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Improved retrieval of land ice topography from CryoSat-2 data and its impact for volume change estimation of the Greenland Ice Sheet

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Abstract

A new methodology for retrieval of glacier and ice sheet elevations and elevation changes from CryoSat-2 data is presented. Surface elevations and elevation changes determined using this approach show significant improvements over ESA's publically available Cryosat-2 elevation product (L2 Baseline-B). This when compared to near-coincident airborne laser altimetry from NASA's Operation IceBridge and seasonal height amplitudes from the Ice, Cloud, and Elevation Satellite (ICESat).

Applying this methodology to CryoSat-2 data collected in Interferometric Synthetic Aperture mode over the high relief regions of the Greenland ice sheet we find an improvement in the root-mean-square-error (RMSE) of 27% and 40% compared to ESA's L2 product in the derived elevation and elevation changes, respectively. In the interior part of the ice sheet, where CryoSat-2 operates in Low Resolution Mode, we find an improvement in the RMSE of 68% and 55% in the derived elevation and elevation changes, respectively. There is also an 86% improvement in the magnitude of the seasonal amplitudes when compared to amplitudes derived from ICESat data. These results indicate that the new methodology provides improved

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tracking of the snow/ice surface with lower sensitivity to changes in near-surface dielectric properties.

To demonstrate the utility of the new processing methodology we produce elevations, elevation changes and total volume changes from Cryosat-2 data for Greenland Ice Sheet during the period Jan-2011 to Jan-2015. We find that the Greenland Ice Sheet decreased in volume at rate of 289 ± 16 km^3 a^{-1}, with high inter-annual variability and spatial heterogeneity in rates of loss. This rate is 65 km^3 a^{-1} more negative than rates determined from ESA's L2 product, highlighting the importance of Cryosat-2 processing methodologies.

1 - Introduction

The European Space Agency (ESA) launched CryoSat-2 in April 2010 tasked with monitoring the changes of the Earth's land and sea ice. CryoSat-2 carries a new type of Doppler/delay radar altimeter (Raney, 1998) referred to as SIRAL (SAR Interferometric Radar Altimeter). SIRAL operates in two different modes over land ice. Over the interior part of the ice sheets it operates as a conventional pulse limited radar system, referred to as the "Low Resolution Mode" (LRM). In more complex high-sloping terrain the system uses a novel second antenna to operate in "Interferometric Synthetic Aperture Radar" (SIN) mode. These new features allow the satellite to monitor changes in complex terrain including ice caps, glaciers and the high relief marginal areas of the ice sheets. Such areas are sensitive to changes in climate and contribute greatly to current rates of sea level rise, e.g., Gardner et al. (2013) and Shepherd et al. (2012).

Ku-band radar altimeters are insensitive to cloud cover providing superior coverage to laser altimeters (e.g., ICESat) but experience significant amounts of volume scattering, the characteristics of which is controlled by the time-evolving dielectric properties of the near-surface snow, firn, and ice (Lacroix et al., 2008; Remy et al., 2012). These effects can have large implications for the determination of mass change over a wide range of both spatial and temporal scales. Changing snow conditions can introduce time-varying biases in the data that, in combination with the radar signals interaction with the surface, introduce large elevation biases (0.5 - 1 m) (Nilsson et al., 2015a). This, combined with other factors such as processing methodology and surface topography, makes it difficult to measure small changes for much of the word's ice covered regions (Arthern et al., 2001; Gray et al., 2015; Nilsson et al., 2015b).

The mitigation of these effects in the processing of radar altimetry data is required for improved accuracy of derived temporal and spatial changes in surface elevation of glaciers and ice sheets. Several studies have proposed different approaches to assess these effects and improve the retrieval process of surface elevation and elevation changes from radar altimetry data. These include different approaches to waveform retracking (Davis, 1993, 1997; Gray et al., 2015; Helm et al., 2014) and empirical corrections to the estimated surface elevation changes (Davis and Ferguson, 2004; Flament and Rémy, 2012; Sørensen et al., 2015; Wingham et al., 2006b; Zwally et al., 2005, 2011). Relatively little work has been done to assess methods from improving elevation and elevation changes derived from ESA's CryoSat-2 data (Abulaitijiang et al., 2015; Gray et al., 2013, 2015; Helm et al., 2014).

Here we conduct a thorough analysis of CryoSat-2 SIN and LRM waveform retracking and geolocation methodologies to design an optimal processing methodology for CryoSat-2 elevation retrieval over both smooth and complex ice-covered terrain. We then analyze several different approaches to determining surface elevation and volume changes from the scattered CryoSat-2 elevation retrievals. The overarching goal of this work is to develop robust and accurate elevation retrieval algorithms that are less sensitive to changes in surface and sub-surface scattering properties.

The new processing scheme is applied to estimate elevation and volume changes of the Greenland Ice Sheet for the period January 2011 to January 2015 that are compared to change estimates obtained from the ESA L2 Baseline-B surface elevation product (Bouzniac et al., 2014), high accuracy airborne data from NASA IceBridge airborne topographic mapper and seasonal height amplitudes estimated from Ice, Cloud, and Elevation Satellite (ICESat) data.

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2 - Surface elevations from CryoSat-2

2.1 - Low Resolution Mode (LRM)

The LRM mode is used over the interior parts of the ice sheet, which mostly consist of low sloping terrain (0-1°). Here, SIRAL operates as a conventional pulse limited radar system with a transmission frequency of 13.6 GHz (Ku-band) and has Pulse-Limited Footprint (PLF) radius of approximately 1.5 km and a beam-limited footprint (BLF) radius of approximately 7.5 km over flat terrain (Bouzinac, 2014). The gentle terrain allows for accurate mapping of the surface elevation of the ice sheet down to decimeter-level (Brenner et al., 2007). Within the LRM waveform we define the location of the surface from the leading edge of the waveform, based on a fraction of the maximum amplitude of the received power. This approach is commonly referred to as a threshold retracker. Following Davis et al. (1997) we use 20% threshold to define the location of the surface. Davis et al. (1997) argued that a 20% threshold represents the best compromise between waveforms that are entirely dominated by either volume or surface scattering, making it suitable for obtaining estimates of surface elevation for most parts of the Greenland Ice Sheet.

The CryoSat-2 LRM radar waveforms suffer from measurement noise, in the form of speckle noise. Furthermore, over the steeper parts of the LRM-area the range gate tracking-loop can loose track of the surface, producing non-usable waveforms. To remove bad or loss of track waveforms the radar waveform (20 Hz) is first filtered using a zero-phase low pass filter to reduce speckle noise on a line-by-line basis. The signal-to-noise-ratio (SNR) of the waveform is then estimated and if the SNR < 0.5 dB the waveform is rejected. The SNR threshold was empirically chosen to obtain a good trade-off between the quality of the measurements and sampling.

Before the waveform can be retracked the first surface return (first major peak) is identified within the range gate window. A copy of the waveform is heavily smoothed to remove small-scale surface roughness signals, keeping the overall surface signal intact. The range gate index of the first peak from the copy is then used to extract the leading edge of the original low pass filtered waveform. Only leading edges with a peak index within the interval of 20-350 are used in the retracking, as peaks before or after that indicate troublesome surface ranging. The extracted leading edge is then oversampled by a factor of 100 (c.f. (Gray et al., 2013; Helm et al., 2014), and the range R between the surface and satellite is determined based on the 20% threshold computed according to Davis et al. (1997). The range is then corrected for several atmospheric and geophysical effects relevant to land ice studies according to Bouzinac (2014). The surface elevation H of the topography, relative to the WGS84 ellipsoid, is estimated as $H = A - R$, where A is the altitude of the satellite.

The measured surface return over a sloping surface does not originate from the satellites nadir location, but from the "Point Of Closest Approach" (POCA) to the spacecraft (Brenner et al., 1983). These off-nadir returns can introduce a large range bias to the surface, depending on the magnitude of surface slope, ranging from 0-120 m (Brenner et al., 1983) as the measured surface height is mapped to an erroneous position (i.e. the nadir position). To mitigate the effect of this error we correct the measured range and location to the POCA point using an a-prior DEM, following the approach of Bamber (1994). In contrast to previous studies we account also for the local surface curvature, as Remy et al. (1989) showed that accounting for surface curvature in addition to surface slope significantly improve results. The surface slope, aspect and curvature are estimated from an a priori DEM. The GIMP elevation model (Howat et al., 2014) was used to derive surface parameters for the slope-induced error correction in the LRM mode. The DEM was resampled to 2 km resolution, using bilinear interpolation, prior to parameter estimation, corresponding to the pulse-limited footprint of the LRM mode.

2.2 - Interferometric Synthetic Aperture Radar Mode (SIN)

The SIN mode is used over the marginal areas of the ice sheets and other smaller glaciated areas. In these areas the SIRAL altimeter operates as a Doppler/Delay radar system (Keith Raney, 1998). The Doppler/Delay radar allows for higher along-track resolution compared to conventional altimetry. SIN modes allows for a 350 m resolution in along track and 1500 m across track. In ordinary SAR operation only the amplitude of the radar echo is measured and the phase content is discarded or ignored. With the inclusion of a second antenna on CryoSat-2, interferometric SAR can also be performed. Difference in the path length between the POCA and the individual antennas introduce a phase shift between the two retrieved signals that can be related to the angle of arrival (look angle). The look angle can in turn be used to resolve the across track (across antenna) location of the echo.

Multi-look processing is applied to ESA's L1B waveform product (Bouzinac, 2014) to reduce the noise in the SIN waveform but it is still affected by speckle-noise, as is the case for the LRM waveforms. To mitigate this effect, and to help identify the leading edge of the first return, we apply speckle reduction filtering and leading edge extraction of the SIN waveforms in the same way as for the LRM processing with minor changes due to differences in range gate resolution. The estimated coherence C of the multi-looked waveforms is then filtered in two stages; (i) all coherence measures larger than one is set to zeros (larger than one coherence exists in the L1B product reason unknown). (ii) the coherence range power image is filtered using a 2D 5x5 Wiener filter to remove high frequency noise. The filtering of the waveform and the coherence is applied to remove noise in the recreation of the interferogram. This is further discussed later.

The measured differential phase ϕ of the return signal is affected by phase ambiguities; a sudden shift of 2π in the measured phase. To reduce phase noise and aid the phase, an unwrapping of the radar interferogram I is performed according to Gray et al. (2013):

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$$I = P \cdot C \cdot e^{-i\phi} \tag{1}$$

The interferogram is then filtered using a wavelet-based de-noising technique, where the real and imaginary parts of the interferogram are filtered separately. The unwrapping of the interferogram allows for indirect filtering of the phase, without being affected by the phase-ambiguities. Phase filtering is an important consideration as it has a direct affect on accuracy of the position of the ground echo. We selected a bi-orthogonal as the mother wavelet to produce the wavelet coefficients decomposed into three levels. Soft thresholding was applied to detail coefficients, using a heuristic threshold rule to remove noise at every level. This was done on a line-by-line basis. The final filtered differential phase was then recovered by:

$$\phi_f = Re\{I_f\} + Im\{I_f\} \tag{2}$$

To resolve the phase ambiguities the filtered phase measurements require unwrapping. The phase unwrapping is done on a line-by-line basis in two directions starting from the center of gravity of the waveform (Wingham et al., 1986).

The return power distribution of a Doppler/Delay radar system shows an important distinction from those from conventional pulse-limited radar systems. Here, the point corresponding to the mean surface is not located at the half-power point on the leading edge, but rather closer to the maximum (Wingham et al., 2006a). Therefore a new retracker has been developed, closely related to the one used in Gray et al. (2013), to allow for adaptive retracking of the upper parts of the leading edge of the SAR waveform. The algorithm follows the main concept of the threshold retracker, developed by (Davis, 1997), but instead of a pre-defined threshold it tracks the maximum gradient of the leading edge of the waveform. We refer to this approach as that “Leading-edge Maximum Gradient retracker” (LMG).

The surface returns are geolocated using the across track look-angle θ estimated from the differential phase at the retrackig point according to (Wingham et al., 2006a). This, in combination with the viewing geometry, is used to define the location of the surface return on

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the ground using basic across track interferometric principles. We correct θ for the interferometer surface slope error by applying the look-angle scaling factor estimated in (Galin et al., 2013).

The along-track differential phase estimate, interpolated to the retracking point, is affected by phase ambiguities not corrected for during the phase unwrapping procedure. To reduce residual phase ambiguities an a priori DEM (GIMP) is used to extract the DEM surface, resampled to 500 m resolution (corresponding to the along-track sampling), elevations at the nadir and echolocation using bilinear interpolation. Over a sloping surface the surface return should always come from a position upslope from the nadir point. Therefore the following relation must hold where ($H_{echo} > H_{nadir}$) or for a more practical application ($H_{echo} - H_{nadir}$) > ε, where ε is the uncertainty of the DEM used. If this relation is violated 2π is added or subtracted to the individual along-track phase estimate, depending on the sign.

A final step is applied to correct for any lingering phase ambiguities not corrected by the a priori DEM. This step uses the assumption that the along-track phase should follow a consistent pattern over most part of the satellite ground track. Hence, any large discrepancies from the overall pattern of the along-track phase would indicate an ambiguity. The ambiguity is detected by computing the residuals of the along-track phase by removing a smoothed version. If any of the residuals have a magnitude larger than π it is considered ambiguous and thus corrected by adding or subtracting 2π.

3 - Surface elevation changes from CryoSat-2

3.1 – Surface fit method

The surface-fitting method is based on fitting a linear model to the elevations as a function of time and space inside a search radius of 1 km (e.g., Howat et al., 2008; Moholdt et al., 2010; Sørensen et al., 2011; Wouters et al., 2015). The linear model consists of a time-invariant

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(static) bi-quadratic surface model to account for variable topography inside the search radius and time-variant part used to extract the temporal change in elevation. The model consists of a total of 7-parameter whereof six of the parameters (0-5) describe the bi-quadratic surface modeling function.

$$H(x, y, t) = H_0 + a_1 x + a_2 y + a_3 xy + a_4 x^2 + a_5 y^2 + a_6 dt \quad (3)$$

The algorithm estimates the elevation change at every echolocation (or grid-node if desired) in the data set. In each solution the signal amplitude and phase are also estimated by fitting a seasonal signal model to the surface-fit elevation residuals, according to:

$$\Delta H = a_0 \cos(wt) + a_1 \sin(wt) \quad (4)$$

where *ΔH* is the elevation residuals estimated from the plan-fit model, $a_{0,1}$ are the model coefficients and t the time. The amplitude *A* is then defined as $A = \sqrt{a_0^2 + a_1^2}$ and the phase *P* as $P = \tan^{-1}\left(\frac{a_1}{a_0}\right)$.

To remove outliers an iterative 3*σ*-filter is used in the full model solution, i.e. the topography, trend and seasonal signal are removed, using a maximum of 5-iterations. For each iteration residuals (full-model) with an absolute value larger than 10 m are removed, as seasonal changes larger than 10 m are not expected (Moholdt et al., 2010; Wenlu Qi and Braun, 2013). The data inside the 1 km cap is weighted according to their distance from the estimation point according to:

$$W = \frac{1}{\left(1 + \left[\frac{d}{\rho}\right]^2\right)} \quad (5)$$

where *W* is the estimated weight, *d* the distance and *ρ* the correlation or resolution parameter set to 500 m. The weighting allows the solution to better reflect local signal dynamics at the prediction point.

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Local elevation time-series are further computed from the elevation residuals and elevation trend from each solution, according to:

$$h(t) = (t - t_0) \cdot \frac{\partial h}{\partial t} + \varepsilon(t) \quad (6)$$

where t is the time epochs inside the search cap, t_0 is the mean time of t, dh/dt is the elevation change rate and $\varepsilon(t)$ is the elevation residual at each time epoch.

The elevation changes estimated from the surface-fitting method are then culled to remove outliers before spatial gridding. Elevation changes with a regression error larger than 15 m a^{-1} are removed. The resulting surface elevations are binned at 5-km resolution for outlier editing purposes. For each cell the local spatial trend is modeled as a bilinear surface, and removed. The residuals are then edited using an iterative 3σ filter until the RMS converges to 2%.

3.2 – Crossover method

The crossover method is used to derive surface elevation at the intersection point between an ascending and descending satellite ground track separated in time (Brenner et al., 2007; Khvorostovsky, 2012; Zwally et al., 1989). The surface elevations and times are then estimated at the crossover location for each track by linear interpolation of the two closest data points. The crossover height difference is then estimated by taking the height difference between the two tracks according to:

$$dH_{12} = H_2 - H_1 + E \quad (7)$$

were H_1 and H_2 are the surface heights at the crossover location at time epoch t_1 and t_2, respectively, and E is the random measurement error, including orbital, range and retracking errors.

This approach produces crossover height differences with scattered time-epochs ranging from 0-4 years. CryoSat-2 has a 369-day repeat orbit configuration with a 30-day sub-cycle meaning that each crossover location will be revisited every 369 days and surrounding

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area every 30 days. This produces annual and sub-annual crossover difference around each crossover location. This fact is used to produce elevation change rates by incorporating all multi-temporal crossover difference within a neighborhood of 2.5-km around each crossover location. The elevation change is then estimated using the same procedure described for the surface-fit method, except that a bilinear model is used to remove any spatial trends in the topography of the crossover elevations according to:

$$dH = a_0 x + a_1 y + a_2 dt \quad (8)$$

where *dH* is the crossover height difference, *dt* the time difference, a_0 and a_1 the across and along-track slope and a_2 the elevation change rate. This produces elevation changes comparable in time and in spatial coverage with the surface-fit method. The same outlier editing schemes is applied to the crossover elevation change rates as for the surface-fit method.

3.3 – DEM method

The DEM method (Helm et al., 2014; Moholdt et al., 2010; Siegfried et al., 2014) is based on removing the underlying local topography from the monthly surface elevations using a DEM made from the full CryoSat-2 data set (jul-2010 to feb-2015). The DEM surface elevations are estimated at each echolocation using bilinear interpolation and differenced to produce elevation residuals. Elevation differences with an absolute value larger than 50 m are removed from the distribution. The resulting observations are culled for outliers by binning the elevation differences into a 5-km resolution cells in which an iterative 3σ filter is applied. To obtain area-average volume changes, the monthly observations are interpolated onto a 1 km grid using Delaunay triangulation and linear interpolation. The volume change rate is estimated by fitting a linear trend to the monthly volume time series, as a function of time.

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3.4 - Gridding of sparse elevation and elevation change data

The gridding is done in a polar-stereographic projection with a latitude of origin at 70°N, central longitude of 45°W and origin at the North Pole. The projection is referenced against the WGS-84 ellipsoid and the grid-resolution. The observations derived from the surface-fit are gridded at a resolution of 1x1-km, due to the high spatial sampling.

The method of Least Squares Collocation (LSC), described in Herzfeld (1992) is used to grid the observations onto a regular grid. LSC is similar to Kriging and allows for optimal interpolation and merging of data with different accuracies, using their inherent covariance structure. The LSC-algorithm uses the 25 closest data points in 8-quadrants surrounding the prediction point to reduce spatial biasing. The prediction equation consists of two terms where the first term is the actual prediction term and the second term accounts for the non-stationary part of the data, as described by:

$$y = C_{xy}(C_{xx} + N)^{-1}x + \left(1 - \sum\left(C_{xy}(C_{xx} + N)^{-1}\right)\right) m(x) \tag{9}$$

were C_{xy} is the cross-covariance, C_{xx} is the auto-covariance, N the diagonal noise-matrix consisting of the a priori RMS-error and $m(x)$ is the median value of the data inside the search neighborhood.

The covariance of the data inside the local neighborhood is modeled as a function of distance away from the prediction point using a third-order Gauss-Markov model described below.

$$C(r) = C_0\left(1 + \frac{r}{\alpha} - \frac{r^2}{2\alpha^2}\right) e^{\left(-\frac{r}{\alpha}\right)} \tag{10}$$

where r is the separation distance, C_0 the local data variance and α is a scaling factor estimated from the correlation length.

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LSC interpolation provides a RMS-error for each prediction point estimated from the modeled covariance of the data according to:

$$C_y = C_0 - C_{xy}(C_{xx} + N)^{-1}C_{xy}^T \tag{11}$$

where the RMSE of the prediction equals to $\sigma_y = (C_y)^{1/2}$ and where $C_{xy}{}^T$ is the transposed cross-covariance matrix.

The elevation changes estimated from the surface-fit and crossover methods are interpolated to a regular grid using their a priori error estimated from the LSC scheme. To avoid unrealistically small errors, common in the regression errors estimated over flat terrain, a minimum error threshold is applied. Error values smaller than a specific threshold are set to the threshold value. The threshold value is representative of the overall precision of the elevation changes over flat terrain and is set to 0.2 m a^{-1}. The data are then gridded using a 50 km correlation length determined from the comparison of CryoSat-2 elevation to airborne measurements (Section 5).

The LSC algorithm is also used to generate a DEM based on the surface elevations generated from the surface-fitting algorithm. The surface elevations generated from the surface-fit were used as input to the gridding-algorithm. The use of surface elevations from the surface-fit provides several advantages compared to the raw observations as they: provide an almost equal number of observations as the raw data, have been screened for gross outliers, have been low-pass filtered using the 1-km search radius, and are all reference to the same time epoch. Further the RMSE error generated from the surface-fit estimated surface height can be used as an a priori error for the LSC gridding procedure.

The DEM is generated using the same approach as for the surface elevation changes, as described previously in the section. Before the gridding procedure is applied elevations $H < 0$ and $H > 3350$ m are removed from the data set. Further, elevations with a standard error larger than 30 m are also removed. The elevations are binned spatially into a resolution of 1000 m and

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inside each cell the local surface trend is removed by fitting of a planar surface, and an iterative 3σ filter is applied to the residuals to remove outliers.

4 - Surface elevations and elevation changes from ICESat

To assess basin-scale patterns of elevation change we compare elevation changes from CryoSat-2 data to elevation changes derived from Ice, Cloud, and Elevation Satellite (ICESat) data. Here we use release 33 (GLA06) data collected over the 2003-2009 period. The ICESat surface heights were used to generate surface elevation changes and seasonal parameters according to method M3 in Sørensen et al. (2011). The derived elevation changes were corrected for the G-C offset (Borsa et al., 2014). Valid elevation retrievals were selected according to Nilsson et al. (2015b). The ICESat elevation, seasonal amplitude and phase, are then used for comparison with CryoSat-2 and to build continuous time series using the surface fit method described in Section 3.1.

5 -Validation

Elevation and elevation change results were generated for the entire Greenland Ice Sheet using CryoSat-2 data collected between Jan-2011 and Jan-2015 using the methodology presented in (Sections 2-3) (JPL product) and by applying the methods of (Section 3) to ESA's CryoSat-2 L2 elevation products (ESA product). Surface elevations and elevation changes were validated against airborne data sets obtained from NASA's Operation Ice-Bridge Airborne Topographic Mapper (ATM). This mission produces both elevation and elevation changes with reported vertical and temporal accuracy in the cm-level (Krabill et al., 2002).

The derived surface elevations from CryoSat-2 are differenced against ATM surface elevations within 50 m of each ATM locations. One month of CryoSat-2 data consistent in time with the ATM elevations are used for the validation to avoid biases due to temporal sampling and to obtain sufficient sample size. A total of four years of campaign data are used for the

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validation of the surface elevations (2011-2014). The residuals are edited using an iterative 3σ filter to remove outliers. The accuracy and precision is estimated as the mean and standard deviation of the differences, respectively. The residual distribution is further binned according to surface slope estimated from the GIMP DEM (Howat et al., 2014) resampled to 500 m. The sensitivity to surface slope (slope error) is then defined by fitting a 1st order polynomial to the slope and height residuals. The rate estimated from the polynomial provides an indication of the magnitude of the slope-induced error over the entire slope interval.

Surface elevation change rates estimated from three different time-periods (2012-2014, 2011-2013 and 2011-2014) of overlapping ATM observations (Krabill, 2014) are used to validate the surface elevation changes estimated from the CryoSat-2 data. The same validation methodology applied to surface elevations is applied to surface elevation changes, with a few minor modifications. First the search radius is increased to 175 m to make it conform to the ATM elevation change resolution of 250 m, as this search radius encloses the entire ATM grid cell. Secondly the estimated mean and standard deviation are multiplied with the individual time-intervals of the validation data sets to make the errors comparable.

For the surface-fit and crossover methods, near-coincident elevation change rates were compared with ATM rates (e.g., April-2011 to April-2014). This provided three validation data sets for the surface-fit method, due to its high spatial coverage. However, only the 2011-2014-validation data set could be used for the crossover method, due to the lower spatial sampling of the crossovers.

The observational error for the DEM-method is estimated similarly as for the surface-fit method, with some modifications. Here, the CryoSat-2 DEM is used to de-trend the corresponding months of CryoSat-2 data consistent with the months used to derive the ATM-elevation changes (e.g., April 2011 to April-2014). However, the monthly spatial CryoSat-2 coverage does not provide adequate number of comparison points. Thus the months of March and May were included in the analysis (i.e. March-May 2011 to March-May 2014) to increase the number of samples.

The overall accuracy and precision for both the surface elevation and elevations changes are then estimated by taking the weighted mean, using the number of observations as weights, for each data set giving an average error for each measurement mode, as seen in Table-2. The weighted average errors for each mode and method have been summarized in Table-1 and Table-2 for both the ESA's and our solutions, where the values for the individual campaigns can be found in the Supplementary material.

The estimated surface elevation changes from the three independent methods were validated separately using near-coincident ATM data. In general we find the same magnitude of improvement observed in the surface elevation validation analysis. The statistics of the elevation change validation have been summarized in Table-2 for each method independently for the two modes of instrument operation. We find the lowest RMSE errors for the surface-fit method, followed by the crossover method and then the DEM method. This differs from the findings of Moholdt et al. (2010) who found lower intrinsic errors for the crossover method, compared to the surface-fit method when applied to ICESat data. The larger search radius used for our application of the crossover method most likely explains the difference in findings between studies. Further, we find that the surface-fit method provides the largest reduction in RMSE for the JPL product, corresponding to 40% and 55% for the SIN and LRM-mode, respectively.

The correlation length used to derive the number of un-correlated grid-cells, which is used to estimate the standard error, was determined from a semi-variogram analysis of the elevation change residuals from CryoSat-2 minus ATM using the data from the surface-fit method. The comparison was done for each mode separately for all the individual campaigns and multiplied with the their individual time span. The semi-variogram was then computed from all the time-invariant residuals, to maximize the spatial coverage, for each mode. Analysis of the

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semi-variogram showed an approximate correlation length of 25 and 35 km for the SIN and LRM-mode respectively. These correlation lengths are slightly lower than those found by Sørensen et al. (2011) for their analysis of ICESat data. To be conservative we chose a correlation length of 50 km for both modes.

Although the main goal of this study is not to derive or compare different types of DEM's they do play a critical part in removing the long-wavelength topography in order to derive the monthly time-series of volume change from the DEM-method. To gain insight into the overall quality of our CryoSat-2 derived DEM (referred to as JPL) we compare it to three other DEM's derived from other data sets. Firstly, we compare it to a DEM derived from ESA CryoSat-2 L2 data (referred to as ESA) gridded in the same manner as our DEM (Section 3.4). Secondly we compare it to a DEM from Helm et al. (2014), also based on CryoSat-2 data from 2011-2014 (referred to as AWI). Thirdly, we compare to a DEM from Howat et al. (2014) (which was used to derive topographical parameters and corrections for the JPL CryoSat-2 data), based on photogrammetry data from 1999-2002 co-registered to ICESat elevation data from 2003-2009 (referred to as GIMP).

These data sets were then compared to IceBridge ATM elevations, spanning the four different campaigns previously used for validation of the CryoSat-2 elevations. The DEM elevation was estimated at each ATM location, using bilinear interpolation, and the elevation difference computed as (DEM-ATM). No attempt was made to account for differences in DEM and ATM epochs. The estimation of the errors of the DEM was determined in the same way as for the individual CryoSat-2 surface heights. The results of the comparison have been summarized in Table-3, as the weighted average of the different campaigns. The values from each individual campaign can be found in the supplementary material.

Analyzing the overall RMSE we find that the AWI produces the lowest RMSE, followed by JPL, ESA and GIMP, due to AWI's lower standard deviation. However, the best accuracy is obtained by the JPL DEM, which shows the lowest elevation bias of all DEM's. The ESA derived

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DEM shows a slightly better standard deviation than the JPL DEM, which can be explained by higher data density in the marginal areas for the ESA data. The difference in density is due to the SNR rejection criterion applied in our elevation processing. This smoothing can explain the lower standard deviations seen for the AWI product. The GIMP data set showed higher degrees of impulse noise than the other products, explaining the higher observed standard deviation. This impulse noise is attributed to that local elevation change rate, which was not accounted for in the creation of the DEM (Howat et al., 2014). Overall we find that the JPL DEM provides a suitable compromise between resolving of local detail and the minimization of bias. Further, modification to the SNR filtering criteria will likely lead to additional improvements in the DEM.

The impacts of the different SIN processing steps were quantified to determine the importance of the different processing steps in the reduction of RMSE. A simple case study over Barnes ice cap, on Baffin Island in the Canadian Arctic, was used to determine the impact of the processing. Barnes was chosen as it saw a major IceBridge ATM campaign in 2011, providing excellent validation coverage. The result of the case study, summarized in (Table-S1), shows that the filtering of the differential phase has the highest effect on the overall RMSE followed by the ambiguity correction. However, the main improvement is still located in the retracking step of the processing. These steps are important to consider, as they have important implications for the accuracy and precision of the measured surface elevations. This is especially true in high relief areas where small changes in the look angle or phase ambiguity issues can produce large elevation errors from 0-100 m in elevation (Brenner et al., 1983).

6 - Error analysis

To compute volume change errors for the three methods we divide the error budget into two main components (1) the observational error (ε_{obs}) and (if appropriate) the interpolation error (ε_{int}). The error budget is estimated using the root-mean-square error (RMSE) of the difference between the CryoSat-2 and airborne elevation and elevation change differences as described in

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Section 5. The overall error is then estimated using the root-sum-square (RSS) of the two error sources. The RMSE is estimated separately from the two different modes, with the total volume change error being computed as the area-weighted sum of the two components, according to:

$$\varepsilon_{vol} = \left(\frac{A_{lrm}}{A_{tot}} \cdot \varepsilon_{lrm} + \frac{A_{sin}}{A_{tot}} \cdot \varepsilon_{sin}\right) \cdot A_{tot} \quad (12)$$

were A_{tot} is the total area of the ice sheet, A_{lrm} and A_{sin} are the corresponding areas covered by each mode summing to A_{tot}. The ε_{lrm} and ε_{sin} are the standard errors of the LRM and SIN computed from the airborne validation data sets.

The observational elevation change error is estimated from the CryoSat-2 - airborne elevation change differences (Table-2) for the three methods. The RMSE from the LRM/SIN errors are computed using Gaussian error propagation producing an observational elevation change error (ε_{obs}). For the surface-fit and the crossover method the interpolation error is estimated as the RMS of the LSC uncertainty grid. For the DEM-method only the observational error is used. The final elevation change error is then estimated by combing the two error sources using RSS according to:

$$\varepsilon_{dh/dt} = \sqrt{\left(\frac{\sigma_{obs}}{\sqrt{N}}\right)^2 + \left(\frac{\sigma_{int}}{\sqrt{N}}\right)^2} \quad (13)$$

Here, N is the number of uncorrelated grid-cells estimated from empirical semi-variogram analysis of the CryoSat-2 and airborne elevation change differences, and estimated according to:

$$N = \frac{A}{\rho_c^2} \quad (14)$$

were A is the total area of the Greenland Ice sheet (~1.7×10^6 km^2) and the correlation length ρ_c of 50 km.

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7 – Results

The measured surface elevations from the two CryoSat-2 products (JPL vs. ESA) showed large differences in both accuracy and precision of the elevation measurements, as seen in Table-1. The average accuracy and precision for the LRM-mode from the two products showed values of 0.00 ± 0.43 m and -1.06 ± 0.89 m for the JPL and ESA products respectively. This corresponds to an average reduction in RMSE of 68% for the JPL product compared to the ESA LRM L2 data. Further, our product shows a 33% lower residual slope error, indicating a lower sensitivity to the degradation of performance as the surface slope increases.

Surface elevations generated from the SIN-mode showed the same type of improvement as for the LRM-mode. Here, an average accuracy and precision was found to be -0.52 ± 0.58 m and -0.90 ± 1.05 m for the JPL and ESA SIN elevation products respectively. This further corresponds to a reduction in the average RMSE of 27% for the JPL product compared to the ESA product. For the SIN-mode the JPL processing produces a slightly lower (23%) residual slope error, compared to the ESA processor.

Larger improvements can be observed if separating the RMSE into its mean and standard deviation, corresponding to the accuracy and precision of the measurements. Using these definitions the analysis found that there is a 45% and 52% increase in precision for the SIN and LRM mode respectively, compared to the ESA L2 product, and a 42% and 99% improvement in accuracy for the respective modes.

The estimated surface elevation changes generated from the surface-fit method also showed improvement in the estimated accuracy and precision, as seen in Table-2. Here, a overall improvement in RMSE of 55% and 40% in the LRM and SIN mode, respectively, was found when comparing against ESA L2 generated elevation changes from the same method. The average accuracy and precision, compared to ATM generated elevation changes, was found to be 0.11 ± 0.67 m (LRM) and 0.30 ± 0.58 m (SIN) for the JPL derived changes. This

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compared to 0.25 ± 1.51 m (LRM) and 0.34 ± 1.06 m (SIN) for the ESA derived changes. This corresponds to an increase in elevation change accuracy of 56% and 12% for the LRM and SIN-mode, respectively, for the JPL product compared to ESA L2 elevation changes. The estimated elevation changes also show an increase in precisions for the JPL product of 56% and 45% for the LRM and SIN-mode, respectively, compared to its ESA counterpart.

The implementation of the LMG SIN retracking algorithm was found to reduce the noise in the retrieved surface elevations compared to conventional threshold retracking. This is exemplified over the Jakobshavn Isbræ area of the Greenland Ice Sheet (Figure S1), where LMG and a leading edge threshold retracker were compared. Though roughly comparable in accuracy, the LMG has a 32% improvement in precision when, compared against elevations from airborne laser altimetry. The adaptive nature of the algorithm provides improved estimates of surface elevation and gives good trade-off between accuracy and precision.

The 20% threshold retracker implemented in the LRM-mode was also found to provide improved estimates of surface elevation (both in accuracy and precision) compared to the model-based ESA-L2 retracker. Further, it also showed lower sensitivity to the 2012 melt event, due to the lower threshold used on the leading edge of the waveform.

The estimated elevation changes of the Greenland Ice Sheet, excluding the peripheral glaciers, over the period January 2011 to January 2015 show significant differences between products (JPL and ESA) in both spatial patterns and the total magnitude (Figures 2 & 3). The estimated volume change rate from the surface-fit method is -289 ± 16 km^3 a^{-1} for the JPL-product and -224 ± 31 km^3 a^{-1} for the ESA-product with a mean difference of 65 km^3 a^{-1}. The surface-fit and crossover-method produced on the order of ~20 million and ~2.5 million usable elevation changes, respectively, providing high spatial sampling. Due to the constraint put into the JPL processor the ESA L2 data produced slightly more surface-fit observations (~10%), as more surface elevations were accepted.

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512 The ESA product produces a more positive elevation change pattern, which can be
513 attributed to the 2012 melt event that introduced a large positive bias with a magnitude of ~0.5
514 m (Nilsson et al., 2015). Larger differences in the marginal areas for the surface-fit methods are
515 also observed. These are particularly noticeable in eastern Greenland (near 73.5 degrees in
516 latitude Figure 2) where the ESA data shows marginal areas of rapid thinning that are not visible
517 in the JPL solution. The positive signal detected in the interior of the ESA surface-fit-solution
518 can also be found in the ESA DEM-solution, correlating well with the timing of the summer of
519 2012 melt event. These results are in agreement with earlier work demonstrating the sensitivity
520 of the ESA retracker to the changes in the volume/surface scattering ratio (Nilsson et al., 2015).
521 The three volume change methods produce consistent results from JPL derived
522 elevation changes, with a difference of less than 7 $km^3 a^{-1}$. The spread between volume change
523 methods is larger (50 $km^3 a^{-1}$) when using ESA L2 data. The larger discrepancy can mostly
524 related to the sensitivity of the various methods to the melt event. The surface-fit method
525 produces the most negative number (least affected by the melt event and the lowest estimated
526 error) and is therefore taken as the most reliable estimate for both the JPL and ESA solution.
527 Comparing the estimated volume change to other studies using CryoSat-2 we find that
528 the JPL product is less negative than that estimated by Helm et al. (2014): -375 ± 24 $km^3 a^{-1}$.
529 This difference can be attributed to difference in processing methodology and to the different
530 epoch of the data used by Helm et al. (2014) of January 2011 to January 2014. Using the
531 corresponding epoch the JPL data gives a volume change estimate, based on the surface-fit
532 method, of -353 ± 21 $km^3 a^{-1}$, well within the stated uncertainty of Helm et al. (2014).
533 To examine the regional behavior of volume change estimates of the Greenland Ice
534 Sheet, gridded values from the three methods were divided into 8-drainage basins according to
535 Zwally et al. (2012). When analyzing the volume change time-series at the basin scale clear
536 differences can be observed in the annual and inter-annual behaviors (Figure 3). The northern
537 and interior basins (1, 2, 7, 8) all exhibit large differences (Table 4: 0 - 30 $km^3 a^{-1}$) in the

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estimated volume change rates due to changes in the scattering regime resulting from the 2012 melt event. In the majority of the southern basins (4, 5, 6, 7), located in areas with higher precipitation, both products show good agreement in both trends and seasonal amplitude estimated from the DEM-method.

The amplitude of the seasonal signal (Equation 4) estimated from the surface-fit (SF) method show large differences in both magnitude and spatial variability (Figure 5). For the surface-fit method a difference in amplitude of 54% is observed between the ESA and JPL products, corresponding to area-averaged amplitude of 0.17 m or the JPL product and of 0.37 m for ESA product. The comparison with ICESat derived amplitudes from 2003-2009 estimated in (Sasgen et al., 2012) using the same methodology as used here produced an area-averaged amplitude of 0.13 m, which is in good agreement with the JPL derived amplitude. This agreement is also spatially consistent, as seen in (Figure 5), indicating low sensitivity to seasonal changes in scattering regime of the upper snowpack. The observed difference in amplitude bias, taking ICESat as the true surface amplitude while acknowledging differences in epochs, is 0.03 ± 0.13 m for the JPL product and 0.21 ± 0.27 m for the ESA product. The smallest differences are observed at high altitudes above 2000 m a.s.l., where the three data sets show almost constant amplitude of 0.1 m (ICE/JPL) and 0.2 m (ESA), providing a factor of two larger amplitude for the ESA product. Below 2000 m a.s.l., corresponding well to the equilibrium-line-altitude (ELA) of the Greenland Ice Sheet, a rapid increase in amplitude is observed for all products. This is especially true for the ESA product, which increases its magnitude by a factor of two.

Analyzing the amplitude patterns on a regional drainage basin level (Figure 5c) we find good agreement between JPL CryoSat-2 and ICESat amplitude with ESA data producing consistently larger amplitudes. Regionally, the highest amplitudes can be observed in the SE of Greenland in basins (3,4,5) and are consistent with regional precipitation patterns that show high average precipitation in these areas (Bales et al., 2009; Ettema et al., 2009).

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The seasonal phase of the peak in amplitude of the seasonal cycle is shown in (Figures 5b and 5c) and shows generally good agreement between the two products, providing the timing of the maximum of the accumulation signal, before the onset of melt, to the months of June/July for both JPL and ESA CryoSat-2 data sets. The ICESat derived seasonal phase shows a higher dependence on elevation where the maximum of the accumulation signal is found in late May below 2000 m and late July/August above 2000 m in elevation. The ICESat discrepancies from the CryoSat-2 data are found in specific basins. Disagreements between the retrieved phase of the peak amplitude from Cryosat-2 and ICESat data are due to differences in temporal sampling as discussed in more detail in Section 8.1.

We used ICESat and CryoSat-2 derived surface heights to generate time series over three regions in Northeast area of Greenland (Zacahariae Isstrøm, Nioghalvfjerdsbrae (N79) and Storstrømmen) for comparison purposes These areas have in recent time shown large and rapid changes, which has been noted by, e.g., Khan et al. (2014). The selected areas were defined using hydrological basins derived by Lewis and Smith (2009), seen in (Figure 6), and were further divided into smaller areas around the termini to highlight performance for areas of rapid change. The ICESat and CryoSat-2 surface heights were then used to generate annual time-series from 2003-2015 using (Equation 6) in the surface fit method. The estimated 12 year time series show overall comparable elevation change rates over both time periods, especially in the terminus areas, providing confidence that CryoSat-2 can actually monitor changes in these areas.

8 - Discussion

The CryoSat-2 processing methodology presented here is found to produce accurate and precise measurements of ice sheet elevation and elevation change. The main improvements have been introduced in the SIN processor with the inclusion of a new type of land ice retracker (LMG), advanced phase filtering and the inclusion of a phase ambiguity correction scheme. This

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processing approach decreased the RMSE in the surface height retrieval by approximately 27% (45% and 42% improvement in precision and accuracy). This improvement further propagated into the quality of the estimated elevation changes for the SIN-mode, with the same magnitude of improvement (Table-2). The described SIN-processing also generated surface elevations and elevation changes with lower sensitivity to the local surface slope, indicating a higher degree of accuracy in the geo-location and surface range estimation.

The SIN processing methodology further includes a phase filtering and phase ambiguity correction scheme. Visual inspections of a large number of tracks have shown more coherent estimation of the surface locations in our product and further the implementation of the phase-ambiguity correction greatly reduced the number of track offsets. It was also noted that a relatively course DEM (~1 km) could be used to resolve phase ambiguities. The detection and correction of phase ambiguities are relatively straightforward and rely mostly on the relative accuracy of the DEM. The implementation of the phase ambiguity correction is particularly important when monitoring smaller ice caps and outlet glaciers, where frequent and large track offsets can bias the estimation of the underlying topography. The statistical effects of these corrections have been analyzed over the Barnes Ice Cap on Ellesmere Island in the Canadian Arctic and are available in Table S1.

The new LRM processing methodology focused on optimal retrieval of surface elevations over the interior parts of the ice sheet. Here the choice of retracking threshold has proven to be the critical factor to acquire high quality surface elevations and elevation changes. The choice of 20% leading edge threshold level reduced the sensitivity to changes in the scattering regime for low slope, high elevation areas. The functional-based retracking algorithm used in the ESA LRM processor corresponds roughly to a 50% threshold level (Wingham et al., 2006a), which appears to suffer from a higher sensitivity to changes in the scattering properties (volume scattering) of the near-surface firn, as the range is reference higher up (later in time) on the leading edge of the waveform. This effect can be seen in (Figure S1a), and that the

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observed negative elevation bias (Table-1) for ESA-LRM (-1.0 m) fit well with the bias for the 50% LRM threshold value shown in Figure S1a. This makes the algorithm more sensitive to annual and sub-annual changes in snow-packs volume/surface scattering ratio, which can produce spurious changes in elevation due to changes in the near surface dielectric properties. This is clearly shown in patterns of ESA product derived elevation changes (Figure 2b) where a large elevation bias was introduced by the 2012-melt event (Nilsson et al., 2015a). The 20% threshold is less sensitive to these types of changes (Table 1 & 2) and is in agreement with previous work that has demonstrated that the 20% threshold best represents the mean surface inside the footprint when exposed to a combination of surface and volume scattering (Davis, 1997).

Surface elevation changes, derived from multi-temporal radar altimetry observations, are typically corrected for their correlation to changes in the radar waveform shape. This is to reduce the effect of changes in the volume/surface scattering ratio of the ice sheets surface (Davis, 2005; Flament and Rémy, 2012; Wingham et al., 2006b; Zwally et al., 2005). This inherently adds to the complexity of the processing and analysis, introducing new biases and error sources in the estimated parameters. For the processing approach presented here many of these steps can be omitted or reduced, as they are an inherent part of the improved waveform retracking. There have been attempts to remove spurious step-changes in elevation resulting from sudden changes in surface scattering characteristics (caused by the 2012 melt event) apparent in the ESA Baseline-B L2 data through post-processing strategies (Nilsson et al., 2015c), but such approaches spread the bias over a longer period of time making the “jumps” less noticeable in the time series by removing the step-change but introduces longer-timescale bias of equal magnitude as the scattering layer is buried by less reflective snow and low-density firn.

The result of the validation procedure shows a larger slope dependent bias in the ESA data, both in the elevation and elevation changes (Figure 1 and Table 1 & 2). This is especially

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true for the surface elevations, which can be seen in the figures of precision and accuracy (Figure 1a and 1c), where both figures show clear linear slope for the ESA surface heights. In comparison, estimated elevations from JPL-product show relatively stable statistics over the entire slope range above 0.2°. The validation of the estimated surface elevation changes, seen in (Figure 1b) and (Figure 1d), shows the effect of the 2012 melt event on the ESA derived elevation changes below 0.2°. Further, the accuracy of the ESA derived changes show a clear negative trend as function of increased surface slope. The derived precision of the surface elevation change increases dramatically above 0.5°, as more complex topography is measured.

The JPL CryoSat-2 processing methodology produces seasonal amplitudes that are in good agreement with those derived from ICESat data, further indicating the processors abilities to track real and physical changes of the ice sheets surface. The current ESA implementation produces noisier estimates of elevation change, as indicated by the larger standard deviations of the residuals in the ESA solutions for the surface-fit and crossover-method. Figure 5 further shows an amplitude bias in the ESA data compared to the corresponding ICESat reference amplitudes. The bias is constant above the Greenland ELA located around 2000 m in altitude but increases linearly as elevations decrease below this. The linear increase in amplitude seems to be connected to the higher and more variable precipitation in the ablation zone where changes in the variable snow cover produces changes in apparent surface height. This is less prominent for the JPL SIN and LRM retrackers. The estimated seasonal phase in Figure 5c and 5d show that both JPL and ESA CryoSat-2 elevation products can adequately resolve the seasonal maximum of the accumulation signal. Both products provide a timing of the maximum to the month of July over the entire ice sheet, independent of elevation. Assessing the CryoSat-2 derived maximum one does however notice a difference between CryoSat-2 and the reference ICESat dataset. This constitutes roughly a ±1 month difference depending on the elevation and the location. The cause of this difference can be attributed to the temporal sampling of the ICESat mission. During the mission, due to degraded laser lifespan, data was only collected in campaign mode during the spring and winter times corresponding to roughly two months of measurements for each period. When the CryoSat-2 data was resampled to coincide with the ICESat temporal sampling the same elevation and spatial pattern in the phase of the maximum seasonal amplitude was observed as determined from the ICESat data. No corresponding change in amplitude was observed. To mimic the temporal sampling of ICESat the each year of the CryoSat-2 data was resampled using the total number of unique months in the ICESat campaign record. This as the specific months used in the ICESat sampling changes with different campaigns.

The three independent methods used to estimate the volume change of the Greenland Ice Sheet produce consistent volume change estimates. This was especially true for volume changes derived from the JPL elevations, with a discrepancy of less than 7 $km^3 a^{-1}$ between methods. All three methods provided the same estimate of integrated volume change but the use of the surface-fit is recommended as it produces higher spatial sampling compared to i.e. the crossover-method and lower errors. The good agreement between the methods further indicates a strong reliability in the estimated volume change rates of the Greenland Ice Sheet over the four-year period. It also shows the ability of CryoSat-2 to capture both small and large-scale spatial patterns in the rugged topography along the coastline and in the interior of Greenland. This is especially true in the major outlet glacier systems (e.g., Zacahariae Isstrøm, Nioghalvfjerdsbrae (N79) and Storstrømmen).

Studying the northern parts of the Greenland Ice Sheet we find that CryoSat-2 captures both intricate and complex behavior in the marginal areas of the ice sheet. This is exemplified in the NE regions of Greenland (Figure 6) near Zacahariæ Isstrøm, Nioghalvfjerdsbrae and Storstrømmen, which all show complex and localized patterns of elevation change. Here, Nioghalvfjerdsbrae shows very small changes in elevation during the observational time-span, while Zacahariae Isstrøm, its major neighbor shows large negative trends in elevation change. The observed behavior agrees with the observations made by in recent studies by Khan et al.

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(2014) and Mouginot et al. (2015) who document rapid retreat and drawdown of the ice-front position of the two systems beginning in 2012. Storstrømmen outlet glacier system also appears to show signs of rapid thinning at low elevations near the ice-front position while a large positive signal is observed roughly 100 km upstream of the terminus. Rates of elevation change from ICESat and CryoSat-2 data show good agreement in basin-scale trends (Figure 6b,c).

The observed volume change rates estimated from this study are within the range of previous studies, ranging from -186 to -309 $km^3\ a^{-1}$ for the time period 2003-2009, summarized by Csatho et al. (2014). A more recent study by Helm et al. (2014: -375 ± 24 $km^3\ a^{-1}$) agrees within uncertainties when differences in observation periods (2011 – 2014 vs. 2011 - 2015) are taken into accounted. Assuming no changes in firn air content over respective study periods and an ice density of 917 kg m^{-3} we compare estimated changes with corresponding estimates of mass change our estimated rate of Greenland glacier volume change. An assessment of changes in firn air content is out of the scope of this paper. Velicogna et al. (2014) estimated mass loss using the Gravity Recovery and Climate Experiment satellites (GRACE) over the time-period 2003-2013 provided (converted from mass) a rate of -305 ± 63 $km^3\ a^{-1}$ for the Greenland which is inclusive of changes in Ice Sheet and peripheral glacier ice mass (–41 ± 8 $km^3\ a^{-1}$, Gardner et al., 2013). The estimated volume change of -265 $km^3\ a^{-1}$ from Csatho et al. (2014) and the estimated rate of -305 $km^3\ a^{-1}$ from Velicogna et al. (2014) spans our estimated rate of -289 $km^3\ a^{-1}$.

9 – Summary and Conclusion

We conclude that the use of an adaptive retracker for the SIN-mode, based on the maximum gradient method, and the use of 20% threshold retracker for the LRM-mode provide improved performance to the retracker currently used for the ESA L2 elevation products. It is further important, especially for the SIN-mode, to apply a leading edge discriminator to identify and track the leading edge of the waveform. The functional model currently employed in the ESA

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processor has, to the author's knowledge, no such discriminator currently implemented. This is important in the SIN-mode, as it often contains multiple surface returns. The single-return model applied in the ESA processor will here have issues fitting a waveform containing multiple surface returns resulting in retrack jitter (Helm et al., 2014).

Using the new CryoSat-2 processing methodology for the LRM and SIN-mode we determine the volume change of the Greenland Ice Sheet to be -289 ± 16 $km^3 a^{-1}$ during the period January 2011 to January 2015. The validation against airborne ATM surface elevations and elevation changes showed an average improvement in the RMSE of the measured elevations of 68% and 27% for the LRM and SIN mode respectively compared to ESA Baseline-B L2 products. The new methodology also provide improved elevation changes with an reduction in RMSE of 55% and 40% for the LRM and SIN mode respectively, compared to their ESA L2 derived counterparts.

The methodology also showed less sensitivity to changes in near-surface scattering properties than equivalent ESA products. The new processing methodology showed little effect of slope-induced errors, providing better performance in the marginal areas of the ice sheets. These improvements to the CryoSat-2 processing mitigate the need for post-processing to correct correlations between changes in surface elevation and changes in the waveform shape (i.e. backscatter and leading edge width etc.) that can introduce biases and add to the complexity of the processing and analysis.

The presented CryoSat-2 processing methodology provides a lower intrinsic error in the measured elevation, elevation change and volume change estimates, all of which will facilitate improved understanding of the geophysical process leading to changes in land ice elevation. Given the pending release of the ESA Baseline-C, which provides improved corrections and processing mainly for the L1B product, further improvements are expected in the near future. The complete set of grids used in this study are available for the public from the main author (J.Nilsson) upon request and are provided in geotiff format.

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Tables:

Table 1: Accuracy (Mean), precision (SD), the total RMS-error (RMSE) and the residual slope error (SE) of surface elevation from CryoSat-2 observations compared to IceBridge ATM elevations. Here, the LRM mode represents the interior of the ice sheet and SIN the marginal high relief areas.

JPL	**Mean (m)**	**SD (m)**	**RMSE (m)**	**SE (m/deg)**
LRM	0.00	0.43	0.45	1.05
SIN	-0.52	0.58	0.82	0.52
ESA	**Mean (m)**	**SD (m)**	**RMSE (m)**	**SE (m/deg)**
LRM	-1.06	0.89	1.40	1.57
SIN	-0.90	1.05	1.13	0.68

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Table 2: Accuracy (Mean), precision (SD), the total RMS-error (RMSE) and the residual slope error (SE) of surface elevation changes from CryoSat-2 derived from three independent methods [Surface Fit (SF), Crossover (XO) and the DEM method (DM)], compared to IceBridge ATM data.

JPL - LRM	**Mean (m)**	**SD (m)**	**RMSE (m)**	**SE (m/deg)**
SF	0.11	0.67	0.70	0.39
XO	0.24	0.72	0.78	1.23
DM	0.21	1.92	1.95	0.68
ESA - LRM	**Mean (m)**	**SD (m)**	**RMSE (m)**	**SE (m/deg)**
SF	0.25	1.51	1.57	1.40
XO	0.60	1.02	1.20	5.01
DM	0.27	2.40	2.56	3.37
JPL - SIN	**Mean (m)**	**SD (m)**	**RMSE (m)**	**SE (m/deg)**
SF	0.30	0.58	0.66	0.52
XO	-0.60	1.26	1.26	0.60
DM	0.26	3.59	3.72	4.97
ESA - SIN	**Mean (m)**	**SD (m)**	**RMSE (m)**	**SE (m/deg)**
SF	0.34	1.06	1.11	0.39
XO	-0.21	1.44	1.44	0.81
DM	0.31	5.01	5.03	3.07

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Table 3: Validation of four different DEMs, compared to IceBridge ATM elevation data. Based on the weighted (number of samples) average of the four different ATM campaigns from 2011 to 2014. Elevation values at each ATM location were estimated by bilinear interpolation for each DEM product.

DEM	**Mean (m)**	**SD (m)**	**RMSE (m)**	**SE (m/deg)**
AWI	-1.35	5.95	6.12	1.35
GIMP	-1.13	7.22	7.32	0.84
JPL	-0.87	6.31	6.39	1.06
ESA	-2.83	6.13	6.76	1.62

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Table 4: Individual basin volume changes (km^{3}a^{-1}) for the Surface-Fit (SF), Crossover (XO) and DEM (DM) method for the JPL and ESA product for the time period Jan-2011 to Jan-2015, with corresponding volumetric error.

Basin	DM - JPL	SF – JPL	XO – JPL	DM - ESA	SF - ESA	XO - ESA
1	-21 ± 26	-26 ± 6	-23 ± 9	-8 ± 34	-9 ± 12	-11 ± 12
2	0 ± 26	5 ± 7	0 ± 10	25 ± 35	31 ± 14	30 ±13
3	-48 ± 27	-38 ± 6	-34 ± 15	-26 ± 36	-46 ± 13	-31 ± 20
4	-39 ± 20	-36 ± 4	-37 ± 11	-46 ± 27	-42 ± 9	-16 ± 14
5	-20 ± 11	-19 ± 2	-27 ± 8	-13 ± 15	-19 ± 4	-6 ± 9
6	-75 ± 22	-72 ± 6	-71 ± 10	-79± 29	-75 ± 11	-79 ± 18
7	-46 ± 20	-56 ± 6	-51 ± 10	-36 ± 27	-41 ± 12	-35 ± 14
8	-45 ± 25	-48 ± 6	-45 ± 10	-21 ± 33	-23 ± 13	-27 ± 14
TOT	**-295 ± 64**	**-289 ± 16**	**-288 ± 30**	**-203 ± 85**	**-224± 31**	**-174 ± 41**

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Figures:

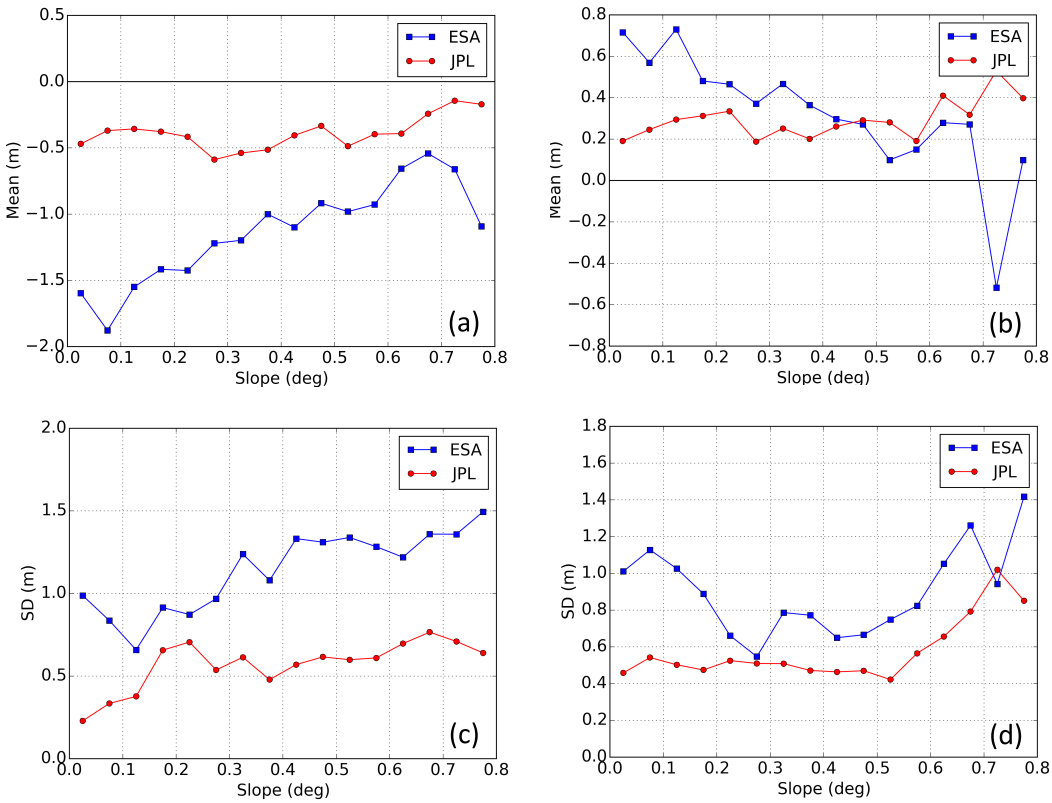

Figure 1: Validation of surface elevations (2012) (a,c) and surface elevation changes (2011-2014) (b,d) compared to IceBridge ATM, as a function of surface slope. The accuracy of the measurement is defined as the mean-value (Mean) of the CryoSat-2-ATM residuals and the precision as the standard deviation (SD).

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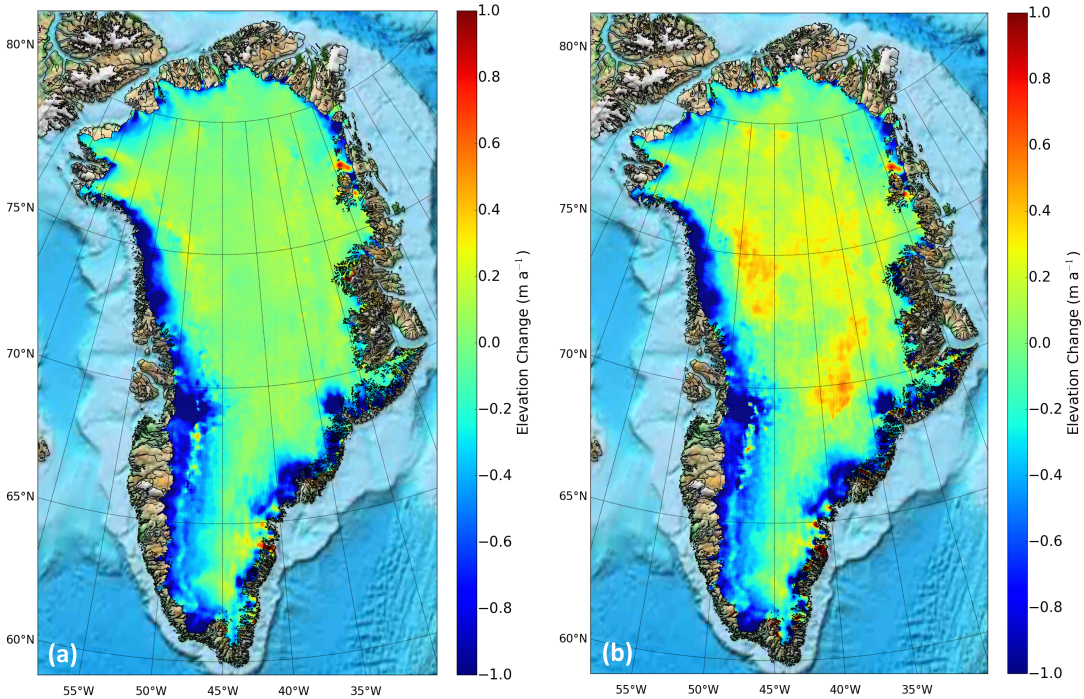
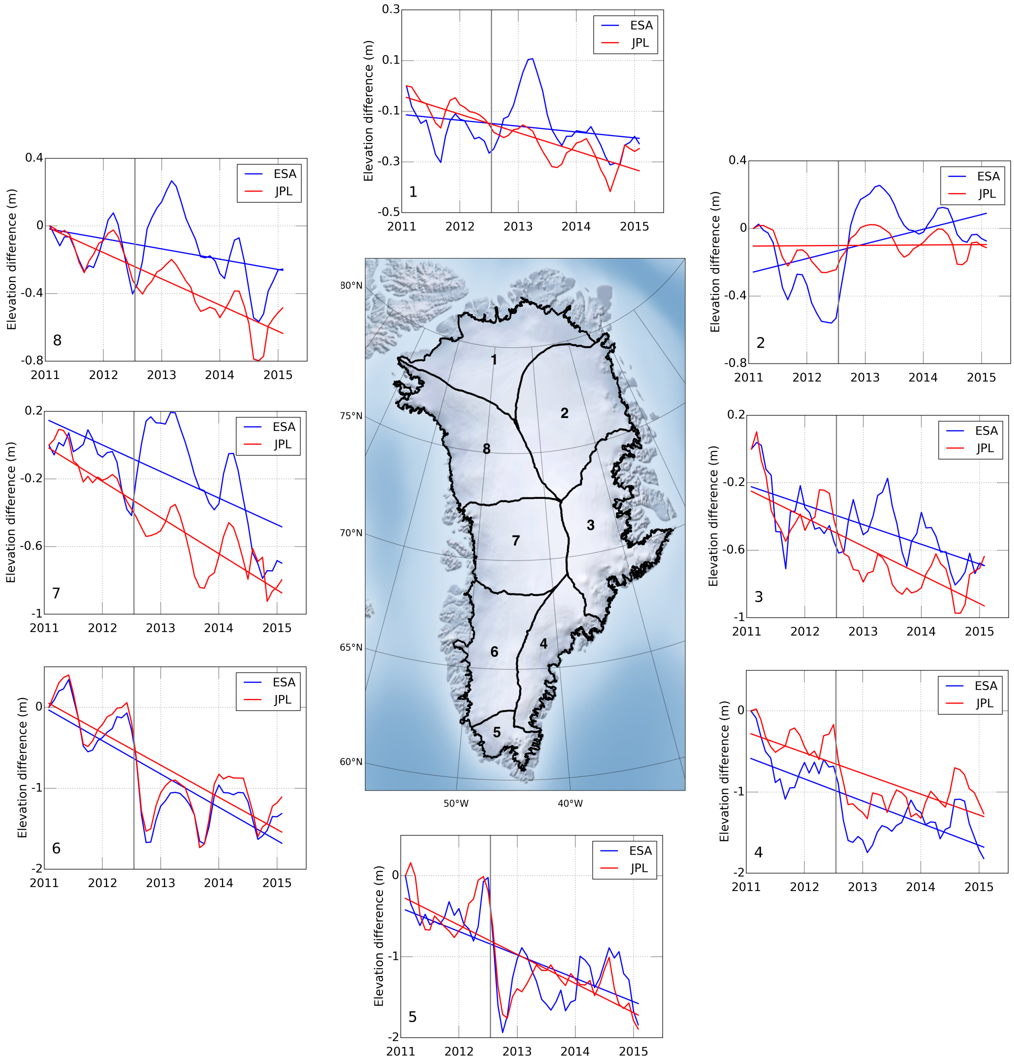


Figure 2: 2011-2015 elevation changes estimated from the surface-fit methods for the JPL (a) and ESA L2 products (b). The JPL product produced a total volume change of -289 ± 16 km^3 a^{-1} while the estimated total volume change of the ESA product totaled -224 ± 31 km^3 a^{-1}. This corresponds to -29 versus 38 km^3 a^{-1} (H > 2000 m) and -259 versus -262 km^3 a^{-1} (H < 2000 m) for the JPL and ESA product respectively. Images have been smoothed with a 5 km median filter for visualization purposes.



955

956 *Figure 3. Monthly elevation change time-series for 8 large drainage basins of the Greenland Ice*

957 *Sheet. Time-series have been smoothed using a 3-month moving average for improved*

958 *visualization. The grey vertical line indicates the timing of the 2012 melt event.*

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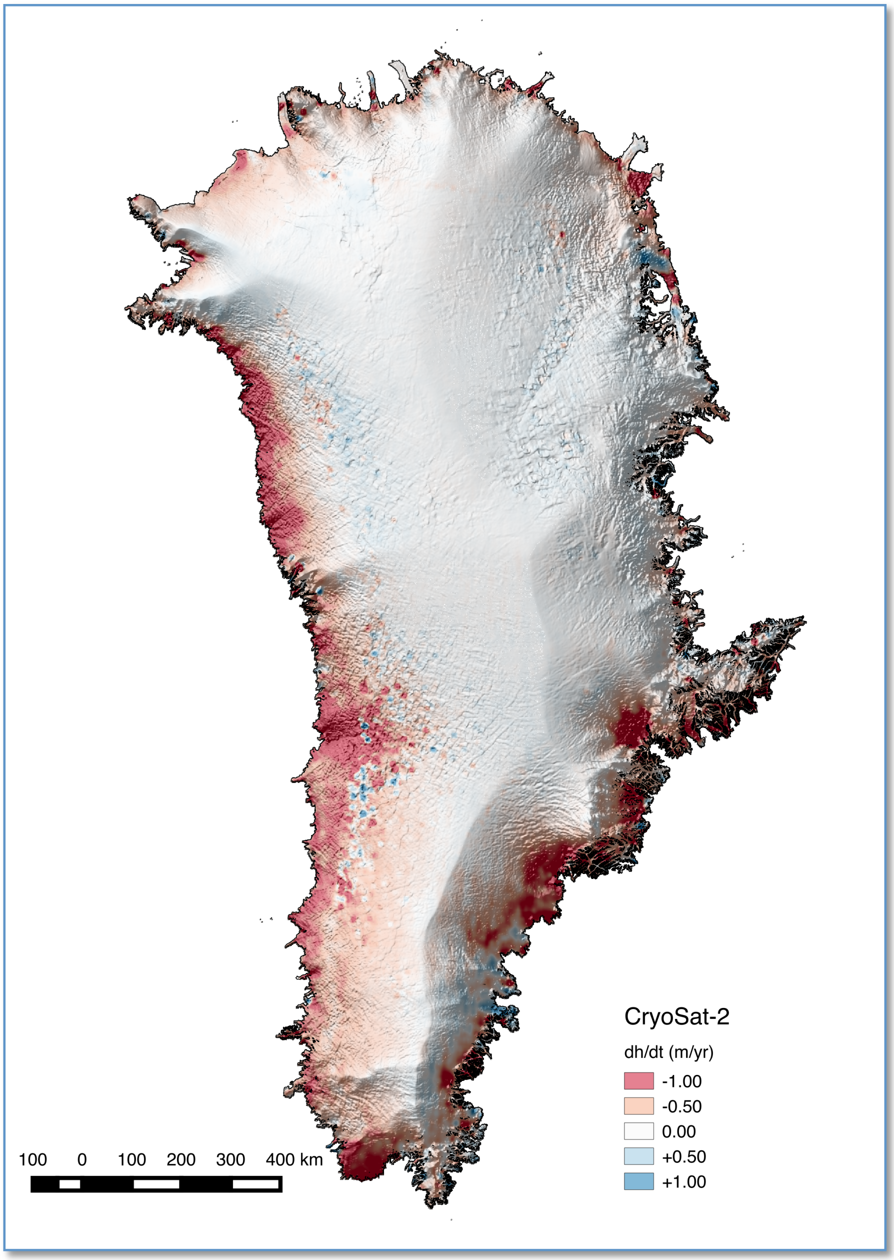

959

960 *Figure 4. Surface elevation changes of the Greenland Ice Sheet from 2011-2015, based on the*

961 *surface-fit method, overlaid on CryoSat-2 hill-shaded DEM (Jul-2010 to Feb-2015).*

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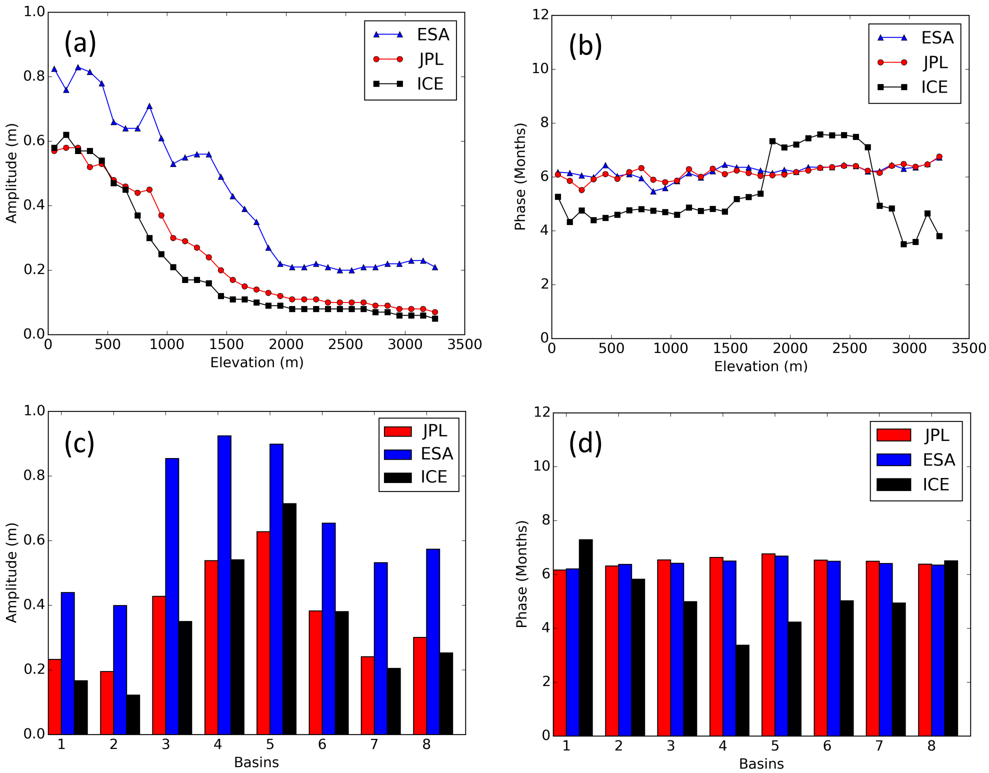

Figure 5: Estimated seasonal amplitude (a,c) and phase of the maximum amplitude (b,d) from the surface-fit method for CryoSat-2 [ESA (blue) and JPL (red)] compared to ICESat (ICE, black)). Values are compared using a search radius of 500 m and the phase offset is referenced from 1st of January

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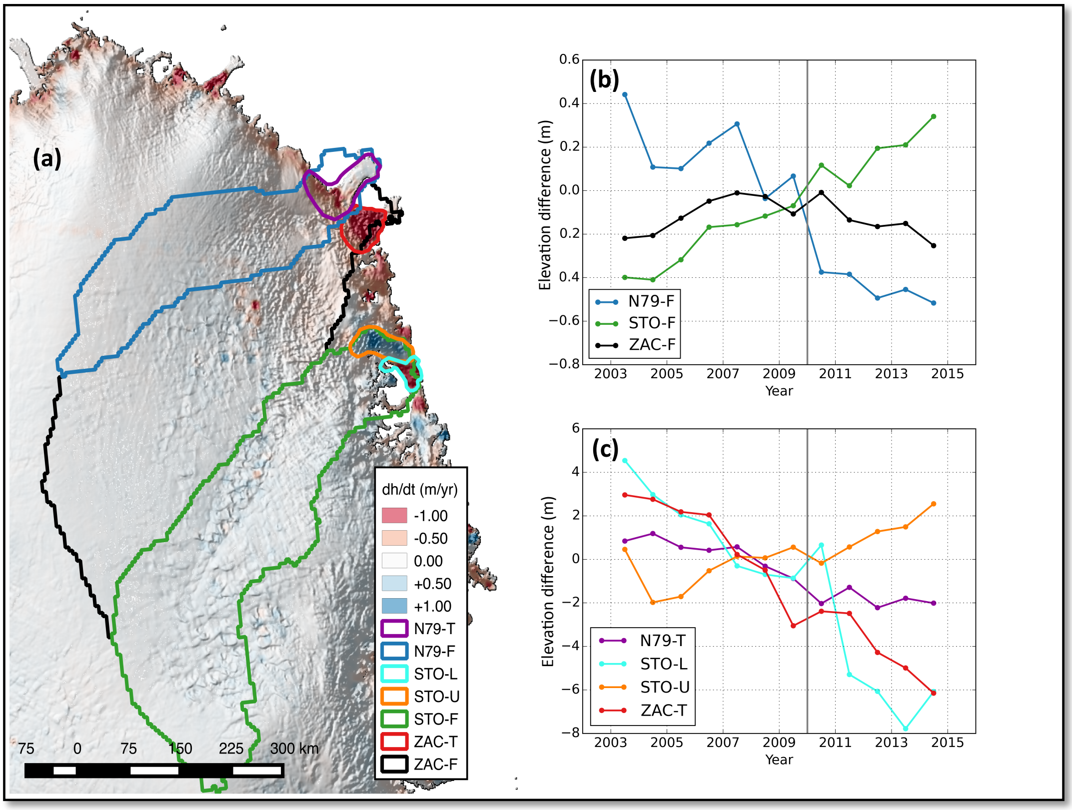

Figure 6: Northeast part of the Greenland Ice Sheet showing surface elevation change (a) from CryoSat-2 JPL-solution (2011-2015), with corresponding hydrological basin outlines. The hydrological basins are separated into full basins size (b) and to the terminus areas (c). Sub-figures (b) and (c) shows a merged 12 year annual elevation time series from ICESat and CryoSat-2 for each color-coded area in (a). The derived elevation time series was formed using the surface-fit method described in Section (3.1). The elevation change map is overlaid onto the CryoSat-2 hill shaded DEM based on surface heights from Jul-2010 to Feb-2015.