ESA Climate Change Initiative (CCI)

Greenland Ice Sheet (GIS) Essential Climate Variable (ECV)

Algorithm Theoretical Baseline Document (ATBD)

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## Signatures page

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## Acronyms

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<td>ASTER</td>
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1 Introduction

1.1 Purpose and Scope

This document is updated for Phase 2 for the "Greenland_Ice_Sheet_cci" (GIS_cci) project in accordance to Contract [AD1] and Statement of Work [AD2].

This Algorithm Theoretical Basis Document (ATBD) is hence based on the Phase 1 ATBD document [RD2] of the "Ice_Sheets_cci" project. The ATBD document is part of Task 2 Algorithm Development deliverables, with deliverable id; D2.1.

The ATBD provides an overview of the algorithms which might be used to generate the ECV parameters 'Surface Elevation Change (SEC)', 'Ice Velocity (IV)', 'Calving Front Location (CFL)', 'Grounding Line Location (GLL)' and 'Gravimetric Mass Balance' in the formats required by the users (see URD and PSD). The ATBD provides a description of the scientific background to an algorithm and a functional description of what the algorithm does' (SoW). For each of the products it reviews the scientific background, the principle of the algorithms, their expected or known accuracy and performance, input and output data, as well as capabilities and limitations.

1.2 Document Structure

This document is structured into an introductory chapter followed by four chapters describing the algorithms for retrieving the ECV parameters of the Greenland_Ice_Sheet_cci project, namely

- Surface elevation change, from radar altimetry, (SEC), section 2
- Ice velocity (IV), section 3
- Calving front location (CFL), section 4
- Grounding line location (GLL), section 5
- Gravimetric Mass Balance (GMB), section 6

Each chapter provides a summary of available algorithms, together with a short description of the respective approach. The capabilities and limitations of the various algorithms for retrieval of the relevant parameter are discussed.

1.3 Applicable and Reference Documents

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**Note:** If not provided, the reference applies to the latest released Issue/Revision/Version
2 Surface elevation change

2.1 Introduction

Satellite altimetry provides estimates of ice sheet elevation changes through repeated measurements of ice sheet surface elevations. The technique has been employed to study both Greenland (Johanessen et al., 2005; Zwally et al., 2005; Zwally et al., 2011; Sørensen et al., 2011; Khvorostovsky 2012) and Antarctica (Wingham et al., 1998; Davis et al., 2005; Zwally et al., 2005), and has the distinct advantage of being able to resolve the detailed pattern of mass imbalance, with frequent (up to monthly) temporal sampling. Radar altimetry, in particular, provides the longest continuous observational record of all geodetic techniques (Wingham et al., 2009).

Altimeters using microwave frequencies are commonly referred to as radar altimetry. At these wavelengths the signal can penetrate cloud cover, making the measurements possible in all weather conditions. In addition, the use of microwaves enables measurements to be made independently from sunlight conditions. The satellites with altimeters on board are placed in repeat orbits (covering a region of up to 1 km on either side of a nominal ground track) enabling systematic monitoring of the Earth. Furthermore, satellite altimetry radars have been in continuous operation since 1991 and new missions have been launched during last few years. There is therefore the availability of long time series and as a consequence the possibility to monitor seasonal to inter-annual variations during the lifetime of these satellites.

The specific objectives of this chapter are:
- to provide the theoretical basis of the algorithms that will be used to generate elevation changes maps from radar altimeter data;
- to assess the accuracy of these products; and
- to evaluate the range of applicability and the limitations of the derived data.

2.2 Review of scientific background

Radar altimeters provide a measure of the time, t_d, of a radio signal to travel from the emitting instrument, reach a target surface, and return/scatter back. The distance from the reflecting target to the radar is given by (Elachi, 1988):

\[ r = \frac{ct_d}{2} \]  

(2.1)

where c is the speed of light. The accuracy with which the distance is measured is given by

\[ \Delta r = \frac{c}{2B} \]  

(2.2)

where B specifies the signal bandwidth. The operating principle of an altimeter is shown in Figure 2.1 (a). Surface elevation h is calculated as the difference between the satellite altitude, a, and the measured range, r:

\[ h = a - r \]  

(2.3)

h is relative to the reference ellipsoid used for determining satellite altitude (see Figure 2.1 a). In addition to measuring range, the altimeter records a sample of the pulse echo return and estimates other parameters, including the magnitude of the return.

The side view representation in Figure 2.1 (b) shows the propagation of a single pulse along the beam of the antenna towards a horizontal and planar surface. The curved lines represent the pulse propagating and the temporal width between the curves is constant and equal to \( \phi \), the duration of the pulse length. A different visualization of the propagation (looking down on the scattering surface from the instrument position) is also provided in Figure 2.1 b) (plane view). When the spherical wavefront first hits the surface at the instant time t_0, the footprint is a point. The area illuminated by the pulse increases to a circular area until the trailing edge of the wavefront reaches the surface, at the instant time t_1. The pulse-limited footprint is the maximum circular area defined as the radius of the leading edge of the pulse when the trailing edge of the pulse first hits the surface. As the pulse propagates, the circle transforms into rings of
equal area (Fu and Cazenave, 2001). The figure shows also a typical return waveform. The power received begins to increase from the time when the wavefront hits the surface, \( t_0 \), and continues to increase for the duration of the pulse. The waveform presents a linear leading edge corresponding to this initial interaction. At the times greater than the pulse duration, the area intercepted by the pulse remains constant with time. But instead of remaining constant, the power of the reflected pulse actually decreases gradually with time according to the illumination pattern of the antenna. The mid-point of the leading edge corresponds to the range to the mean surface within the pulse-limited footprint.

Information about surface roughness can be obtained from waveform analysis. When a pulse scatters from a surface, the returned echo has a shape reflecting the (statistical) properties of the surface. In the case of the ocean, where the surface is homogeneous, the height statistics are the main factors in determining the pulse shape. In the case of terrain, the surface composition varies across the antenna footprint and its statistical properties need to be taken into account. For a perfectly smooth surface, the echo is a mirror
image of the incident pulse. If the surface has some roughness, some return occurs in the backscatter direction at slight off-vertical angles as the pulse footprint spreads on the surface. This results in a slight spread in time of the echo. If the surface is very rough, some of the energy is scattered when the radio pulse intercepts the peaks of the surface and more energy is scattered as the pulse intercepts areas at various heights of the surface. This leads to a larger multi-path spread of energy which results in noticeable rise in the echo leading edge. The rise is used to measure the surface roughness.

The propagation of the pulse with time, as described above, assumes the forming of the returns by scattering from the surface only. However, it has been shown that ice sheet returns consist of a combination of surface and sub-surface volume scattering due to penetration of part of the radar signal through the snow surface (Ridley & Partington, 1988). Volume scattering mainly results from the presence of inhomogeneities in the host medium, like ice grains, air bubbles, and ice inclusions, whose size, shape, density, dielectric constant, and orientation affect the scattering. They cause a redistribution of the energy of the transmitted wave into other directions and results in a loss in the transmitted wave (Ulaby et al., 1982). Signal penetration is largest in the dry snow zone of the ice sheets and can exceed 5 m (Davis and Poznyak, 1993; Legresy and Remy, 1998).

Over ice sheet surfaces, the on-board tracker is generally unable to keep the leading edge of the waveform centred on the tracking point of the waveform window, and waveform retracking is to be applied to determine this offset. Several methods were developed for retracking ice sheet radar altimeter data (e.g. Bamber, 1994; Davis, 1997; Zwally & Brenner, 2001; Legresy et al., 2005). Retracking algorithms are based on defining the point where the waveform exceeds a certain percentage of the maximum power (threshold retrackers) or on functional fits to model waveform shape. All retrackers have their advantages and disadvantages, and selection of the retracker will affect taking of topography and volume scattering into account. Functional-fit retrackers more accurately produce individual elevation estimates, while threshold retrackers could be preferred for elevation change studies because they give more repeatable elevations.

### 2.3 Elevation Biases

One thing which needs to be taken into account when generating time-series of elevation change from different satellite sensors is that they cannot be directly compared since a relative-elevation value is a relative measure and the satellite sensor itself introduces an error into the measurement. Since it is the goal of the is_cci to compare relative-elevation time-series from two or more satellite sensors, it is necessary to identify the relative relationship between the systematic errors of the satellites so that there is no offset between the different time-series when plotting elevation (change) versus time. It is assumed that the systematic error in the satellite measurement is constant in time but may be spatially dependent.

There are several approaches that can be applied to perform cross-calibration. One approach is based on the comparison of all available cycles (or any selected time intervals) of elevation time-series for overlap period between successive mission phases. Then the bias is estimated as the difference between relative-elevation $dh$ data from both time-series (e.g. Davis et al., 2005; Khorostovsky, 2012).

In order to maximize the number of cycles (intervals) involved in estimation of the bias and in construction of the merged time series for overlap period between successive mission phases the following steps are performed:

1. The matrixes of the differences $\Delta dh_{ij} = dh_i - dh_j$ is formed from both time-series $(i = 1, ..., N; j = 1, ..., N; N$ – number of time intervals within overlap period)

2. A merged matrix of the differences $\Delta dh$ is created: if matrix element $\Delta dh_{ij}$ is available for both time-series their mean is used; if matrix element $\Delta dh_{ij}$ is available only for one time-series a value from corresponding matrix is used; if matrix element $\Delta dh_{ij}$ is not available for both time-series it is set to NaN.

3. A merged time-series for overlap period is formed by averaging matrix elements in each column $j$ of the merged $\Delta dh$ matrix. Elements from $i$-th row of the matrix is involved in averaging only if $i$-th element is available for both $dh$ time series. This ensures that data in the resulting time series for overlap period are referenced to the same time interval.

4. The biases between the merged time-series for overlap period and two $dh$ time series from successive missions are estimated as the mean differences between overlapping data. Then the bias between the $dh$ time series of two successive missions is equal to the difference between these two biases.
This approach adjusts time series of relative-elevations \( dh \) together and creates continuous time series avoiding the necessity to determine and apply absolute inter-satellite biases. However, it ignores a large amount of inter-satellite measurement pairs, which can be used in estimating of elevation changes especially when applying crossover method. Therefore, second approach is based on estimating inter-satellite biases using the comparison of inter-satellite measurement differences (Zwally et al., 2005; Johannessen et al., 2005; Khvorostovsky 2012). Inter-satellite biases can be determined by averaging crossover differences with small time intervals between measurements from different satellites, i.e. using crossovers available from overlap periods between successive mission. However, the number of these crossovers within individual grid cells is not sufficient for calculating statistically significant estimates of the bias over large parts of the southern and margin areas of the Greenland ice sheet.

Therefore two methods can be applied to involve more measurements in the analysis:

- Use crossovers over large areas centred in the grid cell (Zwally et al., 2005). In this case the bias is equal to the mean of inter-satellite measurement differences.
- Use of regression fit of inter-satellite elevation crossover differences \( (dH) \) to corresponding time intervals \( (dt) \) including crossovers with larger \( dt \) (Johannessen et al., 2005; Khvorostovsky 2012); In this case the bias is calculated as an offset of the fit from the origin at the point where \( dt = 0 \).

However, applicability of the second approach is limited by comparatively low spatial resolution of tens of kilometres, which is not sufficient to meet user requirements for the SEC product. Thus, it can be applied only as an optional method for inter-comparison or sensitivity studies.

### 2.4 Algorithms

Elevation change rate is determined from altimeter measurements using crossover or along track analysis. A crossover is the elevation difference at the intersection of two ground-tracks (one ascending and one descending track). The elevation difference is assumed to be an actual surface elevation change at the crossover location (when any potential biases between ascending and descending tracks have been considered), since it occurs at the same location but different times.

In along track analysis, elevation measurements from an altimetry satellite with near-repeat ground-tracks are used to estimate surface characteristics, such as elevation change rate. As the ground-tracks will not repeat exactly, the across-track change in topography should be accounted for in repeat track analysis, so that this is not interpreted as a surface elevation change.

Crossover and along track analyses are primarily used for the retrieval of elevation change or elevation change rates. However, in addition to information on elevation change, an elevation derived from a radar altimeter contains effects from physical processes and measurement errors. Both methods use different techniques to remove error terms present within the derived elevations.

#### 2.4.1 The crossover method

Crossovers are used to determine elevation change at the intersection of two altimeter passes. In this section, a detailed description of crossover theory will be given.

##### 2.4.1.1 Principle of method

Surface elevation differences \( (dH) \) are computed from surface heights \( (H) \) at the intersection or “crossover” between two satellite ground-tracks or spot track, one ascending and one descending (Figure 2.2). The measured elevation difference at a crossover point is:

\[
dH(t) = H_2 - H_1 + E
\]  

(2.4)

where \( H_2 \) and \( H_1 \) are the surface elevations at times \( t_2 \) and \( t_1 \), respectively, and \( E \) is the random measurement error, which includes errors in the altimeter-range measurement and in the determination of the vertical position of the orbit (Zwally et al., 1989). In the case of radar altimeters, the range measured is the average surface elevation in the pulse-limited footprint, whereas for a laser altimeter the elevation is obtained by fitting Gaussian functions to the returned waveforms. The maximum amplitude of the return marks the two-way travel time, which translates into distance from the satellite. As satellite altimeters make discrete, non-continuous measurements, it will be rare to have measurements exactly at the crossover location, which means that surface elevations at the crossover points are obtained by interpolation.

There are two methods to obtain the rate of change of surface elevation from a set of crossover measurements. The first method, known as the time series method, consists in comparison of
measurements related to two distinct periods separated by a relatively large time interval ($\Delta t$). It is used when a sufficiently large number of measurements is available. The surface changing rate is the average crossover height difference divided by the time interval:

$$\frac{\sum (H_2 - H_1)}{N} / \Delta t$$

(2.5)

where $(H_2 - H_1)_i$ is the elevation difference at the $i$-th crossover and $N$ is the number of crossovers.

The second method, known as the $dH/dt$ method, is appropriate for a set of crossovers that tend to have randomly distributed time intervals. The slope of a linear fit to the crossover differences, $dH_i = (H_2 - H_1)_i$ versus the time intervals, $dt_i = (t_2 - t_1)_i$, gives the thickening ($dH/dt > 0$) or thinning rate ($dH/dt < 0$).

The second method gives a more valuable result of elevation change for the entire time-interval considered, while the first one allows investigating its temporal variability. The time series method proposed by Wingham et al. takes into account the ascending/descending bias and is known as the dual crossover method (Wingham et al., 1998). It was first used to define elevation changes in Antarctica, using ERS-1 measurements. Instead of combining ascending and descending tracks from a single orbital cycle, the elevation measurements from pairs of orbital cycles, acquired at times $t_1$ and $t_2$ are compared.

In this manner two pairs of elevations can be considered: ascending track elevation $h_{At_1}$ (measured during orbit cycle 1 at time $t = t_1$) and descending track elevation $h_{Dt_2}$ (measured during orbit cycle 2 at time $t = t_2$), as well $h_{At_2}$ (ascending track, orbit cycle 2) and $h_{Dt_1}$ (descending track, orbit cycle 1). The change in elevation between two orbital cycles at each point $x$ is defined as

$$\Delta h(x,t_1,t_2) = \frac{1}{2} [(h_{At_1} - h_{Dt_2}) + (h_{At_2} - h_{Dt_1})]$$

(2.6)

The elevation change during one orbit cycle is supposed to be negligible. To see how the surface elevation changes over time, one of the orbital cycles can be chosen as a reference cycle. Pairing all other orbital cycles with the reference cycle, a time series of values of surface elevation $\Delta h(x,t,t_{ref})$ relative to the surface elevation measured during reference orbital cycle can be obtained from:

$$\Delta h(x,t,t_{ref}) = \frac{1}{2} [(h_{At} - h_{Dt_{ref}}) + (h_{At_{ref}} - h_{Dt})]_{n \in \text{X}}$$

(2.7)
Methods for deriving elevation time series from crossover data have evolved steadily over time. Early studies calculated crossovers between an early "reference" period and later measurements; the time-series of these differences gave the evolution of the surface height, which was fit with a linear model. These studies often simultaneously solved for a sinusoidal seasonal cycle, assumed to be constant from year to year (see e.g. Zwally et al. 1983).

More recently, researchers have fit auto-regressive models to the data (Ferguson et al., 2004), a technique which finds smoothly-varying functions that match the data to within tolerances appropriate to the accuracy of the data. These techniques can identify both seasonal and longer-term height variations, and are claimed to be less prone to a bias induced by seasonal cycles than linear-fitting techniques. In the recent studies (e.g. Zwally et al., 2005, Li and Davis, 2006; Khvorostovsky, 2012) all time periods are considered as reference in order to use all available crossovers.

2.4.1.2 Input / Output

Waveform product for ERS-1/2, i.e. level 1b data, and GDR level 2 Envisat data of radar altimeter measurements will be used for producing surface elevation change estimates. In addition level 1 and 2 data processed by GSFC/NASA will be used for inter-comparison of different algorithms for processing Level-1b data (Level-1 in GSFC). As noted above it is assumed that when all necessary corrections are applied the crossover method provides assessment of actual elevation changes as they occur at the same location (Figure 2.3). The output is a gridded elevation changes in NetCDF format (optionally an ASCII table).

![Crossover Processing Diagram]

Figure 2.3: Schematic of the SEC processing line using the crossover method.

2.4.1.3 Accuracy and performance

Although accuracy of crossover differences degrades with the surface slope (Brenner et al., 2007), crossover method implies comparison of elevation over the same location and therefore allows avoiding of strong influence of slope-induced errors on elevation change estimates. It allows using this method to validate elevation change estimates obtained by along-track methods. In addition accuracy and precision of the measurements used in ice sheet studies depends on mode of altimeter operation that is different for different satellites. Envisat fine mode altimeter measurements available over vast central areas of Greenland ice sheets have better accuracy and precision as compared with lower resolution measurements from ERS satellites in ice mode used for ice sheet studies.

Comparison of elevation changes derived from ERS-2 RA data against results derived from airborne and ICESat laser altimetry over Greenland have been shown that radar altimetry gives more rapid thickening
over central parts of ice sheet and underestimate ice thinning over margins by several cm/year (Thomas et al., 2008). For central Greenland it has been associated with the effect of lifting of the radar-reflection horizon as a response to increased surface melting, and problems of penetration of the radar altimeter signal in the snow. Conversely, over the margins the proposed reasons of differing results are limited capabilities of radar altimetry to measure over rough sloping surfaces. However, limited spatial and temporal coverage of interior regions of Greenland by airborne data also may be the cause of observed difference. Adjustment of elevation time series for the errors associated with penetration of part of the radar altimeter signal through the surface are based on taken dependence of elevation on waveform parameters (Wingham et al., 1998; Davis and Fergusson, 2004; Khvorostovsky 2012). Detailed comparison of radar and laser data is planned within the project and will allow validating techniques for correcting the effect of temporally varying signal penetration.

### 2.4.1.4 Capabilities and limitations

For a radar altimeter, the antenna is pointed at nadir and the pulse-limited footprint tends to be located at the closest surface lying within the beam-limited footprint, namely the first part of the reflected echo will come from the part of the surface within the beam-limited footprint that is closest to the satellite. Over flat surfaces, the closest point on the surface is at nadir, whereas over sloping or rough surface it is offset from nadir. Thus, the determination of the absolute surface elevation at nadir requires correction for slope-induced offset of the pulse-limited footprint from nadir. However, for the purpose of studying elevation changes, correction for slope-induced errors is not necessary because the pulse-limited footprint is usually located at the same place on the surface due to the fact that the satellite is at the same position (the intersection of the two orbits). Thus the slope errors cancel, and any difference between the two measurements reflects elevation change. Another advantage of this method as compared with along-track method is that it does not need repeat phase available is able to analyse data in geodetic phase (e.g. ERS-1 operation over period from April 1994 to March 1995).

The main drawback is that the method provides information on elevation change only on crossover points. In addition, the location of the measurements is still determined by the surface slope, the footprint width and the pointing angle, so significant difficulty remains in measuring elevation changes for outlet glaciers, where the crossovers are preferentially located on the surrounding high points. The error in elevation determination can be caused by non-modelled or incorrectly modelled atmospheric errors. Orbit error, or uncertainty in the altimeter position and inaccuracies in the model of Earth’s gravity field also introduce errors into the derived elevations.

Since elevation data are discrete and not continuous, there will rarely be two elevation measurements at a crossover location, which means that interpolation is required to determine elevation values at the crossover location. Interpolation error can contribute to error in elevation determination, as the interpolation process may produce elevations inconsistent with the actual topography; this error will be higher in areas with higher topographic variability.

In addition, as mentioned in section 2.2, further complication arises as microwaves can penetrate up to several metres into the snow pack. In addition, the measurements show difference due to volume echo induced errors between ascending and descending tracks.

### 2.4.1.5 Error characterisations

The main error sources, when using crossover method, are associated with penetration of radar altimeter signal in the snowpack and changes of volume scattering in time (Ridley and Partington, 1988). This error is corrected by applying proper retracking and backscatter corrections.

Availability of elevation differences only for crossover locations limits the coverage density of the SEC estimates. At the same time since the crossover method estimates elevation changes occurring at the same location the slope errors cancel and do not significantly contribute to the SEC error budget.

The SEC error budget is also affected by uncertainties in satellite orbit, atmospheric (tropospheric and ionospheric) and tide (solid earth and ocean load) corrections as well as corrections for Glacial isostatic adjustment representing response of the Earth to past changes in the ice loading.

### 2.4.2 The along-track methods

Along-track analysis is a technique used to detect elevation change rate by comparing elevations from different times along the same repeat portion of a ground-track or spot track from one orbital revolution of the altimeter, also referred to as track segment (Zlotnicki et al., 1989). The track segments will not repeat exactly, so along-track analysis is an attempt to reduce the associated errors by estimating parameters that influence the elevation.
2.4.2.1 Principle of method

In an ideal case, ground-tracks or spot tracks for altimeter satellites in repeat-track orbits (like Envisat 2003-10, and ICESat) would repeat exactly so that elevations along the track at one time could be directly compared to elevations along the same track obtained at a different time. However, differences in the altimeter pointing angle and orbital perturbations will cause across-track differences, which should therefore be compensated for within the repeat track analysis.

The unmeasured topography between near repeat-tracks needs to be considered when comparing elevations from different tracks. Slobbe et al. (2008) used a DEM to correct for the surface slope between the centre points of overlapping footprints. Using only overlapping footprints limits the slope-induced error, but it also limits the amount of data available for comparison. A method, which uses along-track interpolation to restrict the DEM slope correction to the cross-track distance between two repeat-tracks, has been applied in Moholdt et al. (2010a and b). For pairs of repeat-tracks, one profile is projected onto the other profile using the corresponding cross-track elevation differences from an independent DEM (Figure 2.4a)). Elevations are then compared at each DEM-projected point by linear interpolation between the two closest footprints in the other profile.

Ideally, a DEM or other external data should not be required to compare near repeat-track elevations. A set of repeat-tracks containing a mixed elevation signal from local topography and temporal elevation changes between the observations. Several methods have been proposed to separate elevation changes from topographic variations. For example, Pritchard et al. (2009) fitted a triangular plane to three elevation observations, and used the plane as reference for measurements falling inside this triangle, while Howat et al. (2008), Moholdt et al. (2010a) and Moholdt et al. (2010b) used rectangular reference planes determined by least squares fitting to segments of repeat-track GLAS data. Regardless of the shape of the plane fitted, this approach eliminates the need of an external DEM. One potential disadvantage is the algorithm’s capability to separate surface slope from real elevation change signal, both of which are parameters derived within the algorithm. However, Moholdt et al. (2010b) showed that both along-track methods discussed here yield consistent results and agree well with the elevation changes calculated using the crossover point method. Sørensen et al., (2011) also obtained similar estimates of elevation changes over Greenland using ICESat data when applying different (i.e. DEM and plane) along-track methods.

Using modelling of the surface within 2 km by 350 m regions Horwath et al., (2012) applied along-track method to Envisat radar data, and achieved high spatial resolution of elevation change estimation over Antarctic of 0.75° by 0.2°.

The applied method for deriving along-track SEC from altimetry must be adapted to the specific mission and time span. During some missions, the satellites were placed in actual repeat-track with a (sub) cycle of ~30 days which was the case for ICESat and for Envisat in the period 2003-10. For these missions, the analysis can be carried out track by track.

After the orbit change of Envisat in 2010, the ground tracks shifted and started drifting, and a different method had to be applied to derive SEC along track after this orbit change. This method can also be used for ERS1/2 which were not kept in actual repeat track either.

Figure 2.4: (a) cross-track DEM projection (HDREF=HD₂+dHDEM) and linear interpolation to compare two repeat-tracks (dh=HDREF–HCREF), and (b) fitting least-squares regression planes to repeat-track observations to estimate slopes and average dh/dt (from Moholdt et al., 2010b).
2.4.2.2 Input / Output

Waveform L2 product for ERS-1/2 and GDR level 2 Envisat data of radar altimeter measurements will be used for producing surface elevation change estimates.

As mentioned earlier, two implementations of the along track method are applied: one method for repeat track data (RT method) and one for data from missions not in repeat track (AT method).

The RT method is applied to data in repeat track, because these often offer a good time and space distribution (Figure 2.5), which makes it possible to solve for both the topography and the SEC at the same time.

For other data sets, there is not a good time and space distribution along track (e.g. for Envisat after the orbit change in 2010). In that case, the AT method is applied instead, which requires a DEM as input to take into account the topography.

For both the AT and RT methods, the estimated SEC must be relocated by a slope correction (Bamber, 1994), which also makes use of the external DEM. As a last step, SEC values from ice-covered areas are selected by applying an ice mask.

The output of the processing lines is along-track point estimates of SEC in NetCDF or ASCII format.

Figure 2.5. Schematic of the RT and AT SEC processing lines.
2.4.2.3 Accuracy and performance

Accuracy of along-track methods is heavily dependent on the accuracy of across-track slope determination. At the same time, agreement between elevation change estimates obtained by the various along-track methods discussed above and the crossover analysis demonstrated good performance capabilities of these methods. Excluding the period of ERS-1 geodetic mission, satellite radar altimeters provide the large amount of repeat tracks that allow creating very detailed models of the ice sheet surface and forming elevation time series.

In another study (Horwath et al., 2012), elevation changes derived from the Envisat over the Antarctic Ice Sheet were compared with results of gravity changes from GRACE. In contrast to Thomas et al. (2008), the comparison showed a good agreement between linear trends and inter-annual variations that reflect surface mass balance changes. Although temporal changes of the surface properties are more pronounced in Greenland than in Antarctica, this result confirms the ability of radar altimetry to provide reasonable elevation change estimates.

2.4.2.4 Capabilities and limitations of algorithms

The main advantages of the along track methods are an increased quantity and spatial distribution of elevation change measurements in comparison to the crossover method. It increases the SNR of the analysis and increases the spatial resolution of the measurements. It also allows working locally along track to seek for local scale phenomena much better than the sparse crossover points. Furthermore, since the methods use data from ascending and descending tracks separately, it is not necessary to take into account any biases between them.

One disadvantage when using radar altimetry over ice sheets is that the radar-tracked surface changes with time; the penetration depth of the radar depends on the surface state. The measured height is then variable according to surface state variations or other volume echo intensity variations (linked to temperature changes impacting the medium’s absorption). A disadvantage of the plane fitting method is that the potential elevation change signal between the two repeat tracks is present in the reference plane.

Figure 2.6 from (Ewert et al., 2012) shows one of the challenges associated with any of the along track methods. The figure shows an along track segment covered by five overpasses at different times (t1-t5) of a satellite. The left sketch shows a good time and space distribution, while the right shows a bad time and space distribution. It is only possible to solve for SEC and topography at the same time if the data in an along track segment has a good time and space distribution.

If this is not the case, the topography have to be subtracted, e.g. by use of a DEM, but this introduces errors in the SEC if the DEM is not perfect.

---

**Figure 2.6:** Reasonable (c) respectively bad (d) time and space distribution of the repeat cycles within a box. from Ewert et al., 2012
2.4.2.5 Error characterisations

There are error sources and uncertainties originating from the data themselves that will influence the radar measured elevations and hence the estimated SEC. Among these are (i) uncertainties in the satellite orbit and attitude; (ii) atmosphere propagation corrections; (ii) solid earth tides and ocean tide to remove the effect of local tidal distortion to the Earth’s crust (REAPER handbook).

The SEC error budget is probably only affected by these EO data uncertainties, but rather dominated by the uncertainty related to the SEC derivation. When using the along track algorithms, it is necessary to make assumptions about the underlying topography, in order to separate this from the temporal changes in elevation (the SEC). It is unlikely that the assumed topography (plane, curved, DEM, etc.) will perfectly represent the actual topography, and this introduces errors in the derived SEC.

In general, a simple topography applies better to the central, flat areas of the Greenland ice sheet than the coastal areas characterized by a more complex topography. Therefore, the error is generally larger in areas with steeper surface slopes. Furthermore, the uncertainty on each individual elevation estimate is also slope dependent (Brenner et al., 2007).

One source of uncertainty which is not reflected by the error estimate is the fact that radar signal penetrates into the snow, and that the penetration depth varies in both space and time, being a function of snow properties. Therefore, it is uncertain exactly how the radar derived SEC relates to the physical snow surface elevation change.

A (small) part of the observed SEC is a result of Glacial Isostatic Adjustment (GIA); the response of the Earth to past changes in the ice loading (Whitehouse, 2009). This signal can be removed from the derived SEC by subtracting a model of the GIA signal, but such models are associated with uncertainties.

2.4.3 Merging – creating SEC grids

2.4.3.1 Principle of method

Collocation (Kriging) (Dermanis, 1984) is an interpolation method used for combining heterogeneous data of different kinds. Here, collocation is used for creating a surface (a grid) from the SEC crossover and along track point data. The interpolation method is described in detail in the DPM.

2.4.3.2 Input / Output

The input to the interpolation is the crossover SEC estimates, their associated error estimates, the along track SEC estimates, their associated error estimates, and an ice mask (a grid defining land cover type).

The output of the interpolation will be a 5km grid covering the entire Greenland ice sheet, and a grid with the corresponding errors for each grid cell.

![Diagram of Interpolation Processing Line]

Figure 2.7: Schematic of the interpolation processing line

2.4.3.3 Error characterisations

One benefit of using collocation is the fact that it takes into account the error estimates of the point data and also creates a grid of errors along with the SEC output grid. At grid points close to the input data locations (satellite tracks and cross overs) the SEC error is controlled by the input error estimates, and it then increases with distance to the input data locations.

2.5 Round Robin conclusions

A Round Robin (RR) exercise was carried out in order to determine which method (along-track or crossover) and which sensor (laser or radar altimeter) is most suited for deriving surface elevation change rates for the Greenland ice sheet, in accordance with GCOS requirements. During the RR, results received from different research groups were inter-compared thus allowing for an evaluation of the repeat-track
versus cross-over techniques as well as the use of laser vs. radar altimetry data for estimating surface elevation changes. Combined with a validation of the results using airborne LiDAR data (IceBridge data supplemented with CryoVEx and Danish PROMICE ice sheet monitoring flights), the outcomes of the RR experiment in the Jakobshavn Isbræ drainage sector of the Greenland Ice Sheet can be summarized as follows:

1) The SEC resolution is best from repeat-track algorithms, and generally better for the limited-footprint laser altimetry (ICESat) than for radar altimetry (Envisat). The comparisons have shown, however, that Envisat data have the potential to map height changes even on the relatively narrow fast-flowing ice stream of Jakobshavn Isbræ. This therefore shows the potential of the upcoming Sentinel-3 (S3) radar altimetry mission to provide reliable SEC values nearly all the way to the ice margin.

2) The SEC accuracy is superior using the cross-over method, at the price of limited resolution. This method is therefore less suitable to monitor changes in height associated with changes in glacier and ice stream flow velocities, but more suitable for estimating drainage basin-scale overall elevation changes.

For the practical implementation in the Ice Sheets CCI project, we therefore propose to implement a combined cross-over and repeat-track methodology. In practice, two separate methods will be implemented on an ice-sheet-wide scale: (i) a cross-over method with pre-defined cross-over points tailored to the specific missions (Envisat, Cryosat, Sentinel, IceSat-2), and (ii) a repeat-track method implementation, employing cross-track estimation for obtaining dense along-track SEC values, which could be improved over time as more and more detailed ice sheet heights are accumulated. Algorithms for this are already implemented for other applications among the CCI project partners. However, major adaption, tuning and optimization efforts will be necessary for implanting a truly operational and transparent system setup, including the application of state-of-the-art radar slope correction and backscatter corrections.

The results from the parallel applications of independent cross-over and repeat-track algorithms would subsequently be combined into a consolidated, interpolated SEC grid product, using optimal gridding procedures (collocation/Kriging) taking into account the specific error estimates of the individual estimation points. In the longer term, the two separate estimation methods could likely be combined into a single "hybrid" estimation method. This will, however, require an extra effort as well as extensive research.

2.6 Cryosat-2 SEC

As showed in the RR exercise, a hybrid estimation is needed ensure accurate SEC estimation. With the Cryosat-2 (CS2) radar altimetry data appearing this issue become more evident than before. The orbit configuration for CS2 differs from the previous orbits of ERS-1/2 and Envisat. Hence, the repeat-track altimetry applied along with the crossover method in the CCI SEC products for ERS-1/2 and Envisat (Nagler, 2015) cannot directly be transferred to CS2 SEC estimation. The crossover method can be applied to the orbit and data acquisition of CS2 but suffers from the low acceptance rate of crossover points in southern GrIS and the catchment area of Jakobshavn Isbræ. Here, we chose to develop a new method based on (Flament & Remy, 2012; S. L. S. Sørensen et al., 2011), but adopted to the orbit and data acquisition of CS2.

2.6.1 Principle of method

The elevation change (SEC) algorithm follows the formulation of the least-square model (LSM) presented by (Flament & Remy, 2012). To adapt this model to the data available in the ESA BL2i data-product for GrIS, we neglect the trailing edge slope (TeS) and add parameters for biases between LRM and SARIn, biases between ascending/descending orbits and seasonality in the SEC. The SEC is derived at predefined grid-nodes \((\text{lon}_c, \text{lat}_c)\).

A perturbation study of the LSMs is presented in the PVASR, and the most suitable LSM to solve for SEC with CS2 has been found to be

\[
\begin{align*}
    & h(\text{lon}_c, \text{lat}_c, t_i) - h_{\text{GIMP}}(\text{lon}_c, \text{lat}_c) = H_0 + \frac{dh}{dt}(t_i - t_0) + A\cos(2\pi t_i) + B\sin(2\pi t_i) \\
    & \quad + b_{\mu D}(-1)^{\mu D} + b_m(-1)^{m_i} \\
    & \quad + dB_s(B_S - B_{\text{S0}}) + dLW(LW_i - LW_{\text{S0}}),
\end{align*}
\]

Here, \(h(\text{lon}_c, \text{lat}_c, t_i)\) is the elevation measured by CS2 to a time \((t_i)\) at the geolocated longitude \((\text{lon}_c)\) and latitude \((\text{lat}_c)\), the surface curvature given by the GIMP DEM \((h_{\text{GIMP}}(\text{lon}_c, \text{lat}_c))\) (Howat, Negrete, & Smith, 2014). \(H_0\) is the elevation offset at the SEC grid point \((\text{lon}_c, \text{lat}_c)\). The seasonality in the elevation change is described as the superposition of sine and cosine, resulting in the amplitude \((A)\) and phase \((\phi)\) given by
L = AD_{descending} = 1$, are estimated as and the GIMP DEM is subtracted from the observations. After the selection of surrounding CS2 observations, an initial solution to the LSM is estimated, which is then evaluated in order to reject outliers in the observations. CS2 data points which cannot be represented by the LSM solution within $3\sigma$ are rejected. The outlier rejection procedure is a convergence of the system, which has proven to be 3 iterations. After the production of the SEC at a fine resolution grid, the results are post-processed by rejecting un-physical solutions and then a weighted averaging is done to transform the result to the final CCI 5km grid requirements.

2.6.2 Input / Output

The level 2 product provided by ESA is directly fed to the LSM algorithm along with the GIMP DEM and land mask. The main workflow of the algorithm is shown in Figure 2.8.

The output is gridded SEC in NetCDF format, following the specifications described in the PSD.

2.6.3 Accuracy and performance

The accuracy and performance follows the above description for conventional radar altimetry. CS2 differs in the effects of its orbit configuration and the new application of the SAR interferometer (SARin). The SARin insures better geolocation of the radar echo in the highly changing topography of the coastal area of the Greenland ice sheet.

2.6.4 Capabilities and limitations of algorithm

The capabilities and limitations follow the above described for the along-track method described in section 2.4.2.4. With the CS2 mission the plane fitting method becomes more vulnerable to the orbit configuration, as shown in Fig. 2.6, as the evaluation of SEC at predefined grid points is not chosen based on the orbit, but rather to insure high spatial coverage.

The radar altimeter is measuring the “highest” point within the footprint of the satellite. Hence, SEC estimate is a solution to the time evolution of the highest points with in the CS2 footprint. This becomes especially evident in the area of Jakobshavn Isbræ, where the fixed grid solution struggles to provide accurate estimates of SEC in the main trough.

2.6.5 Error characterisations

As shown in section 2.4.2.5 the general error sources affect the CS2 SEC, including the issue of snow penetration. The perturbation study shows the need of applying a reference DEM. This should remove the errors associated with estimating the topography in the plan fitting method.
2.7 References


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3 Ice velocity

3.1 Introduction

Based on the URD, the IV algorithms must be able to generate ice-sheet wide “static” maps of mean-seasonal velocities as well as time-series of velocity measurements. The term velocity should be understood as “velocity vector”, since users require all its components to be measured. These and other requirements are summarized in Table 3.1, where also the GCOS requirements are reported for convenience. The core measurement techniques for IV are quite well established and reviewed in section 3.3. In order to meet the above requirements table, an algorithm combining these core-techniques is required.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>URD</th>
<th>GCOS</th>
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</thead>
<tbody>
<tr>
<td>Velocity components</td>
<td>Cartesian (ENU)</td>
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<tr>
<td>Spatial coverage</td>
<td>Whole ice-sheet</td>
<td>-</td>
</tr>
<tr>
<td>Minimum spatial resolution</td>
<td>100m-1km</td>
<td>1 km</td>
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<tr>
<td>Optimum spatial resolution</td>
<td>50m</td>
<td></td>
</tr>
<tr>
<td>Minimum temporal resolution</td>
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<td>Monthly</td>
</tr>
<tr>
<td>Optimum temporal resolution</td>
<td>Monthly</td>
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<tr>
<td>Minimum accuracy</td>
<td>30 m/y</td>
<td>10 m/y</td>
</tr>
<tr>
<td>Optimum accuracy</td>
<td>10 m/y</td>
<td></td>
</tr>
</tbody>
</table>

3.2 Review of scientific background

Space-borne IV measurements are currently carried out with SAR or with optical sensors. Measurement techniques based on SAR data are either phase-based (DInSAR and MAI) or offset-based (speckle-, feature- and coherence-tracking), whereas those based on optical data are only offset-based (feature tracking). The main characteristics of each method are described in the following sub-sections.

3.3 Algorithms

Three methods are commonly used to derive ice velocity from satellite image data:

- Differential SAR Interferometry DInSAR
- Multiple Aperture Interferometry (MAI)
- Offset Tracking

Generally, a combination of the core measurement algorithms will be used to derive the foreseen IV ECV parameters. An overview of the processing chain is sketched in the block diagram of Figure 3.1, where processing chain branches for the optical and SAR data respectively. The input data consists of imagery acquired by an arbitrary number of SAR and optical sensors. SAR images will in general be acquired from a number N of different ground tracks, either partially overlapping or non-overlapping among each other. The same applies for optical images.

For the SAR processing chain, the core techniques of sections 3.3.1 and 3.3.2 shall be combined to generate slant-range and azimuth velocities, denoted as V_r and V_a respectively in Figure 3.1, with associated error standard deviation maps. The combination of the different algorithms shall exploit the known synergies and complementarities between the methods (Joughin, 2002). Cartesian velocity components shall subsequently be computed from all possible sources, which may provide two or more independent measurement components, namely:

- V_r and/or V_a measurements from the same pass (e.g. descending);
- V_r and/or V_a measurements from intersecting ascending and descending passes.
In case less than three independent velocity measurements are available, the Surface Parallel Flow assumption can be applied for deriving the Cartesian velocity components.

In view of the future Sentinel missions, that will provide a denser archive of acquisitions with shorter temporal baseline, a multi-temporal algorithm could be used. The latter would process a batch of acquisitions from a common ground-track, using a redundant set of measurements and spatio-temporal filtering methods to provide higher accuracies, at the same time reducing operator time compared to processing individual image pairs.

### 3.3.1 DInSAR

#### 3.3.1.1 Principle of method

DInSAR techniques include DEM elimination, DEME, (Joughin et al., 1996) and Double Difference, DD, (Kwok and Fahnestock, 1996). Like all InSAR methods, both techniques exploit the fact that the phase difference between two SAR acquisitions is sharply peaked around a value proportional to the differential range from the surface, provided the radar speckle is statistically similar (i.e. the images are coherent).

The differential range contains a topographic as well as a displacement contribution, which needs to be isolated. In DEME a single InSAR pair is required, and the contribution of topographic variations to the interferometric phase, is estimated from an external DEM. In DD, two InSAR pairs are required, and velocity and height are co-estimated, under the assumption that the underlying velocity is constant in time.

Both methods measure only the line-of-sight component of the velocity. Temporally the latter is referred to the period between the two SAR acquisitions, for DEME, and between the two acquisitions forming each InSAR pair for DD.

#### 3.3.1.2 Input / Output

DEME requires one coherent image pair and an external DEM, and provides a velocity measurement in the SAR Line-of-Sight (LoS).

DD requires two coherent image pairs and provides a velocity measurement in the SAR LoS as well as a height measurement. An external DEM is not mandatory, but if available can improve the accuracy of the results, by aiding the phase unwrapping process.
Both methods also require a set of Ground Control Points (GCPs) on input, i.e. points of known height and velocity, which are required to estimate the absolute phase and image-wide error trends, due to orbital uncertainties and long-wavelength atmospheric signals.

![Differential Interferometry processing chain](image.png)

| Figure 3.2: Differential Interferometry processing chain. Blue boxes represent functional blocks, whereas red ones represent input data. In the DEME algorithm only one input image pair is required, whereas in the DD two must be provided. The implementation of the geophysical inversion step will differ for the two methods, although conceptually the same task is carried out. |

### 3.3.1.3 Accuracy and performance

DInSAR methods suffer from bias and random errors (Mohr and Merryman, 2008). The former are introduced in the so called “baseline calibration” procedure, in which GCPs are used to refine the orbital information based on the satellite state-vectors, as well as in the phase-unwrapping procedure.

Baseline calibration errors are spatially variable, and depend on the distribution and accuracy of the available GCPs (Joughin et al., 1996). Phase unwrapping errors may cause different areas of the displacement map to have a relative error of multiples of half the radar wavelength. Since baseline calibration is performed on the unwrapped phase, errors can propagate if GCPs are chosen in erroneously unwrapped areas.

Random errors, with a zero mean, are due to decorrelation, atmospheric propagation and, for DEME, to the accuracy of the external DEM. Decorrelation is mainly due to temporal changes of the ice-surface properties, to penetration beneath the surface (volume scattering), and displacement within the radar range resolution cell.

A detailed error analysis is considered outside the scope of this document. Assuming no bias errors, the accuracy of DInSAR phase measurements is in the order of $\pi$ rad (Joughin, 2002), which translates to a ground-range velocity accuracy in m/y of:

$$\sigma_{vgr} = \frac{1}{T} \cdot \frac{\lambda}{4} \cdot \frac{365}{\sin \theta}$$  \hspace{1cm} (3.1)

where $\theta$ = SAR incidence angle, $T$ = temporal baseline [days], $\lambda$ = SAR wavelength [m].

For the shortest temporal baseline in the archive, $T = 1$ day, and ERS sensor properties, $\lambda = 5.6$ cm, $\theta = 23$ deg, the expected accuracy is in the order of $\sigma_{vgr} = 13$ m/y.
3.3.1.4 Capabilities and limitations

Applicability of DInSAR techniques is limited by the following factors:

- The temporal separation of the image acquisition pairs must be short enough for ice-sheet properties not to change, at the penetration depth of the radar wave.
- The perpendicular component of the spatial baseline must be small enough to reduce the decorrelation due to volume scattering.
- The displacement gradients must not exceed \( \lambda/2 \) in a range resolution cell, or correlation will be completely lost.

For Sentinel-1 (S1) TOPS mode acquisitions, interferometric processing is complicated by the fact that the azimuth-varying Doppler centroid in TOPS mode causes azimuth phase ramps in the impulse response function in the presence of azimuth misregistration. The cause of this can be timing errors and/or actual physical motion. This leads to interferometric phase errors, and sets stringent requirements on the azimuth coregistration accuracy (Scheiber et al., 2015). Phase unwrapping is also complicated by the fact that phase discontinuities occur at burst edges where the line-of-sight jumps abruptly.

At C-band, the largest temporal and perpendicular baselines for the Greenland ice-sheet are in the order of 30 days and 300 m respectively (Hoen and Zebker, 2000), although these values are site-dependent. Concerning the displacement gradient limitation, since large gradients are typically associated to large absolute speeds, this translates in a limit on the latter. For Radarsat-1 24-day pairs, maximum measurable speeds of 125 m/y are reported (Gray et al., 2001). This approximately corresponds to the velocity gradient over a 3000 m distance. Assuming this as a general criterion, the maximum measurable ground-range velocity in m/y would be

\[ v_{gr,max} = \frac{\lambda}{2\delta_\nu} \times \frac{365}{36} \times 3000 \]  

where \( \delta_\nu \) is the slant-range resolution of the SAR. For ERS 1-day, 3-day and 6-day pairs, \( v_{gr,max} = 3 \) km/y, 1 km/y and 500 m/y respectively.

3.3.1.5 Error characterisations

Ionospheric propagation errors arise due to spatial fluctuations (scintillations) in the ionosphere Total Electron Content within the synthetic aperture length (i.e. km-scale variations). This is especially a problem in the near-polar regions. For a given image pixel, these fluctuations cause an azimuth variation in the raw signal phase, which is not accounted for by the SAR focusing, resulting in an azimuth shift of the focused pixel. The varying propagation naturally also causes a shift in the range direction, but these shifts are much smaller (typically on the centimeter-level) than those observed in the azimuth direction (comparable to the azimuth pixel size, i.e. several meters) (Mattar, 2002). The shifts vary along the scene according to the ionospheric conditions along the satellite flight path, often being present in only parts of the scene. Also the observed shifts are strongly correlated in the range direction, appearing as linear or slightly curved “streaks” in the azimuth shift map and the interferometric phase and coherence. The shifts observed are generally more severe at lower frequencies, with range shifts of up to 7 cm observed at L-band (JERS-1) and up to 1 cm at C-band (ERS/RADARSAT) (Mattar, 2002).

For the DInSAR algorithms, the main impact of ionospheric errors is the loss of coherence associated with the azimuth misregistration, which requires an estimation of azimuth misregistration on a dense grid and a resampling of the two SLC’s prior to coregistration, using the (perhaps range-filtered) azimuth shift map directly, rather than a polynomial shift model. The error introduced by ionospheric effects on the ice velocity estimate is inversely proportional to the temporal baseline. With the maximum range shift of 1 cm for ERS/RADARSAT mentioned above, this translates to a maximum ground range velocity error of approximately 10 m/yr assuming a 1-day ERS tandem interferogram and 23\(^\circ\) incidence.

Phase unwrapping errors are introduced typically when segments of the image are connected by low coherence regions (e.g. open water/sea ice/fast decorrelating glaciers/layover). Often, these segments can be identified and unwrapped separately using segmentation based on image amplitude and coherence, and an absolute phase can be estimated for each segment if ground control points (stationary or with known velocity) are available. Typical phase unwrapping errors are single cycle errors, leading to range shifts of \( \lambda/2 \), but if GCP’s are placed on incorrectly unwrapped segments, the errors can propagate into the baseline estimate.

In the DEME algorithm, DEM errors cause a displacement error proportional to the perpendicular baseline. If the displacement due to DEM error is large enough, it can affect the coregistration and thus degrade the
coherence, which can be a problem for high resolution sensors like TerraSAR-X. The displacement can cause an error on the ice velocity estimate of

$$\Delta \nu_{gr} = \frac{2\pi}{\gamma} \cdot \frac{r}{\sin \theta} \Delta h$$

For ERS interferometry, usually baselines up to 300 m are used. For a tandem ERS interferogram with this baseline, a DEM error of 100 m would translate into an ice velocity error of 40 m/yr, assuming $r=700$ km, and 23° incidence.

### 3.3.2 MAI

#### 3.3.2.1 Principle of method
MAI measures the azimuth displacement component between two acquisitions, by analysing the difference between a forward and a backward looking interferogram. These are formed using two appropriately chosen squint angles and bandwidths (Bechor and Zebker, 2006), so that the line-of-sight contribution to the interferometric phase cancels out in the difference, while retaining coherence.

#### 3.3.2.2 Input / Output
MAI requires one coherent image pair and provides a velocity measurement in the SAR flight direction (azimuth).

An external DEM, if available, can be used to improve accuracy of the results.

A set of Ground Control Points (GCPs) is required, i.e. points of known height and velocity, in order to estimate the absolute phase and image-wide error trends, due to orbital uncertainties and long-wavelength atmospheric signals.

Figure 3.3: Multi Aperture Interferometry processing chain. Blue boxes represent functional blocks, whereas red ones represent input data.

#### 3.3.2.3 Accuracy and performance
MAI suffers from the same error sources discussed in the DInSAR section, 3.3.1.3. As an order of magnitude, the measurement accuracy of MAI is halfway between InSAR and offset-tracking. Assuming no bias errors, and a sufficiently high coherence and phase averaging, in order for the phase estimates to
have an accuracy better than $\pi/40$ rad, based on (Bechor and Zebker, 2006), the velocity accuracy in m/y is given by:

$$
\sigma_{vga} = \frac{1}{T} \cdot \frac{l}{80} \cdot 365
$$

(3.4)

where $T$ = temporal baseline [days], $l$ = SAR antenna length [m].

An issue poorly documented in literature, and which will need to be investigated, is the sensitivity of this technique to km scale variations in ionospheric Total Electron Content.

### 3.3.2.4 Capabilities and limitations

The MAI technique offers higher spatial resolutions, for a given measurement accuracy, or vice-versa, compared to the azimuth velocities measured with offset-tracking.

It requires coherence, and in this respect suffers from the same limitations previously discussed for DInSAR. This includes a low limit on the maximum measurable velocities, which is expected to limit its application to rapidly flowing outlet glaciers. MAI is expected to fail for displacements in the order of a significant fraction of the SAR antenna length. Extending the results observed for the ERS system, (Bechor and Zebker, 2006), the maximum measureable velocity in m/y would be:

$$
v_{a,max} = \frac{l}{5} \cdot \frac{365}{T}
$$

(2.5)

where $T$ = temporal baseline [days], $l$ = SAR antenna length [m].

### 3.3.2.5 Error characterisations

Since MAI was not selected as a method for IV retrieval within this project, an error characterisation is not presented here. It is noted that the effect of ionospheric scintillations is expected to be a significant problem when applying MAI in the polar areas.

### 3.3.3 Offset Tracking

#### 3.3.3.1 Principle of method

The term Offset Tracking is used here to refer to a family of methods, which includes speckle-tracking (Gray et al. 2001; Joughin, 2002), coherence-tracking (Derauw, 1999) and feature-tracking (Lucchitta et al., 1995; Michel and Rignot, 1999; De Lange et al., 2007). The same approach is applied for SAR and optical images and will therefore not be described separately. However, SAR data have the advantage of an active sensor that is not affected by solar illumination (day/night) or cloud coverage which is special value in high latitudes.

All methods estimate the spatially varying misalignment (offset) of a pair of SAR or optical images, due to surface motion, and provide velocity measurements in sensor geometry (slant-range and azimuth for SAR, horizontal for optical).

In speckle- and feature-tracking, offsets are determined by cross-correlating image patches. The same approach is used for optical data, but with different parameterisations (e.g. template size). Correlation of complex values in SAR images allows for smaller cross-correlation window sizes, e.g. 24 x 24 pixels (Joughin, 2002), compared to amplitude or intensity correlations, e.g. 192 x 192 pixels (Joughin, 2002), 64 x 256 (De Lange et al., 2007), 256 x 256 pixels (Floricioiu et al. 2010). On the other hand, it provides an undesirable sensitivity to phase gradients, which, if uncompensated, may obscure the cross-correlation peak.

In coherence tracking, the offsets which maximize interferometric coherence within a certain window size are determined. Typical window sizes range from 5 x 5 pixels (Derauw, 1999) to 8 x 8 pixels (Strozzi et al., 2002). Due to the estimation window sizes, and subsequent averaging of the offset values to reduce spatially uncorrelated error sources, offset tracking techniques offer a coarser spatial resolution compared to DInSAR (hundreds of meters vs. tens of meters).

For Sentinel-1 TOPS mode acquisitions, offset-tracking algorithms are essentially the same, but must support burst handling and stitching. Also, the azimuth-varying phase introduced by the TOPS beam steering must be accounted, both for complex offset tracking and for incoherent offset-tracking, where a factor of two oversampling of the SLCs is required prior to detection.

#### 3.3.3.2 Input / Output

All offset tracking methods require one image pair acquired from the same ground track (optical or SAR). Additionally, speckle- and coherence-tracking have the same coherence requirements described for
DInSAR methods. All methods provide line-of-sight as well as azimuth velocity measurements when applied on SAR data, and the 2D horizontal displacement when applied on optical data.

An external DEM is required to geocode the measurements and will improve the accuracy of slant-range measurements, in particular for high-resolution sensors.

All methods also require a set of Ground Control Points (GCPs) on input, i.e. points of known height and velocity, which are required to estimate the image-wide co-registration error trends, due to orbital uncertainties and/or long-wavelength atmospheric signals.

![Offset tracking processing chain](image)

**Figure 3.4:** Offset tracking processing chain. Blue boxes represent functional blocks, whereas red ones represent input data. Offset calculation may be performed by complex or intensity cross-correlations, by coherence maximization, or by a combined approach.

### 3.3.3.3 Accuracy and performance

Speckle- and coherence-tracking accuracies depend on the level of coherence and on the correlation window size (Bamler and Eineder, 2005). For a coherence of 0.3 and a complex correlation window of 24 x 24, offsets of 1/20th of the resolution cell can be achieved in ground-range and in azimuth (20 cm in azimuth and 1.5 m in ground range for ERS-AMI or ASAR). Correlation of intensity images can be applied to coherent as well as to incoherent image pairs and optical data. In the latter cases cross-correlation peaks are due to common features in the two images. Accuracies are typically 2 times worse than for speckle-tracking (De Lange et al., 2007).

### 3.3.3.4 Capabilities and limitations

Applicability of speckle- and coherence-tracking techniques is limited by the same factors described in 3.3.1.4, which limit phase coherence. In practice though, compared to DInSAR, measurements can be carried out for lower coherence levels, because no unwrapping procedure is required. Sensitivity to DEM uncertainties is lower for this family of methods, compared to DInSAR.

Feature-tracking is not affected by low or absent coherence, nor does it pose any practical limit on the maximum measurable velocity magnitude, within the range of velocities of the Greenland and Antarctic ice-sheets (< 13 km/y). However, it requires common features to be present in the input SAR or optical images, which causes the method not to be applicable in the interior of the Greenland ice-sheet.

### 3.3.3.5 Error characterisations

Since the main effect of ionospheric scintillations are the azimuth shifts introduced by the fluctuating electron density along the sensor path, this has a large impact on offset-tracking, which retrieves both range and azimuth offsets. Misregistration is not a problem, but the large azimuth shifts will be interpreted as ice motion, and observed offsets can exceed the azimuth pixel size. For ERS, with a 35-day temporal baseline, a shift of one azimuth pixel (4 m), would be interpreted as an ice velocity of 40 m/yr, and shorter temporal baselines would suffer even more.

DEM errors have less impact on the algorithm performance compared to DInSAR, since there is no phase unwrapping. The effect of DEM errors on the ice velocity estimate is equivalent to that for DEME DInSAR, given equivalent temporal baselines. However, since offset-tracking can be applied with longer temporal baselines, and even though larger spatial baselines can be used for offset-tracking, in the typical scenario the sensitivity to DEM errors is much smaller.
3.4 Round Robin conclusions

The open Round Robin experiment has provided important input to the algorithm selection. Few surprising findings were made, but the fact that almost all foremost groups took part in the exercise means that the algorithm selection can be made on a more solid basis.

The overall conclusion is that the offset tracking should be the primary ice velocity algorithm. Where, needed and where feasible, DInSAR can be used as a supplement, whereas MAI will not be implemented at all.

The offset tracking techniques have several advantages. They
1) provide two velocity components from one image pair
2) require less operator interaction than the interferometric techniques, i.e. they can be more easily and reliably automated
3) are more robust in the sense that they do not involve phase unwrapping, they require less (or no) coherence, they can be applied to data with longer temporal baselines (though 35 days is generally too long) and to glaciers with larger velocity gradients (faster glaciers).

The Round Robin experiment demonstrated that an adaptive implementation involving coherent cross-correlation of small windows and intensity or amplitude cross-correlation of larger windows is preferable, as a better coverage was obtained by a group using such an adaptive approach.

On the other hand, the Round Robin experiment clearly demonstrated that, where applicable and successful, interferometric techniques provide more precise ice velocities. The lack of robustness and the need for skilled operator interaction was illustrated by a Round Robin example where a GCP in an erroneously unwrapped region led to large errors (which were fortunately flagged by the error prediction map). In the Round Robin experiment, DInSAR was applied to ERS tandem data with a temporal baseline of 1 day, but for faster glaciers, DInSAR would fail due to decorrelation and unwrapping problems.

MAI, has the same disadvantages as DInSAR, but in addition it provides less precise ice velocities, and it is more sensitive to ionospheric scintillations, as demonstrated in the Round Robin experiment.

3.5 References


4 Calving front location

4.1 Introduction
The Calving Front Location (CFL) of outlet glaciers from ice sheets is a basic parameter for ice dynamic modelling, for computing the mass fluxes at the calving gate, and for mapping glacier length or area change. From the ice velocity at the calving front and the time sequence of Calving Front Locations the iceberg calving rate can be computed which is a direct measure of the ice mass lost to the ocean.

4.2 Review of scientific background
To date, no automatic method of picking CFL from SAR images has been reported in the literature. There is only one study using an automated method for extracting termini positions based on optical MODIS images (Seale et al., 2011). Further, Joughin et al. (2008) applied a “simple edge detection algorithm” on MODIS data, but without integration into an automated workflow. For Greenland, CFL has been quantified and investigated for single outlets (Luckman et al., 2006; Stearns & Hamilton, 2007; Joughin et al., 2008), limited regions (Howat et al., 2008; Murray et al., 2010, McFadden et al., 2011; Box & Decker, 2011, Walsh et al., 2012), or temporal snapshots (Moon & Joughin, 2008; Howat et al., 2010, Howat & Eddy, 2011). All the studies are based on the manual digitisation of the termini positions of the investigated glaciers from various types of satellite images (optical and SAR, very high to medium resolution).

4.3 Algorithms
Currently manual or semi-automated techniques are applied to map calving fronts of glaciers and ice streams.

<table>
<thead>
<tr>
<th>Manual CFL extraction</th>
<th>Automatic CFL extraction</th>
<th>SAR data</th>
<th>Optical data</th>
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4.3.1 Manual Delineation of CFL

4.3.1.1 Principle of method
This method is based on manually identifying CFL based on spatial patterns in reflectance, backscatter and texture in the underlying satellite image. In general it utilises geocoded images in a GIS environment where the frontal positions are digitised in the same reference frame.

4.3.1.2 Input / Output
SAR and optical satellite sensors provide excellent capabilities for monitoring CFL. However, SAR data (ERS, ASAR, and ALOS) will be utilised as primary data source for ECV parameter production, as CFL extraction is a sub-branch of the IV processing chain which is mainly designed for SAR data input. The processing chain will start from Level 0 data with SAR data focussing as the first processing step and is outlined in Figure 4.1. The output is as series of latitude longitude vertices stored as vector line in
standard GIS format or optionally as ASCII list. Additionally, metadata information on the sensor and processing steps are stored in the corresponding attribute table (compare PSD for details on the format).

In addition, the SAR images used for CFL production have the advantage of being independent of solar illumination and cloudiness, therefore allowing monitoring the full annual cycle including the arctic winter. Optical images (from Landsat-8, Sentinel-2 (S2) and other high resolution sensors) can be used for filling gaps in the temporal sequence of ice velocity. Sentinel-1A and 1B will provide an operational coverage of the outlet glaciers providing the data for monitoring the ice dynamics and temporal changes. For CFL mapping only the bursts covering the glacier terminus will be extracted from the slices which speeds up the processing time.

4.3.1.3 Accuracy and performance

Primary sources of error for manual CFL mapping are errors in manual selection of the ice front position, image co-registration errors due to terrain or other factors, and uncertainty in feature locations due to pixel resolution (Moon & Joughin, 2008, Howat & Eddy, 2011). McFadden et al. (2011) assessed the relative positioning error between images by > 20 off-ice control points. They report average errors of ±74 m, ±87 m, and ±25 m for Landsat, RADARSAT, and SPIRIT images respectively relative to ASTER images. For medium resolution MODIS images, Box & Decker (2011) digitised the same glacier terminus three to five times and find that inter-annual area changes exceed two standard deviations of the repeat samples in 63% of cases.

4.3.1.4 Capabilities and limitations

Manual delineation as well as any automatic / semi-automatic method works well when the calving front can be clearly discriminated against open (calm) water due to different reflectivity, texture and shape in SAR and optical imagery, as indicated in the right panel of Figure 4.2. Additionally, the identification of the calving front benefits from a better spatial resolution (Figure 4.3); however, the presence of the ice melange in front of the calving cliff can impede the detection of the frontal position. This is shown with the comparison of the two TerraSAR-X images in Figure 4.2. Further, the ice melange can cause ambiguities in the interpretation at any spatial scale as highlighted in Figure 4.3 with the red frame in the comparatively high resolution TerraSAR-X quicklook.
Figure 4.2: TerraSAR-X images of an area at the East coast of Greenland (centre approx. at 67.55N, 33.54E) with (left) and without (right) ice melange accumulated in front of the calving cliffs (images © DLR).

Figure 4.3: Compilation of satellite images from different sensors over the same outlet glaciers in West Greenland (approx. at 69.94N, 50.2E). Top left Landsat-7 ETM (© USGS), Top right SPOT-5 HRS (© CNES), bottom left ENVISAT-ASAR WSM (© ESA), bottom middle ENVISAT-ASAR (© ESA), bottom right TerraSAR-X (© DLR).
Furthermore, precise geocoding as well as the spatial resolution of the original satellite data are the main issues for the quality of the product, the extracted vector line of the calving front position (shape file) at a given date (see section 4.3.1.3).

There is hardly any in-situ validation data available for validating CFL. Therefore we plan to use cross-checks with very high resolution optical data (SPOT-5 HRS, 5 m resolution, from SPIRIT campaign) which is shown in in the top right panel of Figure 4.3.

4.3.2 Automatic extraction of CFL

Only Seale et al. (2011) describe an automated method for extracting termini positions based on optical MODIS images. No automatic method exists at the moment for picking CFL from SAR images, also it is planned within Ice_Sheets_cci project to develop a change detection algorithm or line detector to map the frontal position automatically based on SAR data. However, no datasets have been processed and the experimental processing line has not been developed and tested at this stage. Therefore, potential developments in the course of the project will be described and included on the follow-up ATBD documents (e.g. ATBDv1).

4.3.2.1 Principle of method

The method described by Seale et al. (2011) allowed processing of daily MODIS images for 32 glaciers, totalling 105,536 images. Although no winter positions could be derived, due to the insufficient solar illumination of large parts of Greenland during that time (2-4 month depending on latitude). The processing chain starts with a quality assessment based on cloud mask products. Then, two edge detection algorithms (Sobel, Brightness profiling) are applied to image chips cut out by pre-defined frames. In several following steps, the quality of the picked line is assessed (scoring scheme and manual removal) with a time series of glacier margin positions as the final output.

4.3.2.2 Input / Output

The method described by Seale et al. (2011) uses MODIS images (250 m spatial resolution) from the Terra satellite, as well as the MODIS cloud identification product and a set of polygon frames (as ‘cookie cutter’) as input.

It is planned within Ice_Sheets_cci to investigate possibilities for an automatic processing line based on ENVISAT-ASAR WSM data. This experimental processing line for extracting CFL from ASAR WSM data is envisaged to start with Level 1 data. However, this approach is under development and cannot be described in detail at this stage (see section 4.3.2).

4.3.2.3 Accuracy and performance

Seale et al. (2011) performed a validation exercise against a dataset of margin position derived by a different group and report a RMS error of 0.19 km which is smaller than the MODIS resolution (0.25 km) the automatic extraction was based on.

4.3.2.4 Capabilities and limitations

Obviously, an automatic picking of glacier terminus positions can be a powerful tool to map frontal variations of a large number of glaciers with a high temporal revisiting rate. However, the only automatic approach that is reported still requires a large amount of manual intervention in the post-processing (Seale et al., 2011). An automatic picking based on SAR data does not exist at the moment.

4.3.2.5 Error characterisations

The error of the CFL location includes various components, (i) error due to automatic processing and (ii) error due to manual delineation of the CFL. The first error source is made of the geolocation accuracy, which depends mainly on the accuracy of the orbit parameters and DEM accuracy. The second error source is the manual delineation. This error source was investigated in the RR tests and depends on the resolution of the SAR images (high resolution images are, the conditions of the ocean surface in front of the calving front, the orientation of the front relative to the imaging geometry. This error can be estimated according to the CFL RR tests (not taking outliers into account).

4.4 Round Robin conclusions

To date, no automatic method of picking CFL from SAR images has been reported in the scientific literature and this is reflected by the contributions received in the Round Robin experiment. All participants manually digitized the termini locations in a GIS environment.

Extracting the calving front position of a marine terminating glacier from SAR is not an easy task. Generally, the errors we found in the RR contributions are about an order of magnitude higher than
expected. However, it has to be mentioned that we deliberately selected some poor data and chose difficult circumstances in order to cover the full range of potential scenarios. Interestingly, the errors seem to be more or less independent from the resolution of the dataset and settle at 5.05 pixels for the SAR scenes with 20 m resolution and 4.13 pixels for the 75 m ASAR wide swath data that were tested.

Semi-automatic procedures might support the detection of the CFL in the future, but this is still under development. For the manual extraction of calving front positions we define a set of guidelines for the ECV production based on the results of the RR experiments that will help to increase quality and accuracy of the product:

- Digitisation should be performed by an experienced operator only.
- High resolution SAR data (e.g. ERS-1/2, ENVISAT-ASAR image mode) are preferable.
- Poor data should be rejected without CFL extraction.
- Sections of the fjord walls on either side of a calving front should be included in the digitisation that can be used for relative quality assessment.
- Data should be projected to the local UTM zone in order get quadratic pixels and therefore minimize any distortion.

4.5 References


5 Grounding Line Location

5.1 Introduction
The grounding line separates the floating part of a glacier from the grounded part. Processes at the grounding lines of floating marine termini of glaciers and ice streams are important for understanding the response of the ice masses to changing boundary conditions and to establish realistic scenarios for the response to climate change (Gladstone et al., 2010; Nick et al., 2010). Further, the migration of the grounding line is a sensitive indicator of ice thickness change and the Grounding Line Location (GLL) is listed as “important parameters for ice sheets” in the IGOS Cryosphere Theme Report.

5.2 Review of scientific background
A substantial collection of literature can be found on detecting GLL in Antarctica (e.g. Fricker & Padman, 2006, Rignot et al., 2011). However, there are only few studies on grounding lines in Greenland on a very limited number of glaciers (e.g. Petermann glacier: Rignot, 1996 and 1998a; Storstrommen glacier: Mohr et al., 1998, Reeh et al., 2003). This is likely due to the fact that only a small number of glaciers with floating tongues can be found in the north of Greenland. It has to be mentioned, that the ocean tide causing the floating part of a glacier or ice shelf to bend up and down is the only (indirect) measure of the grounding line, either from space or on the ground (e.g. Vaughan, 1995, Smith, 1991, Rabus, 2002). There are no direct observations of a grounding line at the bottom of a floating tongue or shelf. The tidal motion of a floating glacier tongue or ice shelf causes a series of surface features and effects that can be summarized as the so-called grounding zone which is illustrated in Figure 5.1.

![Figure 5.1](image)

Figure 5.1: Schematic of features found in the grounding zone (Fricker et al., 2009). F is the landward limit of ice flexure from tidal movement, G is the limit of ice flotation, i.e. the grounding line, $I_b$ is the break in slope, $I_m$ is the local elevation minimum; and H is the landward limit of the hydrostatic zone of free-floating ice shelf, or the seaward limit of ice flexure. The region between F and H is defined as the grounding zone, which is typically several kilometres wide.

5.3 Algorithms
Several approaches have been reported to measure the Grounding Line Location:
- InSAR
- Optical satellite data
- Altimetry

A selection of studies on the different approaches can be found in Table 5.1.
Table 5.1: Selection of GLL related studies with the applied methods and location.

<table>
<thead>
<tr>
<th>Study</th>
<th>InSAR</th>
<th>Optical</th>
<th>Altimetry</th>
<th>in situ / model</th>
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5.3.1 InSAR

5.3.1.1 Principle of method

The principles of InSAR processing are described in section 3.3.1.1. In order to isolate the tidal motion from the interferogram, the other contributions to the signal need to be eliminated. This can be done by combining two repeat satellite overflights where coherence is retained and form a radar interferogram, which is then corrected for surface topography using a prior-determined precision digital elevation model (DEM). Two topography-corrected interferograms spanning the same time interval are then differenced to yield an interferogram which measures only the surface deformation associated with the tidal motion of the glacier (Rignot, 1996, 1998a).

5.3.1.2 Input / Output

GLL extraction requires interferometric processing and therefore coherent data which is only available during the three ERS ice phases (Dec 1991 – Mar 1992, Dec 1993 – Apr 1994, Mar 2011 – Jul 2011) as well the ERS-1/2 tandem phase (Mar 1995 – Jun 1996). GLL processing relies on the same data set as IV, and also SAR focusing and InSAR processing is part of the IV processing line. InSAR can only deliver the surface expression of the so-called grounding zone which is the area of the glacier or ice shelf affected by tidally induced flexure (Rignot, 1996, Rabus & Lang, 2002). Further details on the input and output data can be found in section 3.3.1.2.

With the launch of Sentinel-1 in April 2014 a new SAR data set is available. The main acquisition mode of Sentinel-1 is Interferometric Wide Swath Mode, which applies TOPS for acquiring the data with 12 days repeat cycle. Due the repeat interval of 12 days low coherence is expected over fast moving outlet glaciers, which does not allow to form interferograms for GLL delineation. Using Sentinel-1A and the upcoming Sentinel-1B with 6 days repeat will provide some improvements in this sense. Further tests will shows the suitability of S1 Tops mode data for GLL mapping.
Figure 5.2: Schematic of the GLL processing line based on DInSAR processing, which is outlined in section 3.3.1.

5.3.1.3 Accuracy and performance

The accuracy and performance of InSAR for detecting ground motion is outlined in section 3.3.1.3.

5.3.1.4 Capabilities and limitations

The general capabilities and limitations of the InSAR approach are described in section 3.3.1.4. However, for applying this method for detecting the GLL, additional analysis steps are required. As mentioned before, InSAR is capable of measuring the ocean-induced vertical displacements related to the grounding zone. The actual grounding line is found below the landward limit of flexure (compare to Figure 5.1) and therefore additional calculations are required to extract this position from the interferogram. This is done by applying buoyancy calculations, elastic beam models, or tide model predictions (e.g. Rignot et al., 2000, Rabus & Lang, 2002). However, these approaches make assumptions on the density of the ice and the ice thickness needs to be known at the grounding zone.

5.3.2 Altimetry

5.3.2.1 Principle of method

The tidally induced elevation changes of a glacier or ice shelf can also be detected with repeat-track analysis of altimeter data, and has been successfully applied in several studies, all focussing on Antarctica (e.g. Fricker & Padman, 2006, Fricker et al., 2009, Brunt et al., 2010). A detailed description of this method can be found in section 2.4.2

5.3.2.2 Input / Output

The input and output data of the repeat-track approach are detailed in section 2.4.2.
5.3.2.3 **Accuracy and performance**

The general accuracy and performance of the repeat-track method are discussed in section 2.4.2.3. It has to be mentioned that hardly any study uses Radar altimeter (Herzfeld *et al.*, 1994) data for detecting the grounding line which might be due to the fact, that the distance between adjacent along-track RA measurements is about 350 m and the footprint of each measurement is several kilometres. Further, the number of available RA measurements over rougher sloping margin areas is very limited. Conversely, ICESat laser altimeter data have a high along-track resolution (50 to 70m diameter footprints every ,172 m) and a per-shot accuracy of ,14 cm; this accuracy and resolution enables ICESat to detect the tide-induced difference in elevation from one repeat pass over ice shelves (Fricker *et al.*, 2009).

5.3.2.4 **Capabilities and limitations**

As with InSAR, the comparison of repeat-tracks of altimetry data can directly measure the tidal flexure in the grounding zone at the surface of the glacier tongue or ice sheet. The general capabilities and limitations of the repeat-track approach are described in section 2.4.2.4. Similar to InSAR, additional calculations and assumptions are required to infer the actual position of the grounding line from the detected band of tidal flexure (see section 5.3.1.4 for more details).

5.3.3 **Optical**

5.3.3.1 **Principle of method**

Another approach based on optical data aims on detecting surface topography changes from single (Scambos *et al.*, 2007) or repeat images (Bindschadler *et al.*, 2010) based on changes in surface shading. This approach takes advantage of a break in surface slope close to the location of the grounding line (I in Figure 5.1). This surface feature is caused by the abrupt change in basal stress at the grounding line associated with the transition from ice/bedrock to ice/water at the base of the glacier or ice sheet. This leads dynamic thinning of the floating ice and therefore a lowering of the surface (e.g. Vaughan, 1995).

5.3.3.2 **Input / Output**

Bindschadler *et al.* (2010) uses image differencing on medium (MODIS, bands 1 and 2) to high resolution (Landsat-7 ERM+, bands 4 and 8; EO-1 ALI, bands 4P and 8). The data were selected based on similar solar illumination properties, co-registered and visually inspected for cloud coverage. By differencing the nearly identical images, subtle slope changes are highlighted.

Scambos *et al.*, derive a mosaic of Antarctica from MODIS bands 1 and 2 acquired during the austral summer 2003-2004. Data were combined following a cumulation scheme and the component images were de-striped, geo-registered, and resampled. They generate a surface morphology composite by multiple filter operations as well as stacking and averaging based on the MODIS band 1 data.

5.3.3.3 **Accuracy and performance**

The break in slope is only a proxy for the grounding line and the association of this feature with the grounding line is not robust. Fricker & Padman (2006) showed that the break in slope can deviate from the actual grounding line by several kilometres. Further, they showed that nearby break in slope features are misleading and might wrongly be associated with the grounding line.

5.3.3.4 **Capabilities and limitations**

While InSAR and altimetry can directly measure the tidal motion at the grounding zone, the approach with optical data aims on finding the break in surface slope (I in Figure 5.1) associated with the grounding zone which is only an indirect measure for the location of the grounding line.

5.4 **Round Robin conclusions**

InSAR provides an excellent tool for directly observing the tidal motion of a marine terminating outlet glacier and the main goal of the round robin experiment was the comparison of extracted grounding zones provided by Round Robin participants. However, no external results were received. Therefore, we performed a qualitative comparison with published results over Petermann glacier in Northern Greenland we tried to reproduce using the same input data (Rignot, 1996 & 1998a). It has to be mentioned that Rignot (1996) additionally applied an elastic beam model and tidal predictions in order to derive the actual grounding line. However, InSAR can only deliver the surface expression of the so-called grounding zone (upper and lower limit of tidally induced flexure) without additional modelling and assumptions. No additional modelling efforts will be undertaken within the Ice_Sheets_cci project, therefore only the upper and lower limit of tidally induced flexure will be delivered in the ECV product.
Despite the difference in the final parameter (grounding line vs. upper and lower limit of tidal flexure) the qualitative comparison of the Ice_Sheets_cci consortium results with the published grounding line position showed a good agreement.

Only InSAR processing will be considered for GLL retrieval within the Ice_Sheets_cci project. The InSAR processing steps for retrieving the grounding line are similar to the method that has been tested and validated for in the Round Robin experiment on IV.

5.5 References


6 Gravimetric Mass Balance

6.1 Introduction

The Gravity Recovery and Climate Experiment (GRACE) (Tapley et al., 2004) satellite mission launched in 2002 allows fluctuations in ice-sheet mass to be estimated through measurement of their changing gravitational attraction. Advantages of the GRACE method are that it provides regional averages without the need for interpolation, measures the effect of mass fluctuations directly, and permits monthly temporal sampling. However, a key challenge is to discriminate fluctuations in ice-sheet mass from changes in the underlying crust and mantle. The spatial resolution of GRACE observations derived from global spherical harmonic solutions of about 300 km in the Polar Regions is coarse in comparison to that of other geodetic techniques. Hence, a further complicating factor is that signals may leak into regional GRACE solutions as a consequence of remote geophysical processes.

In 2005, Velicogna and Wahr (2005) showed for the first time the possibility of using data from the GRACE mission to determine the mass balance of the Greenland ice sheet. Since then many mass balance estimates of both Greenland and Antarctica have been published, both on ice sheet scale (Chen et al., 2006b; Ramillien et al., 2006; Forsberg and Reeh, 2007; Barletta et al., 2008; Velicogna, 2009; Shepherd et al. 2012) and drainage basin scale (Luthcke et al., 2006; Wouters et al., 2008; Schrama and Wouters, 2011; Sasgen et al., 2012; Barletta et al. 2013).

6.2 Review of scientific background

The GRACE mission has two identical spacecrafts flying about 220 km apart in a polar originally at 500 km above the Earth. GRACE maps the Earth's gravity fields by making accurate measurements of the distance between the two satellites, using GPS and a microwave ranging system.

The gravity field of a body depends on its mass and shape. For a spherical and uniform body, the gravity field is uniform in any direction. However the mass distribution of the Earth is irregular and 'lumpy'. The Earth's mantle flows to drive tectonic plate motion, enormous quantities of water are exchanged between the ocean and land, and atmospheric masses are also in continuous movement.

While the satellites move through this irregular gravity field, the orbits of each satellite are slightly disturbed, and this in turns affects the distance between the two spacecraft. GRACE's uniquely precise microwave ranging system measures changes in the approximately 220 km distance between the satellites with an accuracy of some microns. Moreover the satellites use the GPS system to determine precisely where and when the measurements were taken. The ultra-precise measurement taken by GRACE, combined with tracking data from the GPS satellites, allows us to map the Earth's gravity field with unprecedented accuracy.

GRACE-derived solutions of the Earth's time variable gravity field are available from different processing facilities like CSR, GFZ or JPL. Level-2 products, provided as spherical harmonic coefficients (Stokes coefficients), are widely used in mass change studies. Only a few approaches are directly based on Level-1B data (Luthcke et al., 2013). Because of differing processing strategies and background models, the comparison of various GRACE solutions reveals non-negligible differences.

With a typical temporal resolution of one month, GRACE Level-2 products allow the investigation of seasonal and inter-annual variations in addition to long-term changes (Horwath et al., 2012). Since small-scale gravity changes are attenuated at satellite altitude, the measurements are unable to resolve mass changes of high spatial resolution. The limited spatial resolution complicates the separation of mass changes from adjacent regions (e.g. neighbouring drainage basins or between ice sheet and ocean) and is the main reason for an error source termed leakage error.

Monthly GRACE solutions are heavily affected by correlated errors. Manifesting themselves in terms of north-south oriented stripes in the spatial domain, they have to be taken into account during the inference of mass change. Various filter approaches have been proposed for minimizing GRACE error effects. These approaches comprise simply Gaussian average methods (e.g. Swenson and Wahr, 2002) as well as more elaborated approaches accounting for the error characteristics (e.g. Swenson and Wahr, 2006; Kusche et al., 2009).

While correlated errors are predominantly affecting the short wavelength (i.e. coefficients of higher spherical harmonic degrees), coefficients of low degrees are also subject to errors. Due to ocean tide aliasing C20 can only be imprecisely resolved from GRACE observations. To overcome this limitation C20 is often replaced by an estimate derived from SLR observations (Cheng et al., 2013). GRACE is completely insensitive to other components of the signal spectrum. Gravity fields derived from GRACE data refer to the Earth's centre of mass (CM), where the combined degree one effect of surface mass changes and deformations of the solid Earth is zero by definition. To avoid the omission of degree one information on
surface mass changes external data sets, like from a combination of GRACE and numerical ocean model output (Swenson et al., 2008), have to be incorporated.

The integrative character of GRACE observations does not allow the vertical separation between super-imposed mass changes. External information is needed to distinguish between ice mass changes and solid Earth mass redistributions caused by Glacial Isostatic Adjustment (GIA). Deficiencies in GIA models add another potential source of errors (see annex CCI CECR document).

6.3 Algorithms

Methods used for the inference of mass changes from GRACE data is divided into two main groups:

1. Inversion approaches

2. Regional integration approaches

The mass inversion method has been adopted for the GMB product generation, primarily in order to be consistent with the ongoing GRACE result release at the national Danish Polarportal (www.polarportal.dk), but also because the mascon inversion approach make it more easy to separate overlapping signals ("leakage") from adjacent ice caps. The inversion method has been previously consolidated and validated in intercomparison exercises (e.g. Shepherd et al. 2012), and give similar results to regional integration approaches (as e.g. used in the Antarctica_CCI, where there are less leakage problems), when similar periods and level-2 data are analyzed.

6.3.1 Mass inversion method

The direct point mass inversion method used here for determining the monthly mass changes from the monthly GRACE data is based on Forsberg and Reeh (2007), Sørensen and Forsberg (2010) and Barletta et al. (2013). Prior to the inversion, corrections (prescribed GIA, C20 and degree one) and filtering (KK3 from Kusche et al., 2009) are applied to the GRACE data. The elastic response of the solid Earth to present-day ice mass changes involves changes in the gravity field. This must be removed from the GRACE data before deriving the surface mass changes from the observed gravity changes.

The elastic corrections based on the model of Farrel (1972), as also presented by Wahr et al. (1998), are used to make “reduced” gravity disturbances for the observation equations in the inversion.

The inversion is performed on a set of $N_r$ observations $\mathbf{y} = \{\delta g_k\}$, $k = 1 \ldots N_r$, i.e. gravity disturbances at the altitude of the satellites located at coordinates $(\theta_k, \phi_k)$, and solved for a point-like mass ensemble $\mathbf{x} = \{m_j\}$, $j = 1 \ldots N_s$ located on the surface of the Earth at coordinates $(\theta_j, \phi_j)$ which define the solution area (SA).

The linear problem $\mathbf{y} = \mathbf{A} \mathbf{x}$ is solved using a generalized least squares inversion with Tikhonov regularization $\mathbf{x} = (\mathbf{A}^T \mathbf{A} + \lambda I)^{-1} \mathbf{A}^T \mathbf{y}$, where $\lambda$ is a smoothing parameter and the observation matrix $\mathbf{A}$ is built upon the attraction of a point mass of the sphere to the measured gravitational attraction at the orbit level by

$$\delta g_k = G \frac{m_j a^2 [(h+a) - a \cos \psi_{kj}]}{r_{kj}^3}$$

so the elements of the matrix $\mathbf{A}$ are

$$A_{kj} = G \frac{a^2 [(h+a) - a \cos \psi_{kj}]}{r_{kj}^3}$$

where $G$ is the gravity Newton constant, $a$ is the mean radius of the Earth, $h$ is the height of the observation, and $r_{kj}$ and $\psi_{kj}$ are the distance and the angle, respectively, between the observation $\delta g_k$ at $(\theta_k, \phi_k)$ and the solution point $m_j$ at $(\theta_j, \phi_j)$. Once we have the observations $\{\delta g_k\}$ from GRACE, we build the matrix $\mathbf{A}$ and we find $m_j$ by solving the positive definite symmetric linear equations using Cholesky decomposition.

The inversion method has been refined, optimized and calibrated for the work presented in Barletta et al. 2013, and also used in the IMBIE comparisons (Shepherd et al. 2012). The solution area is optimized by using icosahedron-based grids (Tegmark, 1996) with disk elements of almost equal area. The solution area is calibrated in order to reduce as much as possible the leakage from the signal outside the region of interest. The inversion method assumes that the gravity signal is negligible outside the region of interest; otherwise such a signal would be forced to be part of the ice mass changes in the solution area. Once this assumption is verified by applying the method on synthetic data, we find that it is able to recover up to 99% of the mass (supplementary material in Barletta et al. 2013).

One strategy for mitigating the effect of ocean mass changes being erroneously modelled as ice sheet changes is to force the ocean signal to be zero, especially in the far field. The signal outside the region of
interest that is farther than some hundreds of km (300 to 500) from the boundary of the solution area can be forced to zero (zero mask) without compromising the signal of interest.

Another strategy is to build a complementary solution area (CSA) around the primary one. The CSA is a belt around the original solution area, but separated by a gap of some hundreds of km (300 to 500), and it accounts for the signal outside the original solution area. We used a combination of the two above strategies, and we calibrated the gap for CSA and the distance for the zero mask. The parameters we chose within the calibration allow us to retrieve about 98% of different kinds of synthetic signals. The regularization parameter (the smoothing parameter) used for this study was also chosen after calibration using a synthetic data set. Note that we do account for the degree-1 also in our calibration process because it does alter the calibration.

6.3.2 Outer glaciers and the drainage basin definition

Figure 6.1 (left panel) shows the solution area for the mass inversion over and around the Greenland ice sheet, and the point-like mass units used as solution area for the inversion are also visible. The closest surrounding ice-covered areas (Ellesmere Island and the outer glaciers) are included in the solution area, to reduce leakage. In fact Ellesmere Island has a strong trend, and it is so close to the northwest Greenland that it cannot be treated as the rest of the surrounding sea and islands around Greenland. Once we obtain a mass grid from the inversion scheme for each month, we integrate over each basin’s area to derive the total mass change for Greenland (see RR annex). The mass estimate of each of the basins is obtained as the sum of the point mass changes within each basin mask definition (Figure 6.1 left panel). Each basin is colour-coded with dark violet being the surrounding glaciers and ice caps not connected to the ice sheet.

The basin definition (Figure 6.1 left panel) covers only the ice sheet, but the outer glaciers contribute to the signal, and therefore have to be taken into account too. In fact the outer glaciers (Figure 6.1 left panel) are so close to the ice sheet that they cannot be resolved by GRACE. If the glaciers and ice caps are not included in the solution area, their potential signal will be pushed into the ice sheet solution, and so we overestimate the GIS-only mass loss (265 Gt/yr for the whole ice sheet, Figure 6.1 middle panel). The inversion method tends to place more mass on the edge and therefore if the outer glaciers are included in the solution area, part of the GIS signal will be pushed in the outer glaciers area. So if their contribution is not included in the total mass balance computation, the GIS mass loss is likely instead underestimated (237 Gt/yr for the whole ice sheet). This problem of leakage is inherent in all GRACE estimation methods, either inversion or regional integration approaches.

For these reasons, using GRACE data only we get a more accurate estimate if we compute the mass balance over the whole Greenland of -265 Gt/yr (Ice sheet and glaciers and ice caps), that is including the outer glaciers (Figure 6.1 right panel). We choose to use this estimate for the whole GIS mass balance.

For the solution over basins which are defined only over the ice sheet we cannot get an accurate solution. In the annex CCI CECR document we explain in details how we choose to provide the leakage error.
6.3.3 Regional integration approach

There are different implementations of the regional integration approach. In general in the regional integration approach (Swenson and Wahr, 2002; Horwath and Dietrich, 2009) the mass change for a certain region \( k \) is estimated as integral over a region of the surface mass density derived from the GRACE solutions \( \Delta m_{i,m} \) with a straightforward formula as shown in Wahr et al. (1998)

Before the GRACE monthly solutions \( \Delta m_{i,m} \) can be used within a regional integration approach, inherent error effects need to be reduced by means of an appropriate filter. Usually the filtering causes an additional smoothing and attenuation of the signal leading to increased leakage. Several leakage corrections have been proposed. For example, the derived mass change can be rescaled using a scaling factor derived by means of a synthetic high-resolution data set, e.g. a geophysical model simulating the mass change under investigation. After limiting the synthetic data set to the spatial resolution of the GRACE data and applying the filter, the scaling factor can be derived by comparing the original and the processed synthetic data (Landerer and Swenson, 2012).

Instead of performing these processing steps individually a tailored sensitivity kernel \( \eta(\lambda, \varphi) \) can be designed, providing a trade-off between the minimisation of both GRACE error effects and leakage errors (Groh & Horwath, 2016) as described in the AIS CCI ATBD document. For the Greenland drainage basins, denoted by indices \( k = 1, \ldots, K \), a coherent set of sensitivity kernels \( \eta_k(\lambda, \varphi) \) \((k = 1, \ldots, K)\) is constructed. The sensitivity kernel for any region needs to be chosen as a compromise between the following conflicting conditions:

(A) Mass changes inside the region are correctly recovered

(B) Mass changes outside the region have zero influence on the regional mass change estimate

(C) Propagated errors of the GRACE solutions have small influence on the estimate.

Conditions (A) and (B) are to minimize leakage effects, while condition (C) is to minimize GRACE error effects. Tailoring of the sensitivity functions is accomplished in a formal least square adjustment procedure. The parameters to be adjusted are the spherical harmonic coefficients of the sensitivity functions, \( \eta_{n,m} \). To control leakage, condition equations on the set of coefficients \( \eta_{n,m} \) need to be established for a large number of mass change patterns, where different weights can be controlled by scaling the mass patterns.

To establish a condition for (C) (control on propagated GRACE error effects), an error variance/covariance model for the GRACE monthly solutions, expressed as a variance/covariance matrix \( C \), is needed. An error variance/covariance model may be derived empirically, e.g. based on the short-term month-to-month scatter of the monthly solutions or it may be provided together with the GRACE monthly solutions.

6.4 Round Robin conclusions – selection of algorithm

The algorithm selection is based on all Round Robin contributions which include both GRACE-derived mass change products and results from prescribed synthetic data sets. It was shown that mass change time series on basin scale from the mass inversion and the regional integration approach using tailored sensitivity kernel and from a forward modelling technique are in good agreement. This agreement comprises the noise level of the time series as well as the observed temporal changes. Both algorithms were able to recover synthetic mass changes on a comparable level. The mass inversion approach has been selected for the generation of GMB products. A detailed summary of the Round Robin results is given in the technical note annexed to this document.

6.5 GMB estimation process

The GMB system is a semi-automatic processing system implemented in Fortran and gnuplot. Figure 6.2 gives a schematic overview of the GMB processing line, which consists of two main modules: the pre-processing and processing modules. Input data download is done manually. In the following, the two modules of the GMB processor are briefly described.

Pre-processing: the GRACE monthly solutions and time series of auxiliary datasets are pre-processed, combined and corrected. This includes filtering/destriping of the Stokes-coefficients, the addition of degree-one coefficients, replacement of degree C20, and the reduction of GIA. All monthly solutions are reduced to a specified reference value. The solution area grid and the basin definition is chosen as described in section 6.3.2.
### Processing (Mass change estimation): Mass change time series for every grid cell of the GMB gridded product and every basin of the GMB basin product are produced by integrating the product of the point mass inversion (Barletta et al., 2013) of each monthly as detailed in section 6.3.1. The statistical error characterization of the time series is derived. Basin averaged time series are used to calculate a mass balance estimate for each basin.

![Flow chart of the GMB estimation process](image)

**Figure 6.2:** Flow chart of the GMB estimation process

### 6.6 Selection of Level-2 data source for the CCI product

We have for the CCI GRACE product selected to use a new 2016 L2 version product from ITSG, TU Graz (Mayer-Gürr et al., 2014). The new ITSG2016 product is unconstrained, as a number of updates in fundamental updates in reference syste, and processing, and forms the core of a new European Gravity Emergency Service project (EGSIEM), where the goal is to make new, even more accurate combined GRACE products. Comparison to results from other L2-processing centers (e.g. CSR-R5) show only minor differences in the derived trends. The ITSG forms also the base for similar GMB product for the Antarctica_CCI project.

### 6.7 References

Barletta, V. R., Sørensen, L. S., and Forsberg, R. (2013). Scatter of mass changes estimates at basin scale for Greenland and Antarctica. The Cryosphere, 7(5), 1411–1432.


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