevation changes using their relation to elevation. This makes the estimation of mass balance heavily dependent on the method used to regionalize the elevation changes. Left without consideration large discrepancies can arise in the mass change estimates and the interpretation of them. In this study we use Ice, Cloud, and land Elevation Satellite (ICESat) derived elevation changes from 2003–2009 and determine the impact of different regionalizing schemes on the mass change estimates of the Arctic ice caps and glaciers. Four different methods, based on interpolation and extrapolation of the elevation changes were used to quantify this effect on the regional mass changes. Secondly, a statistical criteria was developed to determine the optimum method for each region in order to derive robust mass changes and reduce the need of external validation data. In this study we found that the range or spread of the estimated mass changes, for the different regions, was highly correlated to the inter-annual variability of the elevation changes, driven by the different climatic conditions of the regions. Regions affected by a maritime climate show a large range in estimated values, on average 1.5–2 times larger than the predicted errors. For regions in a continental regime the opposite was observed, and the range of the values lies well inside the error estimates. We also found that the extrapolation methods tend on average to produce more negative values than the interpolation methods and that our four methods do not fully reproduce the original histogram. Instead, they produce more negative distributions than the original which may indicate that previous and these current estimates using ICESat observations might be overestimate by as much as 4–19 %, depending on region. This should therefore be taken into account when deriving regional mass balance from satellite altimetry in regions which show

high inter-annual variability of elevation changes. In these regions several different independent methods should be used to capture the elevation change pattern and then analyzed to determine the most suitable method. For regions in a continental climate regime, and with low variability of elevation changes, a single method may be sufficient to capture the regional elevation change pattern and hence mass balance.

1 Introduction

The most recent assessment from the International Panel on Climate Change (IPPC) indicates that contributions to future sea level rise in the 21st century will be driven primarily by mass loss from ice caps and glaciers Meehl et al. (2007).

Measurements of regional and global mass balance have primarily been derived from a series of local glaciological records, Radić and Hock. (2011). These measurement are both sparse and biased to the area where they are measured Gardner et al. (2013). This makes both regional and global mass balance estimates prone to a large uncertainties, Kaser et al. (2006). With the introduction of satellite remote sensing, such as satellite altimetry, a new era has opened up. It is now possible to determine the mass changes of vast and remote areas, such as Greenland and Antarctica Shepherd et al. (2012).

The use of satellite altimetry to determine geodetic mass balance of the major ice sheets has been possible since the mid 1980's and early 1990's, by Zwally et al. (1987); Wingham et al. (1998) and others. Deriving geodetic mass balance on glacier and ice caps on the other hand is inherently more difficult due to their size and geometry, and the fact that the spatial coverage of altimetric data across these areas is usually poor. This can make the estimation of the regional mass balance of ice caps and glaciers heavily dependent on the method used to regionalize the observations, which can lead to large discrepancies in the estimated mass changes, depending on the method and region.

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Deriving geodetic mass balance from satellite altimetry is done by measuring the elevation change using repeat passes of the satellite over the same geographical region. The apparent height change measurements are then converted into volume change and mass change based on the knowledge of the glaciated regions areas and densities. Using this approach the major assumptions and generalizations lie in deriving the regional elevation changes and the conversion to mass, Huss et al. (2013). Previous altimetric studies of the ice caps and glaciers in the Arctic area, (Gardner et al., 2011; Moholdt et al., 2010a, 2012; Arendt et al., 2002; Abdalati et al., 2004), aimed at quantifying the rate of mass change, have in most cases used a single or possibly two method to derive the regional geodetic mass balance. But minor analysis have only been done to determine or quantify the effect of different regionalizing schemes for elevation changes and how they impact the regional mass balance.

The main objective of this study is to determine and quantify the effect that different methods of deriving regional elevation changes have on the mass change estimates, this by analysing the range of the different results. Mass changes will be estimated from elevation changes obtained from the Ice, Cloud, and land Elevation Satellite (ICESat) Schutz et al. (2005) during the period from 2003–2009. A total of four methods will be used to derive regional elevation changes based on interpolation and extrapolation of the elevation changes to the glaciated areas. The second objective is to determine an optimum method for deriving ice mass changes across the various Arctic regions. This assessment will be based on statistical analysis and inter-comparison of the different mass change estimates produced by the methods.

Results from this study will allow us to determine robust mass change estimates for the Arctic region, reducing the need of external validation data. It will also quantify the impact of the different regionalization procedures on the Arctic mass change estimates, and an important insight into where and how these regionalization methods should be applied to ice caps and glaciers in the Arctic region.

2 Study areas and data

To determine the regional geodetic mass balance the Arctic was divided into six different regions. In this study Greenland was excluded due to that the peripheral regions of the Greenland ice sheet have been studied in more detail in other studies, such as Bolsh et al. (2013).

The Arctic was divided into the following regions: Iceland (ICEL), Svalbard (SVLB), Russian high Arctic (RUS), Gulf of Alaska and Eastern Canada (GoA), Canadian Arctic, north and south (CAN and CAS). Due to their different climatic conditions and geographical separation the Russian high Arctic was divided into three sub-regions (Novaya Semelya, Franz Joseph Island and Severnaya Semelya). The glacier outlines for these glaciated areas was extracted from glacier shapefiles, obtained from the "Randolph Glacier Inventory" (RGI), (http://www.glims.org/RGI).

Digital elevation models (DEM's) were used to retrieve regional elevations for each region. For Iceland and Svalbard they were obtained from the National Geospatial-Intelligence Agency (NGA), with a resolution of approximately 1 km. For CAN, CAS, RUS and GoA the GTOPO30 DEM was used, also with an 1 km resolution (http://www1.gsi.go.jp/geowww/globalmap-gsi/gtopo30/gtopo30.html).

ICESat carries the Geoscience Laser Altimetry System (GLAS). The system was operating from 2003–2009 and has a repeat cycle of 96 days with a 33 day sub-cycle. The system derives range from the delay time of the transmitted laser pulse and the recieved return echo. GLAS has an average ground-track sample spacing of 172 m (along-track) and a ground footprint of approximately 70 m in diameter. The ICESat elevation data were obtained from the the National Snow and Ice Data Center (NSIDC), (http://nsidc.org/data/icesat/index.html), in the form of the GLA06 L1B global surface elevation data product. We used the latest to date product release (R33) to estimate the surface elevation change for the regions during the period of 2003–2009.

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3 Data processing

Data was processed in three main steps: In the initial step the ICESat GLA06 product was filtered using the quality flags and rejection parameters included in the product release. From this global data set we then extracted the regional data of interest. Glacier outlines were then used to extract data over the glaciated areas in each region.

Due to the fact that ICESat ground track does not have perfect spatial repetition (there can be large offsets of individual tracks from the main ground-track cluster, up to the size of a degree) we developed a graphical user interface (GUI) to visually edit out tracks with large separation. This was done to produce more robust elevation changes, from the method described in Sect. 4.1.

A cleaning procedure was initially applied to the estimated elevation changes. The first step is to remove samples with a standard deviation (estimated from the elevation change algorithm) outside the 95 % confidence bound of the entire data set. In the second step a moving hampel filter Pearson et al. (2002), was used to identify and correct outliers in the elevation changes. The filtering was done in the elevation change vs. elevation domain to easier detect outliers. If the data set contains a low number of data points, the outlier is set to the median value of the local window (to preserve data), otherwise it is removed.

The third step was to apply an along-track smoothing filter to the elevation change data. The filter was an unweighed 5-point moving average filter (with a corresponding physical filter distance of 2.5 km). The smoothing was done to remove noise in the elevation change estimates, to aid the fitting procedure for the extrapolation and surface fitting for the interpolation methods (described in Sects. 4.2.1 and 4.2.1).

4 Methods

4.1 Volume and mass change

To estimate surface elevation changes the ICESat repeat ground tracks was divided into 500 m segments. In each segment the mean elevation change is estimated by least square regression (if 6 yr of data is avaliable) and the seasonal signal is also removed. This method is described in detail in Sørensen et al. (2011) (refered to in that paper as the M3 method).

To estimate the regional elevation change, the elevation changes were regionalized and re-sampled onto a regular grid, with a grid spacing of 0.01 latitude and 0.025 longitude. This was done using several regionalization methods based on interpolation and extrapolation. The glaciated areas were then extracted using a glacier mask constructed from the RGI polygons.

The mass change is estimated by multiplying the volume change with an appropriate density of snow/ice. Due to the fact that density varies across elevation, location and time applying it is not straight forward. In this study we retrieved the mass balance by assuming that positive elevation change values are due to accumulation. Due to that the density of firn varies between 400–830 $\mathrm{kg\,m^{-3}}$, we assigned it a average density of $\rho_{\mathrm{firn}} = 500\,\mathrm{kg\,m^{-3}}$. While negative elevation change values are assumed to be due to ablation and assigned a average density of ice $\rho_{\mathrm{ice}} = 900\,\mathrm{kg\,m^{-3}}$. This assumes that the mass changes are due to effects such as major snow accumulation and ice melt, while ignoring such effects like dynamic thinning and thinning at higher elevations due to enhanced firn densification. This is a very simplified view and is not always valid, which makes it a large source of uncertainty.

4.1.1 Spatial interpolation

The first method (referred to as M1) fits a smooth surface to the scattered elevation change estimates. To obtain the regional volume change, the individual interpolated

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grid elements are summed up and multiplied with the grid cell (pixel) area.

$$\dot{V} = \sum (\dot{h}_i) \cdot A_\mathrm{p} \tag{1}$$

where $\dot{h}_i$ are the individual elevation change elements and A_p is the grid cell (pixel) area.

Due to data processing and data editing there is a loss of spatial resolution and data gaps in the along track elevation change estimates. The second method (referred to as M2) tries to overcome this by increasing the along-track resolution. To increase the number of along track samples we re-sampled the 500 m along-track data points to a new separation of 100 m, using linear interpolation. To include both the spatial and elevation dependent variations we parametrize the elevation changes using the following function:

$$\dot{h} = a_0 + a_1\Phi + a_2\lambda + a_3h + ... + a_Nh^N \tag{2}$$

where $\dot{h}$ is the parametrized elevation change value, a is the model coefficients (solved by least squares regression), h is the DEM elevation, N is the model order, Φ is the latitude and λ is the longitude. This relation was then used to estimate the elevation change at the new along-track positions. Ordinary interpolation could have been used to estimate the elevation change from the original track, but this would be less robust in areas where there is sparse track coverage or large data gaps.

The interpolation was done using least squares collocation (as implemented in the GRAVSOFT program GEOGRID, Forsberg et al. (2008). The interpolation uses a quadrant based nearest neighbour search to search for the N closest points in every quadrant around the prediction point. The data points are then interpolated by applying a second order Markov covariance model. The covariance scale of the model is found from the data and the correlation distance is varied until a smooth error surface is reached (initial correlation length taken as half the track spacing). For the six regions $N_q = 5$ number of points in each quadrant was used for the interpolation, and a correlation length of 50 km gave a sufficiently smooth error surface.

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4.1.2 Spatial extrapolation

An approach using hypsometric averaging, see Nuth et al. (2010) and Moholdt et al. (2010a), was used for regional extrapolation of elevation change estimates to regional volume change. Hypsometric averaging is based on parametrizing the elevation changes as a function of elevation where the glaciated areas is divided into elevation bands or bins and each band is assigned a representative elevation change value estimated from the parametrization to elevation. Each elevation band, with its corresponding elevation change value, is then multiplied with the glacier area-elevation distribution to obtain volume change.

$$\dot{V} = \sum \left(\dot{h}(z) \cdot A(z)\right) \tag{3}$$

where $\dot{h}$ is the specific elevation change value evaluated at the elevation bin z. Where z is defined as mid elevation of the bin (i.e 25 m if bin range 0–50 m). $A(z)$ is the area-elevation distribution at the corresponding elevation z. The area-elevation is estimated by the number of pixels inside each elevation bin.

The first method used (refereed to as M3) to parametrize the elevation changes is by fitting a polynomial function to the elevation change as a function of elevation, as in Nuth et al. (2010) and Moholdt et al. (2010a). The elevations are obtained from the glacier masked DEM's for every region (see Sect. 2.). The DEM's are divided into elevation bins and the elevation change for each bin is estimated from the polynomial function. Hypsometric averaging is then used to extrapolate the elevation changes regionally.

The second extrapolation method (refereed to as M4) used to parametrize the relation to elevation is by binning the elevation changes according to elevation (as in M3), but instead of estimating the centre bin elevation change from a continuous function we instead use the median value of the elevation changes inside the bin, as in Abdalati et al. (2004). Empty bins (due to lack of data at that specific elevation band) are estimate by linear interpolation. DEM elevations not covered by the ICESat data (usually low and high elevations) are extrapolated by fitting a linear function to these bins (estimated from the entire data set).

To determine the degree and the number of terms in the polynomial, we need a measure of how much variance the model is able to account for. The more variability that can incorporate into the model the better it will explain the underlying dynamics of the measured data. We use the adjusted R^2 statistics as a measure of incorporated variance, see Moholdt et al. (2010a). The degree of the polynomial and the number of parameters are then increased until a convergence of this ratio is reached. For all regions, except Svalbard, a linear fit ($D = 1$) was sufficient to parametrize the relation. For Svalbard a third order polynomial ($D = 3$) fit the distribution best in a R^2 sense (same used by Moholdt et al., 2010a). Varying the elevation bin size for the hypsometry had only a small effect on the regional geodetic balance. So an elevation bin size of 50 m was chosen for all regions. This was also seen by Gardner et al. (2011).

4.2 Determining optimum regional method

To determine the optimum regional method for estimating the geodetic mass balance we define a statistical criterion. This criterion measures the absolute shift in mean value (shift in histogram) away from the mean of the original elevation change distribution from ICESat.

$$\Delta\mu = |\mu_o - \mu_m| \tag{4}$$

where $\Delta\mu$ is the absolute shift between the mean values, μ_o is the original mean and μ_m is the mean of the inter or extrapolated values.

Measuring the shift produced by the methods we can determine the optimum method(s) as follows: (1) By choosing the method with the smallest mean shift to the original mean. (2) If there is a large range of values a combination of different methods can be chosen that gives the smallest shift.

This approach assumes that the ice cap or glacier geometry is fully resolved by the ICESat ground track, so that the histogram produced from the ICESat elevation changes is representative of the entire region. This assumption is not always valid for all regions, as it is a function of spatial coverage and the number of samples. For most Arctic regions it is a valid assumption due to that the high latitude makes the satellite ground tracks converge.

5 Error analysis

We have based the error analysis on two main concepts central in error analysis: The standard deviation around the mean and the standard error of the data, which follows Nuth et al. (2010) and Moholdt et al. (2010a) approach. There have been extensive studies to quantify the individual point measurement error for ICESat over ice covered regions. Brenner et al. (2007) found that the ICESat error over ice sheets are in the range of 0.14–0.5 m. We have taken a more conservative approach assume an error of $\varepsilon_{\text{icesat}} = 1$ m, as in Nuth et al. (2010). There also exist a inter-campaign bias in the ICESat data set Siegfried et al. (2011). This bias has not been quantified in R33, but we assume that this should be included in the conservative error estimate of the ICESat error.

There is also a need to estimate the error from the elevation change estimation procedure. Here we use the standard deviation calculated from the least square solution as a measure of how trustworthy the individual elevation change measurements are Sørensen et al. (2011). ICESat elevation changes are correlated along track due to the similarity in surface topography, atmosphere and satellite related effects. To reduce this correlation effect previous studies have either: Divided the tracks into segments (Moholdt et al., 2010a) and assumed that the data are fully correlated inside the segment and that the individual segments are uncorrelated. Or divided them into elevation bins and assumed that the individual bins are uncorrelated Nuth et al. (2010). In this study we combined them both and divided the tracks into elevation bin segments, which are assumed to be un-correlated. The number of segments can then be used to estimate

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the standard error.

$$\varepsilon_{\mathrm{d}h/\mathrm{d}t} = \frac{\sigma_{\mathrm{d}h/\mathrm{d}t}}{\sqrt{N}} \tag{5}$$

where N is the number of un-correlated segments and $\sigma_{\mathrm{d}h/\mathrm{d}t}$ is the mean standard deviation. Here the $\sigma_{\mathrm{d}h/\mathrm{d}t}$ is previously reduced by a factor $1/\sqrt{N_\mathrm{s}}$ due to the along track smoothing, were N_s is the size of the smoothing filter.

Next we need to estimate the error from the collocation (interpolation) procedure. We estimated this by computing the mean standard deviation from the total number of prediction points. The standard error is then estimated by using the number of sub-rectangles containing data created by the GEOGRID algorithm. These individual sub-rectangles are then assumed to be uncorrelated.

$$\varepsilon_{\mathrm{int}} = \frac{\sigma_{\mathrm{int}}}{\sqrt{N}} \tag{6}$$

where N is the number of sub-rectangles containing data and σ_{int} is the individual standard deviation from the collocation prediction of non-zero data.

Then we quantify the parametrization error from the fitting of the polynomial function. This error can be estimated by calculating the root mean square error (RMSE) between the original elevation changes and the predicted values.

$$\varepsilon_{\mathrm{fit}} = \frac{\sigma_{\mathrm{fit}}}{\sqrt{N-D}} \tag{7}$$

where σ_{fit} is the RMSE between the original and modelled data, $\sqrt{N-D}$ is the adjustment due to the degree of the polynomial.

After this we need to quantify the extrapolation error $\varepsilon_{\mathrm{ext}}$. We have in this study adopted the approach used by Nuth et al. (2010) were the extrapolation error is the root sum square (RSS) difference of the fitted error minus the elevation change error.

$$\varepsilon_{\text{ext}} = \sqrt{\varepsilon_{\text{fit}}^2 - \varepsilon_{\text{d}h/\text{d}t}^2} \tag{8}$$

The extrapolation error for the median binning method is also to be determined. This will be refereed to as the binning error, ε_{bin}, so not to confuse it with ε_{ext}. This error is defined as the standard deviation inside every elevation bin. The standard error is then calculated by assuming that the individual bins are uncorrelated

$$\varepsilon_{\text{bin}} = \frac{\sigma_{\text{bin}}}{\sqrt{N}} \tag{9}$$

We also have to add an error term for the density ratio. Because the density is not uniform over the area of the glaciated region. It varies as a function of position and time Moholdt et al. (2010b).

$$\varepsilon_{\rho} = \frac{1}{2}(\rho_{\text{ice}} - \rho_{\text{firn}}) \tag{10}$$

where ρ_{ice} and ρ_{firn} is the density of ice and firn respectivly.

The corresponding height error ε_{h} is then estimated by RSS of the individual error sources described in Table 1.

The volumetric error can then be estimated by multiplying the height error with the regional area.

$$\varepsilon_{\text{vol}} = \varepsilon_{\text{h}} \cdot A \tag{11}$$

Finally we can estimate the mass balance error as follows:

$$\varepsilon_{\text{mass}} = \sqrt{(\varepsilon_{\text{vol}} \cdot \rho)^2 + (\dot{V} \cdot \varepsilon_{\rho})^2} \tag{12}$$

where $\dot{V}$ is the estimated volume change.

6 Results

The Arctic ice caps show a consistent pattern of large peripheral thinning and thickening in the interior regions (see Figs. 3 and 4). The thinning is mostly located in the low elevation areas of the ice caps and glaciers ($h < 500$–800 m) and becomes more positive as the elevation increases. This pattern has previously been observed in studies of Greenland, Antarctica and Svalbard, see Sørensen et al. (2011), Wingham et al. (1998), Krabill et al. (2000), Pritchard et al. (2009), Bamber et al. (2004).

The lower elevations in every region show large variability ($h < 500$–800 m) which are clustered around the coastal regions. These clusters are located in areas which comprise of relatively fast flowing outlet glaciers.

The different regions exhibits very different rates and patterns of elevation change (see Table 2). Regions such as RUS, CAN and CAS all show low variability and low area-averaged volume change. While regions such like ICEL and GoA show a high variability and high rate of area-averaged volume change. SVLB exhibits its own unique behaviour where it shows little area-average volume change but has a large variability compared to other regions. Comparing the area-averaged volume change we observe that the highest rate of change are located in GoA and ICEL while we observe the lowest rates in SVLB and RUS. Examples of this can be seen in Fig. 1. where regions located in the lower latitudes of the Arctic regions (such as ICEL and GoA) show a large spread in the elevation change estimates. While the high(er) Arctic show a lower and more Gaussian (symmetric) distribution of elevation changes.

Studying ICESat's sampling in both the spatial and elevation domain (Figs. 2 and 3) we observe that the data density is skewed to the mid-elevation range (illustrated by the red curve in Fig. 2). Worth noting is that in low elevation bands (where most of the variability is located) we usually have quite poor sampling. Regions excluded from this are SVLB and RUS which both show higher sampling of the low elevation regions ($h < 500$ m).

Examining the geodetic mass balance results from the different methods (see Table 4), we can see that range of the results varies to large degree. Regions with high variability such as SVLB, GoA and ICEL also show a large geodetic mass balance range. Where the typical range is on the order of the error estimates or higher. They are on average two times higher for these areas. Regions with a more uniform pattern and lower variability of elevation changes show typically a range well below the error estimates. The largest range of geodetic mass balance can be seen for GoA ($-55\,\mathrm{Gt\,a^{-1}}$). This region also has the largest spatial variability ($1.3\,\mathrm{m\,a^{-1}}$). RUS shows the smallest geodetic range of only $0.7\,\mathrm{Gt\,a^{-1}}$ for the period. It also has the lowest variability of all the regions ($0.33\,\mathrm{m\,a^{-1}}$). There is a strong relation between the range of the geodetic mass balance and the variability of the elevation change estimates. This strength of the relation (using all range values) corresponds to a correlation coefficient of $\rho_{\mathrm{corr}} = 0.67$. If Alaska (with its extreme magnitude of range compared to the other regions) is excluded this relation increases up to $\rho_{\mathrm{corr}} = 0.75$.

Studying Table 3 we find that there is a systematic bias between the extrapolation and interpolation methods. The extrapolation methods tends in general to give more negative estimates of the geodetic mass balance compared to the interpolation methods. This pattern is broken for SVLB which shows the inverse of this relation. On average (when excluding SVLB) we find that the extrapolation methods tend to give a 12 % more negative value of the geodetic mass balance. This pattern is also consistent within the mean values produced by the different methods. Where they all produce a mean value more negative than the original mean.

The impact of correcting for variable density on the regional geodetic balance showed only minor effects for most regions. This is due to the fact that most positive elevation change estimates are located at higher elevation and these higher elevations only account for a small percentage of the total area. The only case this was not true was for Svalbard. Its more complex elevation change pattern (see Fig. 2) gave rise to an almost 50 % difference in geodetic balance, compared to using a density of $\rho = 900\,\mathrm{kg\,m^{-3}}$.

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The optimum regional results in Table 3 was determined using Eq. (4) and the following combination(s) of methods: SVLB: M4, ICEL: M1, CAN: all, CAS: all, GoA: M1 and RUS: all.

7 Discussion

The large degree of variability seen in the lower elevations in Fig. 3 for most regions indicates that dynamic behaviour has been captured in the data (as it decreases as a function of elevation). The location of these data ($h < 500$–800 m) are mostly clustered along the coast and to areas with large drainage systems. These areas are usually highly active and show the most rapid changes.

These dynamical components can help explain the systematic bias that we observe between the interpolation and extrapolation methods. The extrapolation methods tend in general to produce a more negative geodetic mass balance due to its parametrization to elevation. The polynomial or linear model can not capture the usually large variability in the lower elevation, due to the large dynamic component. By only capturing the trend of the data in this region the area-elevation distribution can have a important effect on the total geodetic mass balance. On the other hand the interpolation method shows on average a lower estimate due to its smoothing criterion. The smoothing criterion prevents us from capturing the full amplitude of the minimum and maximum values of the data when fitting the interpolation surface.

As listed in Table 3, we find that we find that both the interpolation and extrapolation methods, determined by the optimum approach, produces higher thinning rates (more negative means) than the original elevation changes. This indicates that the mass loss of the different regions, from ICESat, might be overestimated on the order of 4–19 % (depending on the region), with respect to the original mean.

The determination of the optimum regional method is based on the assumption that the ICESat sampling fully resolves the ice cap geometry which in many cases is usually not the case. This mostly in areas in the highest elevations in the DEM (not covered by

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ICESat points). These areas have to be interpolated or modelled but due to there small size they play a insignificant part in the geodetic mass balance. So for many areas this assumption is mostly valid. This is not true however for Iceland and Alaska.

Due to its low latitude, Iceland shows only a few ICESat tracks that sparsely covers the ice cap geometry. Thus there is a low number of samples to determine its elevation change pattern (seen in Fig. 2a). Observed in Fig. 2a is a large variability of the elevation changes in both the lower ($h < 1000$ m) and the higher ($h > 1500$ m) elevations. The low number of data samples and large variability of the data causes the optimum method to break down as there are too few data values to create robust statistics.

Comparing the results to Björnsson et al. (2013) we find that the extrapolation methods clearly overestimates the mass balance while the interpolation methods captures the (even with the poor spatial sampling) elevation change pattern. This is likely due to: (1) Even though the ICESat spatial sampling is poor the ground tracks transects the ice caps geometry evenly. This reduced the risk of biased observations due to location and elevation. (2) There is a clear symmetry of the elevation changes (clear linear relation in elevation) which creates a bell like geometry that is easier to fit a surface to. This would indicate that the interpolation methods should be used here. They give better agreement with the in-situ derived mass balance.

The same arguments can be applied to Alaska with its large variability of elevation change. Most prominent here is the lack of spatial sampling, due to the regions low latitude and mountain/valley type glaciers. This low sampling (2643 points) gives rise to large uncertainties in the estimated geodetic mass balance which can be clearly seen by observing the range of the different estimates. The optimum estimate derived from our study falls inside previous estimated results by Arendt et al. (2002) and Luthcke et al. (2008). A more previous study Berthier et al. (2010) have though indicated that the these values might be overestimated by roughly 35 % due to the spatial sampling issue.

There is a strong relationship (positive correlation) between the range of the geodetic mass balance estimates and the effects of inter-annual variability (ice dynamics, snow accumulation and surface melt) on the elevation changes observed, where the inter-annual variability is a function of the climate regime of the different regions. Regions such as the Canadian Arctic and the Russian High Arctic are regions with a continental climate regime (dry and cold). These regions are characterized by a low variability ($\approx 0.35\,\mathrm{m\,a^{-1}}$) and a spatially uniform pattern of elevation change. With the low variability and uniform pattern the different methods tend to converge to a small range of values of mass balance.

This knowledge can be used to determine the sensitivity of the geodetic mass balance due to different methods. One should first consider the regional climate type as a first indicator and then the variability of the elevation changes. If the variability of the elevation changes are larger than $0.5\,\mathrm{m\,a^{-1}}$ one should consider using several independent methods in the estimation of the regional geodetic mass balance.

Converting volume change to mass change can introduce significant uncertainty when estimating mass balance from satellite altimetry. The approach taken here assumes a ELA-like relationship not based on elevation but on sign of the surface elevation changes. For this study we have altimetry data from nine years of observations. This makes the assumption that a positive sign of the elevation change actually indicates a accumulation zone and not just a random change in surface elevation more stringent. The regional approach also ensures that the error introduced, by these randomized variations in the pattern, have less impact on the distribution. For most regions this has little impact but for Svalbard with its more complex pattern of elevation change this plays a substantial role. One should investigate the elevation change pattern before choosing the density conversion scheme. This can lead to significant differences in geodetic mass balance given the same data sets.

Even though smoothing has been performed on the elevation changes, there are still clear dynamic signals evident in the elevation changes in Fig. 3. These signals are located in the lower ranges of elevation in the form of high variability, which is usually the main issue for the extrapolation methods. Parametrizing the elevation changes as a function of elevation the polynomial captures very little of this variability (captures the

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trend). These low elevation variability usually holds a substantial part of the glaciated area. This usually produces a result that for many cases is overestimated. So for many cases the interpolation methods would yield more realistic results due to its ability to account for more of the spatial variability (given enough spatial sampling). An improvement can though be made in the case of the parametrization using a continuous function (M3). This would be to include a spatial dependency of the data, as done in Eq. (2). Thus including more of the variability, due to position, of the elevation changes.

In general the Arctic shows large regional differences in both variability and rates of change. This propagates into the geodetic mass balance due to different methods tend to capture different parts of the elevation change pattern. In our study we have shown that the range of the results are for many regions larger than the error estimates. This is an important fact to consider when calculating the geodetic mass balance. One should implement caution when choosing a particular methods to derive regional geodetic balance. This can have an significant impact on the result and the following interpretation. We suggest that several methods should be used to derive the regional geodetic balance, especially in areas with maritime climate and with large spatial variability of the elevation changes.

8 Conclusions

In this study we quantified the impact on mass change estimates due to different regionalization schemes. We further determined optimum methods for each region based on statistical analysis of the different estimated results.

We found that the range in the different values of mass change is strongly correlated and mostly driven by the inter-annual variability of the elevation changes, which is a function of the regional climatic regime. Regions with a maritime climate showed a large spread in the estimated mass changes, which were on average two times larger than the estimated errors. For regions in a continental climate regime the opposite was observed and the range of mass change estimates lie well within the error estimates.

Also noted in the study was that the extrapolation methods on average produces more negative mass balance estimates compared to the interpolation methods. Where both the extrapolation and interpolation produce histograms with more negative means than the original mean. This would indicate that the mass change estimates from previous and this current study, using ICESat observations might be overestimated with as much as 4–19 %, depending on the region.

One should consider the following when deriving regional geodetic mass balance from satellite altimetry: (1) Climatic regime and (2) the variability of the elevation changes. If the variability of the elevation changes exceeds $0.5\,\mathrm{m\,a^{-1}}$ and the region exhibits a maritime climate, one should use several different independent methods to derive the geodetic mass balance. On the other hand if the region is affected by a continental climate regime one method is usually sufficient to capture the regional elevation change pattern.

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Table 1. The number of error terms present in each method. These error are then combined into a height error, using RSS.

Method	Error Terms
M1	$\varepsilon_{\text{icesat}}$, ε_{int}, $\varepsilon_{\text{d}h/\text{d}t}$
M2	$\varepsilon_{\text{icesat}}$, ε_{int}, $\varepsilon_{\text{d}h/\text{d}t}$, ε_{fit}
M3	$\varepsilon_{\text{icesat}}$, ε_{ext}, $\varepsilon_{\text{d}h/\text{d}t}$
M4	$\varepsilon_{\text{icesat}}$, ε_{bin}, $\varepsilon_{\text{d}h/\text{d}t}$

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Table 2. Statistics for all Arctic regions. Where μ is the mean elevation change of the region, σ the standard deviation, A is the area, $\dot{h}_A$ the area average thinning rate (of the optimum methods) and N is the number of elevation changes.

Region	μ [m a^{-1}]	σ [m a^{-1}]	A [km^2]	$\dot{h}_A$ [m a^{-1}]	N
SVLB	−0.02	0.66	33 673	−0.025	5287
ICEL	−0.77	1.26	10 989	−0.85	943
CAN	−0.37	0.34	103 990	−0.40	18 472
CAS	−0.67	0.39	40 601	−0.70	3499
GoA	−0.95	1.30	84 926	−1.00	2643
RUS	−0.15	0.33	51 161	−0.16	9213

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Table 3. Geodetic mass balance $\dot{m}$ from the four methods for the different Arctic regions, with there corresponding errors (σ).

Region	Method	$\dot{m}$ [$\mathrm{Gt\,a^{-1}}$]	σ [$\mathrm{Gt\,a^{-1}}$]
Svalbard	M1	−4.5	1.9
	M2	−5.9	2.1
	M3	−3.8	1.9
	M4	−2.8	1.7
	Range	**3.1**	**0.4**
	Optimum	**−2.8**	**1.7**
Iceland	M1	−8.7	2.1
	M2	−10.1	2.4
	M3	−12.5	3.0
	M4	−13.1	3.1
	Range	**4.4**	**1**
	Optimum	**−8.7**	**2.1**
Canadian Arctic North	M1	−36.7	8.2
	M2	−34.8	7.8
	M3	−37.9	8.6
	M4	−38.1	8.6
	Range	**3.3**	**0.8**
	Optimum	**−37**	**8.3**
Canadian Arctic South	M1	−24.8	5.7
	M2	−25.4	5.7
	M3	−26.0	5.9
	M4	−26.0	5.9
	Range	**1.2**	**0.2**
	Optimum	**−25.5**	**5.8**
Gulf of Alaska	M1	−77.2	18
	M2	−110.0	25
	M3	−132.2	30
	M4	−84.7	20
	Range	**55**	**12**
	Optimum	**−77.2**	**18**
Russian High Arctic	M1	−8.9	2.4
	M2	−9.2	2.4
	M3	−9.6	1.2
	M4	−9.2	1.2
	Range	**0.7**	**1.3**
	Optimum	**−9.2**	**1.9**

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Table 4. Final geodetic mass change estimates derived using the optimum approach described in Sect. (4.2).

Region	$\dot{m}$ [$\mathrm{Gt\,a^{-1}}$]	σ [$\mathrm{Gt\,a^{-1}}$]
SVLB	−2.8	1.7
ICEL	−8.7	2.1
CAN	−37.0	8.3
CAS	−25.5	5.8
GoA	−77.2	18
RUS	−9.2	1.9

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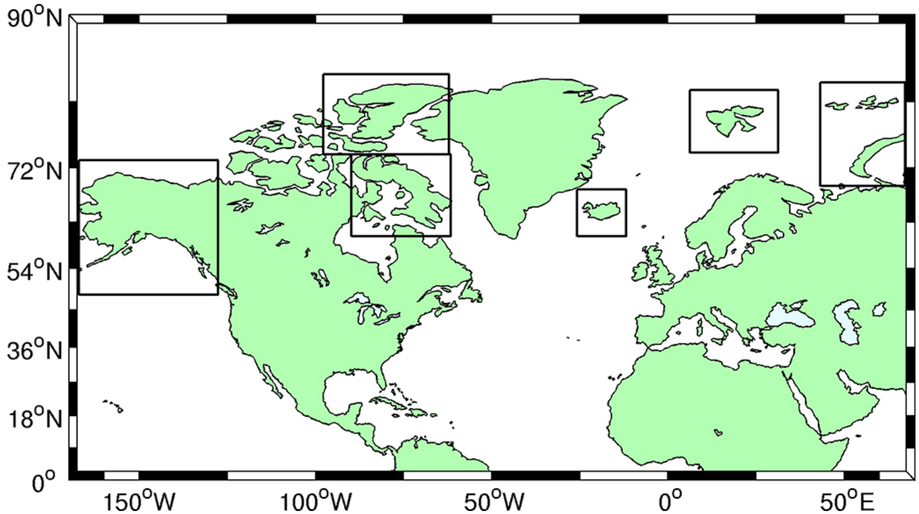

Fig. 1. Selected geographical areas in the Arctic region.

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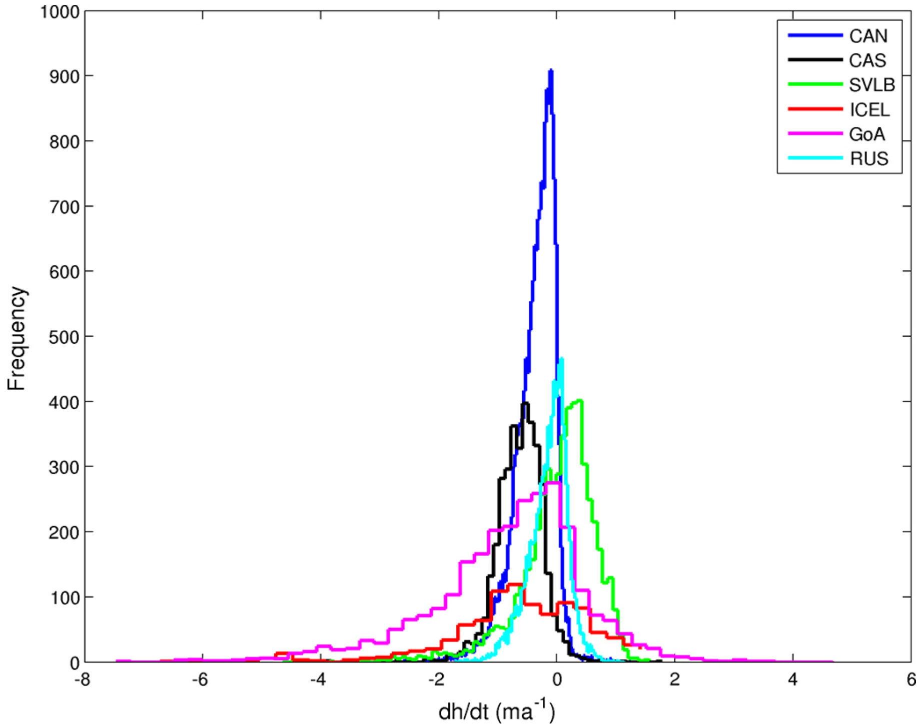

Fig. 2. Histogram of elevation changes for the different Arctic regions. The Russian High Arctic (RUS) is treated as one region.

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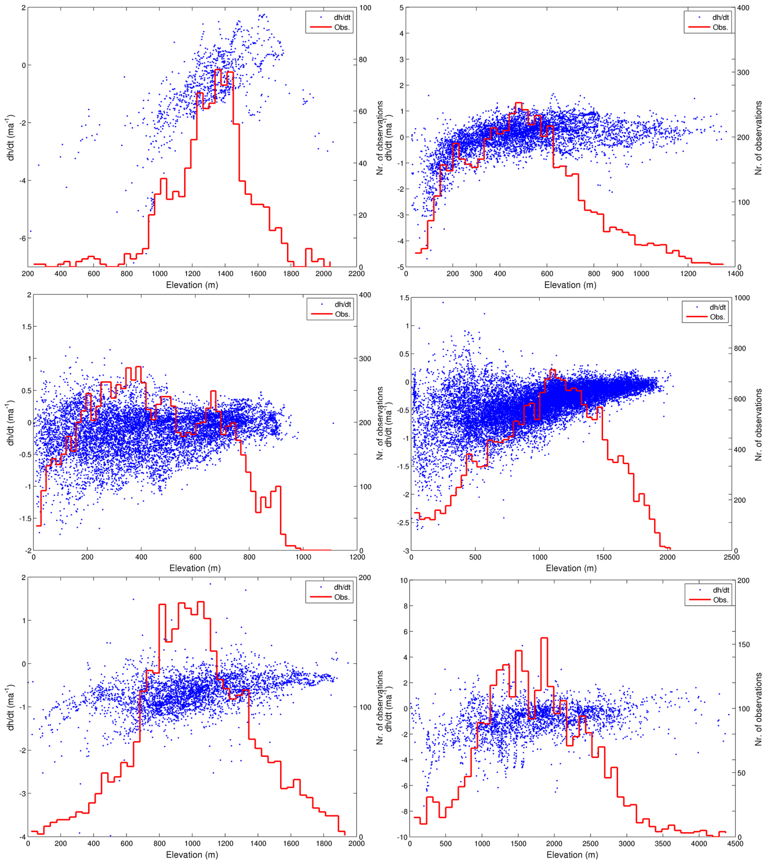

Fig. 3. Elevation change (blue points) as a function of elevation for the different Arctic regions which are used for regional extrapolation. The red curve represent the density of ICESat's elevation sampling.

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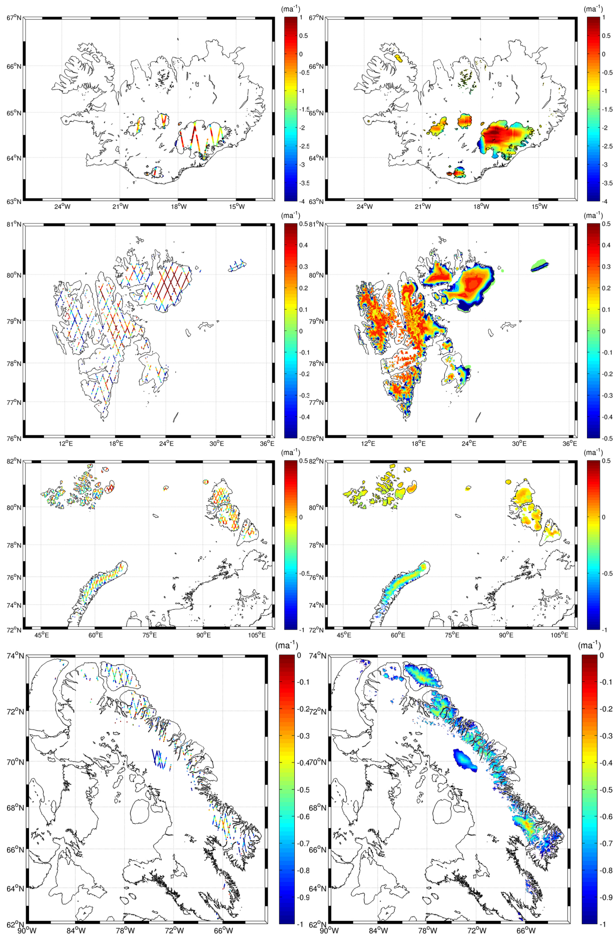

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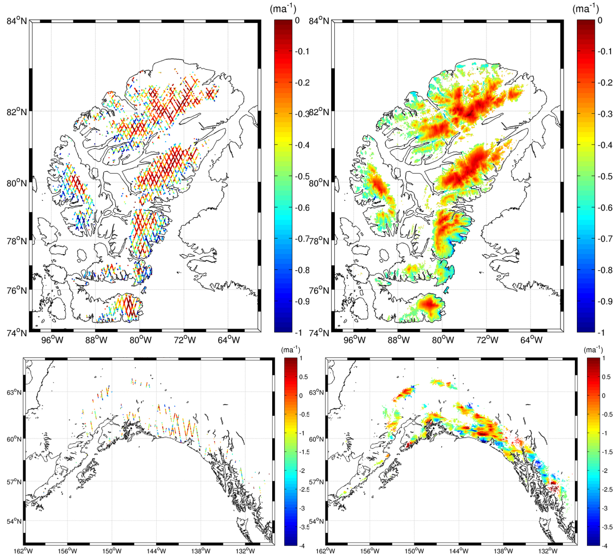

Fig. 4. Spatial pattern of elevation changes in the Arctic region in the form of satellite ground track coverage and gridded values, where the gridded values are estimated using the optimum method approach (see Sect. 4.2). (**a** and **b**): Iceland, (**c** and **d**): Svalbard, (**e** and **f**): Russian High Arctic, (**g** and **h**): Canadian Arctic South, (**i** and **j**): Canadian Arctic North and (**k** and **l**): Gulf of Alaska.

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