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**Estimates of glacier ice discharge to the ocean
combining synthetic-aperture radar-derived
velocities and ground-penetrating radar-derived ice
thickness. Applications to Arctic Glaciers**

Tesis Doctoral

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I would like to dedicate this thesis to my family, especially to my parents. Without their support this endeavor would not have been possible to tackle.

Declaration

I hereby declare that except where specific reference is made to the work of others, the contents of this dissertation are original and have not been submitted in whole or in part for consideration for any other degree or qualification in this, or any other University. This dissertation is my own work and contains nothing which is the outcome of work done in collaboration with others.

Pablo Sánchez-Gómez
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I would like to acknowledge the special contribution of Prof. Francisco J. Navarro to my research. His advice and expertise was an invaluable asset that helped to bring this effort to a successful ending.

Abstract

Ice discharge to the ocean is an important component of the mass balance of tidewater glaciers and marine-terminating ice caps. In this thesis we develop methodologies to improve the ice discharge calculations from remotely-sensed glacier velocities and radar-retrieved ice-thickness data, and apply them to provide updated estimates of ice discharge for Canadian High Arctic glaciers and the Academy of Sciences Ice Cap in Severnaya Zemlya, Russian Arctic.

Following an overview of the state-of-art methodologies to retrieve glacier surface velocity fields from remotely-sensed data, and presenting the basics of Synthetic Aperture Radar (SAR) data processing in chapters 1 and 2, the core results of this thesis are presented in chapters 3, 4 and 5, and finally the conclusions and outlook are summarized in Chapter 6.

Focusing on the core of the thesis, we firstly provide an improvement of the intensity offset tracking methodology for estimating glacier surface velocities. We optimise the offset tracking technique by omitting the azimuth offsets, using instead only range offsets from ascending and descending passes. By doing so, we are able to improve the final resolution of the velocity product, as the Terrain Observation by Progressive Scans (TOPS) acquisition mode of the Sentinel-1 mission provides resolutions of 5 m in range and 20 m in azimuth. Simultaneously, we avoid the undesired ionospheric effect manifested in the data as azimuth streaks. We apply the developed technique to retrieve glacier-surface velocities from the southern Ellesmere Island ice caps, Canadian High Arctic. We additionally use Differential Interferometric Synthetic Aperture Radar (D-InSAR) techniques, and show that the latter shows its merits when applied to slow-moving areas, while offset tracking is more suitable for fast-moving areas. Both methods are thus complementary, and the use of both to determine glacier velocities is better than only using one or the other. We observe glacier surface velocities of up to 1200 m year^{-1} for the fastest tidewater glaciers. The land-terminating glaciers show typical velocities between 12 and 33 m year^{-1} , though with peaks up to 150 m year^{-1} in narrowing zones of the confining valleys.

Secondly, we analyse the various error sources in the estimation of ice discharge through flux gates, distinguishing the cases with ice-thickness data available for glacier cross-sections or only along the centreline. For the latter, we analyse the performance of three different

U-shaped cross-sectional approaches. We apply this methodology to glaciers of the Canadian High Arctic. The velocity field is the main error source for small and medium-size glaciers (discharge $<100 \text{ Mt a}^{-1}$) with low velocities ($<100 \text{ m a}^{-1}$), while for large glaciers (discharge $>100 \text{ Mt a}^{-1}$) with high velocities ($>100 \text{ m a}^{-1}$) the error in cross-sectional area dominates. Thinning/thickening between ice thickness and velocity measurements should be considered, as it implies systematic errors up to 8% in our study. The U-shaped parabolic approach, which allows for an adjusted estimation when the ice-thickness measurement point is displaced from the glacier centreline, performs best, with small bias and admissible standard error. We observe an increase of ice discharge from the main glaciers (Trinity and Wykeham) of the Prince of Wales Icefield from 2015 to 2016, by 5% and 20%, respectively, followed by a decrease in 2017, by 10% and 15% respectively. Belcher Glacier, of the Devon Ice Cap, maintains similar discharges during 2015-2017.

Thirdly, we apply the developed methodologies, together with other state-of-art techniques, to the investigation of the dynamics and mass balance of the Academy of Sciences Ice Cap in the Russian Arctic, analysing its seasonal and intra-annual, as well as inter-annual, variations of velocity and ice discharge. We also analyse the contributions to the total mass balance of the ice cap of surface mass balance and frontal ablation (approximated here by the calving determined as ice discharge through flux gates close to the calving fronts), and the partitioning of total ablation into surface ablation and frontal ablation. With these aims, we process, using feature tracking, 54 pairs of Sentinel-1 synthetic-aperture radar images of the Academy of Sciences Ice Cap, acquired from November 2016 to November 2017. Seasonal velocity variations up to 10% (20% peak-to-peak) of the yearly-averaged velocity are observed. Shorter-term intra-annual velocity variations have average deviations up to 16% and maximum up to 32% (32% and 64% peak-to-peak). This gives an indication of the errors incurred when extrapolating to the whole year discharge values determined from a single pair of SAR images. Average ice discharge for 2016-2017 was $1.93 \pm 0.12 \text{ Gt a}^{-1}$. The difference from an estimate of $\sim 1.4 \text{ Gt a}^{-1}$ for 2003-2009 is attributed to the initiation of ice stream flow in a southern basin. The total geodetic mass balance for the ice cap over 2012-2016 is $-1.72 \pm 0.67 \text{ Gt a}^{-1}$ ($-0.31 \pm 0.12 \text{ m w.e. a}^{-1}$). The climatic mass balance is not significantly different from zero, at $0.21 \pm 0.68 \text{ Gt a}^{-1}$ ($0.04 \pm 0.12 \text{ m w.e. a}^{-1}$), and has remained at this level for the last four decades, so the total mass balance is governed by the variations in ice discharge, whose long-term changes do not appear to respond to environmental changes but to the intrinsic characteristics of the ice cap.

Resumen

La descarga de hielo glaciar al océano es una componente importante del balance de masas de glaciares y casquetes de hielo terminados en mar. En esta tesis desarrollamos metodologías para mejorar el cálculo de la descarga glaciar empleando velocidades del hielo determinadas empleando sensores remotos y espesores glaciares obtenidos empleando radar. Usando estas metodologías realizamos estimaciones actualizadas de descarga glaciar para el Ártico Canadiense y el casquete glaciar de la Academia de Ciencias situado en Sévernaya Zemliá, en el Ártico Ruso.

En los capítulos 1 y 2 presentamos una visión de conjunto del estado del arte sobre las metodologías para el cálculo de campos de velocidad superficial empleando datos recogidos por sensores remotos, y una introducción al procesamiento de datos de Radar de Apertura Sintética (SAR). A continuación, los resultados fundamentales de la tesis se presentan en los capítulos 3, 4 y 5, y, finalmente, las conclusiones y las perspectivas de futuro se resumen en el capítulo 6.

En el capítulo 3 se presenta una mejora del método de estimación de los desplazamientos empleando intensidades de la señal SAR para calcular la velocidad en la superficie de los glaciares. La idea fundamental es la optimización de la técnica para estimación de desplazamientos conocida como offset tracking omitiendo el empleo de los offsets en la dirección de desplazamiento del satélite, empleando en su lugar únicamente offsets en rango en las direcciones tanto ascendente como descendente del satélite. Con esto se logra mejora la resolución del campo de velocidades del glaciar, ya que la técnica de adquisición conocida como Observación del Terreno mediante Escaneos Progresivos (TOPS) del satélite Sentinel-1 proporciona una resolución asimétrica, de 5 metros en rango y de 20 metros en la dirección de desplazamiento. Simultáneamente, se evita el efecto ionosférico adverso que se manifiesta en los datos finas franjas de ruido perpendiculares a la dirección del azimut. Esta técnica se aplica en la tesis para calcular las velocidades superficiales de los casquetes glaciares situados en la Islas Ellesmere del Ártico Canadiense. Adicionalmente, se emplea la técnica de Interferometría Diferencial del Radar de Apertura Sintética (D-InSAR), mostrándose que con esta técnica se logran mejores resultados en zonas de poco movimiento, mientras que la metodología de seguimiento por offsets es más adecuada para zonas con grandes

desplazamientos. De esta forma, ambas metodologías resultan ser complementarias, y el empleo de ambas para la estimación de las deformaciones en la superficie glaciar es mejor que el uso de una sola técnica. Se observan velocidades de hasta 1200 m año⁻¹ para algunos de los glaciares más rápidos. Los glaciares terminados en tierra muestran velocidades típicas comprendidas entre 12 y 33 m año⁻¹, aunque se observan picos de hasta 150 m año⁻¹ en los estrechamientos de los valles de confinamiento.

En el capítulo 4 se analizan las fuentes de error en la estimación de la descarga glaciar a través de puertas de flujo, distinguiendo el caso en el que se dispone de datos de espesores para secciones transversales del glaciar de aquél en el que únicamente se cuenta con datos de espesores a lo largo de la línea central del glaciar. Para este último caso, se analizan tres aproximaciones a la sección transversal del glaciar. Esta metodología se aplica a glaciares del Ártico Canadiense. Los resultados del análisis muestran que el campo de velocidades es la mayor fuente de error para glaciares de tamaños pequeño y mediano (descarga <100 Mt año⁻¹) con bajas velocidades (<100 m año⁻¹), mientras que para glaciares de gran tamaño (descarga > 100 Mt año⁻¹) con altas velocidades (>100 m año⁻¹) el error en el área de la sección transversal es el factor dominante. El posible engrosamiento/adelgazamiento del glaciar producido entre la observación de los espesores y las mediciones de velocidad debe ser tenido en cuenta, ya que implica errores sistemáticos de hasta el 8% en nuestro estudio. La aproximación parabólica de la sección transversal, que permite una estimación ajustada cuando la medición del espesor se encuentra desplazada del centro del glaciar, presenta los mejores resultados, con un pequeño sesgo y una desviación estándar admisible. Se observa un incremento en la descarga para los principales glaciares (Trinity y Wykeham) del campo de hielo Price of Wales entre 2015 y 2016, de un 5% y un 20%, respectivamente, seguido de un descenso en 2017, de un 10% y un 15%, respectivamente. El Glaciar Belcher, en el casquete de hielo Devon, mantiene descargas similares durante el período 2015-2017.

Finalmente, en el capítulo 5, las citadas metodologías se aplican, junto con otras técnicas innovadoras, en la investigación de la dinámica y el balance de masas del casquete de hielo de la Academia de las Ciencias en el Ártico Ruso, analizando sus variaciones intra e interanuales en velocidad y descarga. También se analizan las contribuciones del balance de masa en superficie y de la ablación frontal al balance de masa total del casquete, y las contribuciones de la ablación en superficie y la ablación frontal a la ablación total. Con estos objetivos, se procesan, empleando la metodología de offset tracking, 54 pares de imágenes Sentinel-1 del casquete, tomadas entre noviembre de 2016 y noviembre de 2017. Las variaciones de velocidad estacionales suponen hasta un 10% (20% entre máximos) de la velocidad media anual observada. Las variaciones intraanuales de corto período tienen desviaciones promedio de hasta el 16% y máximas de hasta el 32% (32% y 64% entre máximos). Esto es

indicativo de los errores que se cometerían extrapolando a todo el año los valores de descarga calculados usando un único par de imágenes. La descarga glaciár promedio para el período 2016-2017 es de $1.93 \pm 0.12 \text{ Gt año}^{-1}$. La diferencia con la estimación de 1.4 Gt año^{-1} para el período 2003-2009 se atribuye al comienzo del flujo en la cuenca glaciár sur. El balance de masa geodésico para el casquete en el período 2012-2016 es de $-1.72 \pm 0.67 \text{ Gt año}^{-1}$ ($-0.31 \pm 0.12 \text{ m w.e. año}^{-1}$) y se ha mantenido a este mismo nivel durante las últimas cuatro décadas. Por lo tanto, el balance de masas total está gobernado por las variaciones en descarga glaciár, cuyos cambios de largo período no parecen responder a cambios climáticos sino a las características intrínsecas del casquete.

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Chapter 1

Introduction

1.1 Motivation

Climate change and its impact on natural and anthropogenic systems is becoming a matter of concern for society and policy makers (IPCC, 2013). Global mean sea-level rise is expected to range between 0.24 and 0.30 metres for the period 2046-2065 and between 0.40 and 0.63 for the period 2081-2100, depending on which of the four Representative Concentration Pathways (RPCs) defined in the IPCC 2013 Report is considered (IPCC, 2013). The main current contributors to sea-level rise are the thermal expansion of the oceans, the mass losses from glaciers and ice caps (henceforth, glaciers) and the ice-sheet mass losses (IPCC, 2013). However, the relative shares of each contributor to sea-level rise varies among studies. The latest research shows a similar contribution to sea-level rise of ice-sheets and glaciers ($\sim 25\%$) while ocean thermal expansion remains as the main source ($\sim 35\%$) (Gardner et al., 2013; Hanna et al., 2013; IPCC, 2013).

Considering the influence of glacier mass loss in sea-level rise and its importance under all future RPCs scenarios, it is fundamental to constrain the shares of surface ablation and frontal ablation to total mass loss as a necessary step for understanding glacier dynamics and how climate change could have an influence on these systems (Melkonian et al., 2016; Osmanoğlu et al., 2013, 2014; Sánchez-Gómez et al., 2018). The climatic mass balance comprises both the surface and internal accumulation and ablation processes (Cogley et al., 2011), but the internal ones, difficult to estimate but assumed much lower than those at surface, are often neglected. In such case, the climatic mass balance reduces to the surface mass balance, which encompasses surface accumulation and surface ablation. Frontal ablation is defined as ice mass losses by calving, subaerial frontal melting and sublimation, and subaqueous frontal melting at the nearly-vertical calving fronts (Cogley et al., 2011). In general, subaerial frontal melting and sublimation are assumed to be small, and, in the case of the near-vertical fronts

of tidewater glaciers, the mass loss by subaqueous frontal melting is assumed to be much lower than that by calving. Consequently, frontal ablation can be considered in our study as nearly equivalent to calving flux. Ice discharge can be computed through any given flux gate in a glacier. However, when the considered flux gate is close to the calving front of a tidewater glacier, ice discharge and calving flux (the latter being defined only at the glacier terminus) are equivalent, except that the latter comprises also the changes in mass due to the possible variations of the calving front position (either advance or retreat).

Frontal ablation is an important component of the mass balance of tidewater glaciers and marine-terminating ice caps (Huss and Hock, 2015). It accounts for up to 30-40% of the total ablation of some Arctic glaciers (Błaszczyk et al., 2009; Dowdeswell et al., 2002, 2008) and up to 50% in some ice caps in the Antarctic periphery (Osmanoğlu et al., 2014). We focus our study on the Arctic glaciers and ice caps. Specifically on the Canadian High Arctic (Sánchez-Gómez and Navarro, 2017, 2018) and the Russian Arctic (Sánchez-Gómez et al., 2018). Recent research has determined that these regions are affected by climate change (Gardner et al., 2013, 2011). The Canadian High Arctic glaciers and ice caps have shown an unprecedented increase in mass loss during the last decade (Gardner et al., 2011). Novaya Zemlya archipelago glaciers have been also affected by climate change, in some cases with signs of retreat and thinning at tidewater glacier termini (Carr et al., 2014, 2013; Melkonian et al., 2016). The rest of the Russian Arctic, namely Severnaya Zemlya and Franz Josef Land, are currently losing little mass. However, they are projected to have increased mass loss rates to the end of the 21st century (Huss and Hock, 2015). These reasons motivated our interest to deepening the knowledge of the mass losses by frontal ablation in both of these regions. In the case of the Russian Arctic, we focused on the Academy of Sciences Ice Cap, located in Severnaya Zemlya, because it is one of the largest ice caps in the whole Arctic region.

Flux across a given surface is calculated as the surface integral, over the entire cross-sectional area, of the inner product of the ice velocity field and the unit vector surface area (element of surface times the unit vector normal to the surface element). In the case of ice flux, the velocity field is determined from the velocity observed at the glacier surface, and the glacier cross-section from the ice thickness determined using, most often, ground-penetrating radar (GPR) data. The current methodologies for estimating the ice surface velocity field rely on the use of either optical or Synthetic Aperture Radar (SAR) sensors (Strozzi et al., 2002). The latter, being an active sensor, is better suited for carrying out continuous analysis of the cryosphere. The use of optical imagery for estimating glacier surface motion is also widespread (Heid and Kääb, 2012; Kääb et al., 2005). However, the presence of clouds makes this methodology less suited for continuous analysis. The possibility of a combination

of both types of surface velocities is outlined in the Global Land Ice Measurements from Space (GLIMS) textbook (Kargel et al., 2014).

The first aim of this thesis is to develop an algorithm able to provide an improved glacier surface velocity field calculated using the intensity offset tracking methodology (Strozzi et al., 2002). We will apply this methodology to Sentinel-1 SAR Single Look Complex (SLC) Interferometric Wide-swath level-1 product, with a resolution of 5 and 20 metres in the across- and along-track directions, respectively (Nagler et al., 2015). Typically, the estimated ice surface velocity field from SAR acquisitions suffers from the presence of azimuth streaks due to the influence of the ionospheric effect on the retrieved signal (Gray et al., 2000). Furthermore, the resolution of the SAR Terrain Observation by Progressive Scans (TOPS) image on the along-track direction is coarser than in the across-track direction (Zan and Guarnieri, 2006). The possibility of using the ascending and descending acquisitions for the Sentinel-1 mission, allows to extract the full vector of displacement only from the across-track offset tracking results (Fallourd et al., 2010). The latter brings up the opportunity of avoiding at the same time the undesired azimuth streaks and using the higher resolution across-track offsets (Sánchez-Gómez and Navarro, 2017).

Global-scale assessments of mass change of the glaciers outside the ice sheets did not include frontal ablation up to IPCC 2007. This is due to the inherent difficulty of assessing and modelling of calving (Benn et al., 2007a,b) and submarine melt (Enderlin and Howat, 2013). The processes involved in frontal ablation are largely non-linear and operate on time scales that are not necessarily linked to regional climate variations (Truffer and Fahnestock, 2007). There is a need to better quantify the dynamic mass losses because they provide a mechanism for glaciers to loss mass much more rapidly than is possible through other means.

There are plenty of ice discharge estimations for the different Arctic regions (Melkonian et al., 2016; Moholdt et al., 2012a; van Wychen et al., 2012; Williamson et al., 2008). In all cases, the estimated values have an associated uncertainty. However, in most cases, this uncertainty estimation is very rough and lacks a proper analysis on how the various error components are propagated into the total error. Gardner et al. (2018) applied error propagation analysis to their Antarctic-wide ice discharge estimates. Nonetheless, for the case of tidewater glaciers, a similar study on how the error budget is distributed among the different error components and how the glacier characteristics affect this error distribution remained to be done. Therefore, the second aim of this thesis is to characterize the error budget estimation and its distribution among the different error sources for valley-confined tidewater glaciers (Sánchez-Gómez and Navarro, 2018). The knowledge of the accuracy of the ice discharge estimates provides the opportunity for better constraining the contribution of frontal ablation to total glacier mass loss.

Data regarding the partitioning of total glacier mass loss into its main components (surface mass balance and frontal ablation) are very scarce, but are needed to derive suitable models for frontal ablation on larger scales and also to calibrate and validate these models. A few studies on marine-terminating ice caps in the Arctic show that frontal ablation might account for roughly 30-40% of the total ablation (Dowdeswell et al., 2002, 2008). Other studies in the Arctic region have also calculated the frontal ablation, but have presented their results as a percentage of the net losses (e.g. Burgess et al. (2005, 2013)). However, computing the share of frontal ablation to total ablation (which always is a mass loss) is very different from computing the share with respect to the net mass budget (that can be either gains or losses). The former approach requires that the partitioning of the budget between mass gains and losses is known, as happens with any method of the input-output type. In contrast, geodetic or gravimetric methods do not allow discriminating the individual components of the mass budget.

The combination of different remotely-sensed information from the same region allows to improve the knowledge of the different geophysical parameters of glaciers and ice caps such as the surface elevation change from ICESat laser altimetry, and the derived mass change (Moholdt et al., 2010, 2012b), or the ice surface velocities determined by image matching between repeat-pass optical satellite imagery (Moholdt et al., 2012a). We applied both types of techniques to the Academy of Sciences Ice Cap, located in Severnaya Zemlya, in the Russian Arctic. The final aim of applying both methodologies to an ice cap is to ascertain the shares of frontal ablation and surface ablation to total mass loss. The surface elevation change was estimated with the combination of ICESat laser altimetry data (Moholdt et al., 2010, 2012b) and inferred digital elevation models (DEMs) (Noh and Howat, 2015; Noh et al., 2016). Combining both sources of information, namely surface elevation changes and surface velocity for the same area, permits to split the total mass balance (estimated as geodetic mass balance) and frontal ablation. Their difference gives the surface mass balance and, provided that we count on an independent estimate of surface accumulation, we are able to infer surface ablation. In this way, we are able to estimate the shares of frontal ablation and surface ablation to total ablation. This technique, which we successfully applied to the Academy of Sciences Ice Cap, has a great potential to be applied in similar studies of other glacier regions.

This dissertation has been developed within the Group of Numerical Simulation in Science and Engineering (GSNCI, using its Spanish acronym) of Universidad Politécnica de Madrid (UPM), to which both the author and the supervisor of this thesis belong. The main research lines of GSNCI are numerical modelling of glacier dynamics modelling, including physical parameters estimation and inversion, ground-penetrating radar applications to glaciology,

aimed to determine ice thickness and physical properties of glacier ice, mass balance studies and, more recently, the application of satellite-based remote sensing techniques for the investigation of glacier dynamics and mass balance. This thesis was developed within the Spanish Research and Development project entitled "Mass discharge from glaciers to the ocean: improving current estimates and forecasting future contributions under a changing climate" (project CTM2014-56473-R). The first objective of this project was to "develop techniques allowing to improve the estimates of ice discharge from glaciers to the ocean, combining remote sensing techniques, surface mass-balance observations, modelling of glacier dynamics and ground-penetrating radar measurements". Therefore, the subject of this dissertation is closely linked to this project, and the results presented in this doctoral thesis have greatly contributed to achieve the main objectives of the mentioned research project.

This PhD thesis work has also been partly developed within the European Commission H2020 project "Integrated Arctic Observation System" (INTAROS, Research and Innovation Action 727890). In particular, within its Work Package 2, Task 2.2, where it is stated that "Methods to estimate ice sheet mass balance, mass changes, and surface velocities from in situ and from remote sensing data will be developed (UPM, DTU, GEUS, US)", and also as part of Work Package 6, Task 6.4, in which it is noted that "UPM, DTU and GEUS will demonstrate the integration of recently available satellite data from the Copernicus programme (Sentinel-1) with enhanced in situ observations of surface mass balance and mass loss and innovative modelling to deliver the contribution of mass losses from the Greenland ice sheet and Svalbard glaciers and ice caps to sea-level rise and a method to separate solid iceberg production from the total marine mass loss of selected glaciers".

1.2 State of the art

Calving flux, as an approach to frontal ablation, is calculated in the literature from the combined use of a surface velocity field and either an observed (Dowdeswell et al., 2002) or estimated ice-thickness field (Farinotti et al., 2009). The surface velocity field can be calculated using different techniques that have their own advantages and shortcomings. Among them, we can enumerate the following well-established methodologies (Kargel et al., 2014):

1. SAR interferometry (InSAR) and differential interferometry (D-InSAR).
2. InSAR applied to ascending and descending passes.
3. SAR multiple aperture interferometry (MAI).

4. SAR intensity offset tracking.
5. SAR coherence offset tracking.
6. Optical satellite imagery matching methods.

In the subsections that follow, we will briefly describe the above methodologies, and comment on their merits and drawbacks.

Regarding the ice-thickness field, observations are normally carried out using ground-penetrating radar (Navarro and Eisen, 2010). Nonetheless, ice-thickness observations are usually very costly to obtain, and their low spatial coverage and poor distribution is commonly a problem. Therefore, methods for inverting ice-thickness from several observable glacier parameters, such as average surface slope or surface velocities, together with mass balance and surface-elevation changes, have been developed in the last years (Farinotti et al., 2017, 2009). Furthermore, in the case of valley-confined tidewater glaciers, cross-sectional estimation of glacier thickness using the parabolic approach from single along-track GPR measurements have been similarly proposed (Harbor, 1992; van Wychen et al., 2012). We will not discuss these techniques here, though several of them will be dealt with in subsequent chapters (in particular, chapters 3-5).

1.2.1 SAR interferometry (InSAR) and differential interferometry (D-InSAR)

SAR interferometry is based on the use of a pair of images of the same area at different times or from different positions. The measurement of the phase shift between acquisitions at the same pixel permits estimating the displacement at that location when the topographic phase is known or the baseline between acquisitions is small (Goldstein et al., 1993), or estimating the terrain elevation when we assume no displacement between images (Fatland and Lingle, 1998). The phase shift signal between acquisitions is made up of the following components (Fatland and Lingle, 1998; Joughin et al., 1998):

$$\Psi_{total} = \Psi_{topography} + \Psi_{geometry} + \Psi_{displacement} + \Psi_{noise} . \quad (1.1)$$

where Ψ_{total} is the total observed phase shift, $\Psi_{topography}$ is the topographic related component, $\Psi_{geometry}$ is the acquisition geometric phase component, $\Psi_{displacement}$ is the ground-displacement component and Ψ_{noise} is the phase noise due to speckle and propagation delays caused by atmospheric variations between passes (Joughin et al., 1998):

$$\Psi_{geom} \approx \frac{2kB_n}{R \tan \alpha} , \quad (1.2)$$

$$\Psi_{topography} \approx \frac{2kB_n \Delta h}{R \sin \alpha} , \quad (1.3)$$

$$\Psi_{displacement} \approx 2k\Delta R . \quad (1.4)$$

where k is the radar wave number $2\pi / \lambda$, λ is the radar carrier wavelength, B_n is the baseline line-of-sight perpendicular component, R is the distance from the sensor to the scene center and α is the image-center radar-incidence angle. Δh is the topographic elevation of a cell relative to a reference elevation and ΔR is the radial distance change of a resolution cell due to ground-displacement (Fatland and Lingle, 1998).

The phase shift can be retrieved when both images are coregistered to a fraction of a resolution cell (Fatland and Lingle, 1998; Joughin et al., 1995). Once both acquisitions are coregistered the interferogram is calculated, but this signal encompasses the phase of the acquisition's geometry and topography (see Eq. 1.1). The intrinsic coherence of the interferogram is limited by the time span between image acquisitions when the scene has ground-displacement or the baseline between passes exceeds a specific threshold (Goldstein et al., 1993; Joughin et al., 1998). Therefore, when coherence between consecutive acquisitions is lost, interferometry can no longer be applied for retrieving ground displacements. This sets a temporal limitation for applying this methodology to two consecutive SAR acquisitions.

We can consider three general cases where interferometry can be applied. The first situation is when there is no ground displacement and, therefore, there is only phase shift due to the acquisition's geometry and topography (Fatland and Lingle, 1998). In this situation, we are only interested in retrieving the phase shift produced by surface elevation. Therefore, we need to isolate the topographic component of the phase. In order to achieve this goal, if no information or only coarse information is available on the baseline vector between satellite passes, the baseline needs to be estimated and further refined in a sequential fashion (Fatland and Lingle, 1998). Using the estimated baselines, the calculated geometric phase allows to isolate the remaining residual topographic phase and further unwrap it and process it for obtaining a digital elevation model. The resolution of the calculated topographic phase is on a scale of several metres (Zebker et al., 1994).

The second-case scenario appears when the baseline vector is very small or close to zero (this situation is very uncommon in reality). If this happens, the phase shift due to both

the acquisition's geometry and topography disappears, leaving only the component due to ground-displacement (see Eq. 1.1).

The last case is defined by a moving surface with topographic relief and a non-zero baseline. This represents the general situation for interferometric analysis (Fatland and Lingle, 1998). We assume that the geometric phase has been removed from the interferogram and the remaining phase is due to topography and ground-displacement. The methodology used for separating both components of the phase shift is called differential interferometry or D-InSAR (Fatland and Lingle, 1998; Vachon et al., 1996). There are two ways of separating both components of the phase, the first is using two pairs of coregistered images and obtaining two interferograms from them (the first with only the topographic phase and the second with the phase due to topography and displacement). The second approach makes the use of an external DEM, for subtracting the topographic phase from the total calculated phase shift (Fatland and Lingle, 1998; Vachon et al., 1996). The ground-displacement obtained in this way corresponds with the projection of the deformation on the radar line-of-sight. The resolution of this measurement is on the scale of a fraction of the sensor wavelength (Fatland and Lingle, 1998).

The main limitation of this methodology is retaining the coherence between images. Therefore, temporal decorrelation between acquisitions could prevent the application of this technique. Further limitations are those related with the differential displacement measurement, as the change in displacement between adjacent resolution cells must be less than $\lambda/2$ (Vachon et al., 1996). The latter sets a limit on the ability of resolving a large velocity gradient.

1.2.2 InSAR applied to ascending and descending passes

Ice-dynamics studies normally require the full three-dimensional surface velocity vector. The projected radar line-of-sight ground displacement obtained using the D-InSAR technique is an information that does not allow to infer the full 3D displacement vector. Hence the need for two further image acquisitions taken from different observation directions (Joughin et al., 1998). However, this is very restrictive, and it is advisable to be able to infer the full displacement vector from only two directions of observation.

$$\mathbf{v} = \mathbf{v}_h + [\nabla_{xy} z_t(x, y)]^T \mathbf{v}_h \hat{\mathbf{z}}. \quad (1.5)$$

Where $z_t(x, y)$ is the glacier surface elevation, $\mathbf{v}_h$ the horizontal velocity vector and $\hat{\mathbf{z}}$ the unit vector in the vertical direction.

The surface-parallel flow assumption (see Eq. 1.5) constrains the ice-flow to be parallel to the glacier or ice-sheet surface and allows to reduce the number of observations needed to two (Joughin et al., 1998; Kumar et al., 2011; Mohr et al., 1998). The use of ascending and descending passes from the same scene is a possible approach for solving the 3D displacement vector (Joughin et al., 1998).

1.2.3 SAR multiple aperture interferometry (MAI)

The high accuracy of the differential interferometric measurements, whose main shortcomings are the temporal limitation between acquisitions and the capability of measuring only along the radar line-of-sight direction, motivated the development of a new technique that allows obtaining along-track direction deformations from a single image pair (Bechor and Zebker, 2006). This methodology is called multiple-aperture InSAR (MAI) and it is a split-beam method.

The deformation in the along-track direction is measured using sub-aperture processing (Bechor and Zebker, 2006). The methodology consists on construction forward- and backward-looking interferograms of the scene. The phase difference between the calculated interferograms will be proportional to the along-track motion (Bechor and Zebker, 2006). Forward- and backward-looking MAI pairs have different perpendicular baselines. Therefore, a small orbital deviation could lead to a significant flat-Earth phase. This effect and the coherence improvement between image pairs has been tackled in the literature (Jung et al., 2009).

The forward-looking interferogram is formed using the forward part of the antenna beam width, and the backward-looking interferogram is formed with the back half of the antenna beam. The formulas that define the phases for the forward and backward interferograms are (Bechor and Zebker, 2006):

$$\Phi_{forward} = -\frac{4\pi x}{\lambda} \sin(\Theta_{SQ} + \frac{\alpha}{4}) , \quad (1.6)$$

$$\Phi_{backward} = -\frac{4\pi x}{\lambda} \sin(\Theta_{SQ} - \frac{\alpha}{4}) , \quad (1.7)$$

$$\Phi_{MAI} = \Phi_{forward} - \Phi_{backward} . \quad (1.8)$$

where α is the antenna angular beam width, Θ_{SQ} the radar nominal squint angle (i.e. the angle that the transmission is offset from the normal of the plane of the antenna) and x is the along-track displacement.

The precision of the calculated along-track displacements outperforms by at least a factor of 2 over the amplitude pixel correlation results using the same image pairs (Bechor and Zebker, 2006; Jung et al., 2009). On the other hand, the limitations on the temporal correlation between acquisitions of the interferometric technique still apply to the current methodology. Similarly, the simultaneous application of both types of methodologies, namely speckle-tracking methods and interferometric SAR data processing, has been dealt with in several instances in the literature (Liu et al., 2007; Scheiber et al., 2015).

1.2.4 SAR intensity offset tracking

The limitations of all the interferometric-related techniques can be overridden by the use of SAR intensity offset tracking. This methodology is more resilient to loss of coherence and therefore to increasing time lapses between image acquisitions (Strozzi et al., 2002). There are two main sources of decorrelation. On one hand, meteorological sources, among which we can find ice and snow surface melt, snowfall and snow redistribution by wind. The second decorrelation source is rapid and incoherent flow. SAR offset-tracking is a good alternative to SAR interferometry for resolving glacier surface displacements when the second type of decorrelation takes place (Strozzi et al., 2002). Normally, using large image patches improves the results of cross-correlation and helps avoiding the correlation loss between scenes separated by large time lapses.

Intensity tracking cross-correlation produces an offset field generated using a normalized cross-correlation between a reference and a movable patch between two real-valued SAR intensity images. The normalized cross-correlation function is defined as (Werner et al., 2005)

$$R(n_1, n_2) = \sum_{m_1=0}^{M_1} \sum_{m_2=0}^{M_2} I_1(m_1 + n_1, m_2 + n_2) I_2^*(m_1, m_2) , \quad (1.9)$$

where I_1 and I_2 are the single-look complex (SLC) intensity images of a reference time and a later acquisition. This function is more efficient when applied in the frequency domain. Therefore,

$$R(n_1, n_2) = FFT^{-1} [\tilde{I}_1(m_1, m_2) \tilde{I}_2^*(m_1, m_2)] . \quad (1.10)$$

Image oversampling increases the sampling of the correlation function proportionally with the oversampling factor (Werner et al., 2005). The correlation function is modelled with a two-dimensional regression fit. The peak of the fitted function gives the estimated offset.

The accuracy of the estimate is given by the comparison of the height of the correlation peak with the average level of the correlation function (Strozzi et al., 2002).

The resilience to image temporal decorrelation is the main advantage of this technique when compared with the interferometric technique. On the other hand, its main disadvantage is the lower accuracy of the estimated offsets. Common figures for the errors in the slant-range and azimuth offsets range between 1/20 and 1/30 of a pixel when using typical image-patch sizes and oversampling factors (Strozzi et al., 2002; Werner et al., 2005).

1.2.5 SAR coherence offset tracking

This algorithm calculates a series of small interferograms with changing offsets. The coherence is estimated for each of these interferograms. The corresponding offset is estimated locating the coherence maximum. The maximum is calculated using a 2D regression function which fits the estimated coherence of the interferograms from the single-look complex SAR image patches to model the coherence peak. The quality of the estimation is given by the magnitude of the coherence maximum relative to the average level (Strozzi et al., 2002).

The main advantage of this technique is the possibility to apply it to areas with few surface features such as the large ice-sheets of Greenland and Antarctica. SAR coherence tracking (also known as speckle tracking) matching algorithm is based in the complex-retrieved signal with the advantage of having larger cross-correlation for low correlation areas in contrast with SAR intensity tracking. Another advantage is the smaller patch sizes for cross-correlation when compared with the intensity offset tracking technique (Joughin, 2002). The main disadvantage of the technique is its lower resilience to surface de-correlation. In other words, phase gradients across the patches being matched, caused by steep topography or regions with high shear, can reduce the correlation peak (Joughin, 2002).

Considering its favorable characteristics for ice-velocity mapping of large featureless areas, this technique has been successfully applied to the whole Antarctic Ice Sheet (Mouginot et al., 2012). Furthermore, research on how to combine this technique with interferometry has also been developed (Joughin, 2002; Liu et al., 2007; Scheiber et al., 2015).

1.2.6 Optical satellite imagery matching methods

The normalized cross-correlation algorithm can be applied to consecutive orthophotos (Kääb and Vollmer, 2000). The main disadvantage of using optical imagery for cross-correlation is the presence of clouds in the acquired scenes. Contrastingly, using SAR images guarantees the acquisition of the scenes under any sky condition. Heid and Kääb (2012) proposed several image matching methods. Among them, we find the normalized cross-correlation

in the spatial domain and in the frequency domain using the Fast Fourier Transform (FFT). The phase correlation in the frequency domain using the FFT method allows to ignore the differences in image intensity. Also, orientation correlation (Fitch et al., 2002) can be used for deriving ice surface velocities. Orientation correlation is illumination-invariant (Fitch et al., 2002). The orientation vector takes zero value in uniform regions and one in non-uniform areas (see Eq. 1.11 and 1.12). This property is well suited for glacier ice because uniform areas are very common (Heid and Kääb, 2012). The latter is also of importance when matching striped Landsat-7 images affected by the failure of the scan line corrector (SLC-off). In this situation, stripes are ignored during correlation (Heid and Kääb, 2012).

$$f_0(x, y) = \text{sgn}\left(\frac{\partial f(x, y)}{\partial x} + i \frac{\partial f(x, y)}{\partial y}\right), \quad (1.11)$$

$$g_0(x, y) = \text{sgn}\left(\frac{\partial g(x, y)}{\partial x} + i \frac{\partial g(x, y)}{\partial y}\right), \quad (1.12)$$

$$\text{sgn}(x) = \begin{cases} 0 & \text{if } |x| = 0 \\ \frac{x}{|x|} & \text{otherwise} \end{cases}. \quad (1.13)$$

where f and g represent the images at times 1 and 2 and f_0 and g_0 their corresponding orientation images that consist of real and imaginary parts (Fitch et al., 2002).

These type of methodologies generate surface velocity fields with an accuracy better than 1/10 of a pixel. Therefore, other factors that affect the image at the same resolution level, such as erroneous topographic corrections, sensor geolocation errors and altitude variations, need to be considered (Heid and Kääb, 2012; Kääb and Vollmer, 2000).

1.3 Objectives

The main objectives of this thesis are summarized in the following bullet points:

1. Development of an algorithm for generating improved glacier surface velocity fields from state-of-the-art TOPS Sentinel-1 SAR images. Surface velocity is one of the two components needed for estimating ice discharge to the ocean; the other is ice thickness. Hence the importance of improving the glacier velocity field estimates. Common intensity offset-tracking methods are well suited for sensors with similar resolutions in the slant-range and azimuth directions. The larger errors of the latter, together with the influence of the ionospheric effect on azimuth offsets, motivates the use of only slant-range offsets from both ascending and descending passes of the Sentinel-1 two-satellite constellation.

2. Characterization and quantification of the different error sources to the total error budget in the estimates of frontal ablation of marine-terminating glaciers. The knowledge of the various error components is relevant for constraining the shares of surface ablation and frontal ablation to total mass loss from glaciers. It is also important to understand how the error sources change their respective contributions to the total error with glacier characteristics such as glacier size. Improving the way to calculate frontal ablation, and to estimate its error, are important contribution of this study.
3. Obtaining new and improved estimates of frontal ablation from the Canadian High Arctic and the Academy of Sciences Ice Cap, in Severnaya Zemlya, Russian Arctic. The Canadian Arctic has been extensively studied by many scientists during the last decades. Therefore it is an ideal scenario to test new observation methodologies (such as those proposed in this thesis) against existing methodologies. Once these techniques are successfully compared, the new technique can be applied to other Arctic regions lacking consistent and continuous observation records such as Severnaya Zemlya, and specifically the Academy of Sciences Ice Cap.
4. Sentinel-1 two-satellite constellation acquires images with a 12-day time-lapse when only one sensor is considered, and with 6 days when both sensors are considered. This temporal resolution, together with the free-of-charge accessibility policy, makes Sentinel-1 an ideally-suited sensor for carrying out studies of temporal changes of glaciers and ice caps. This motivates our analysis of the time variability of the ice surface velocity field of the Academy of Sciences Ice Cap and the corresponding variability of its frontal ablation.
5. Calculating the shares of frontal and surface ablation to total mass loss of the Academy of Sciences Ice Cap. Few studies reveal the partitioning of both contributions to total mass loss and hence the relevance of such an analysis. This partitioning is of interest for calibrating/validating suitable models of frontal ablation on larger scales.

1.4 Thesis organization

This dissertation is divided into 6 chapters. The first chapter is of an introductory nature, giving an overall view on the research topic that has been tackled in the thesis and the main motivations behind it. A brief overview on the current status of the methodologies used nowadays for retrieving ice surface velocities from remote sensing data has also been discussed in the state-of-the-art section. The second chapter is devoted to the description

of the Synthetic Aperture Radar sensors, how a synthetic image is formed and how image coregistration and image cross-correlation are performed. The following 3 chapters, from chapter 3 to chapter 5, are literal transcriptions of two already-published papers and a further contribution that, at the time of writing this manuscript, was under review in the *Journal of Glaciology*. In Chapter 3 a novel methodology for calculating a glacier surface velocity field from Sentinel-1 imagery is described. Chapter 4 is devoted to the analysis of the total error budget in frontal ablation and the contribution to the total error by the various error sources; the methods are applied to glaciers of the Canadian High Arctic. Chapter 5 deals with the analysis of the frontal ablation, and its seasonal and inter-annual variability, of the Academy of Sciences Ice Cap, in Severnaya Zemlya, Canadian Arctic. The shares of surface ablation and frontal ablation to the total mass loss of the ice cap are also quantified. Finally, chapter 6 tackles the general conclusions of the research involved in this thesis. Also, an outlook is drawn on the research lines that can be followed in the future.

Chapter 2

Methodology

2.1 Introduction

Synthetic Aperture Radar (SAR) is an active sensor that emits radiation in the microwave region of the electromagnetic spectrum (commonly between 0.25 and 40 GHz, and from 120 to ~ 1 cm in frequency and wavelength, respectively; see Table 2.1). Vision is based on the reception, sensing and processing in the brain of the visible part of the electromagnetic spectrum (between 400 and 700 nm). Vision is a trait that is shared by many creatures and represents a breakthrough in evolution directly linked with the presence of this type of radiation due to the atmospheric window that allows the visible wavelengths of the spectrum to penetrate the atmosphere (Soumekh, 1999). The angular resolution is related with the Rayleigh criterion for calculating the minimum resolvable detail of a device. From Equation (2.1), it is immediate to notice the relationship between the wavelength and the resolving capability. That is to say, the human eye or a telescope are prepared to resolve objects orders of magnitude smaller than a radar-based device.

$$Resolution = \frac{R\lambda}{D}, \quad (2.1)$$

where R is the range to the object to be resolved, D is the diameter of the telescope or radar antenna and λ the radiation wavelength.

Radar active systems, as human vision for the visible part of the spectrum, are devised to exploit the atmospheric window of the microwaves (Curlander and McDonough, 1991; Soumekh, 1999). Furthermore, as opposed to optical-based systems, radar is capable of penetrating an atmosphere with clouds. The resolving capabilities of a common radar system are different in the range and cross-range directions. This disparity in resolution motivated the search for a way of improving the resolution in the lateral direction. The way to tackle

Table 2.1 Frequency and wavelength ranges for SAR systems

Frequency band	Frequency (GHz)	Wavelength (cm)
P	0.25-0.5	120-60
L	1-2	30-15
S	2-3.75	15-8
C	3.75-7.5	8-4
X	7.5-12	4-2.5
Ku	12-17.6	2.5-1.7
Ka	25-40	1.2-0.75

the problem was the use of a synthetic aperture, which is a lateral motion-based technique that exploits the Doppler spectrum of the retrieved signal to increase the resolution in that direction.

The Radar Synthetic Aperture was invented in 1951 by the mathematician Carl A. Wiley while working at Goodyear Aircraft Company, when he was working on a guidance system for the Atlas ICBM program (Wiley, 1985). In the year 1952, Carl Wiley, Fred Heisley and Bill Welty designed the DOUSER or "Doppler Unbeamed Search Radar". During the following two decades Goodyear Aircraft introduced many new technologies to SAR-thriving systems.

2.2 Range Imaging

The range imaging problem consists in calculating the target range distance to the sensor and the reflectivity from the echoed signal. The discussion in the whole section will cope with a simplified one-dimensional (range direction) target area. A system model will provide the theoretical background for achieving this goal. The actual procedure of calculating the target range and reflectivity from the retrieved signal is called reconstruction or inversion of the system model (Soumekh, 1999). The parameter utilized for calculating the target range x_n and reflectivity σ_n is the radar signal frequency. Equation (2.2) shows the *ideal* target function:

$$f_0(x) = \sum_n \sigma_n \delta(x - x_n), \quad (2.2)$$

where σ_n and x_n are the reflectivity and cross-range of the target, and the only unknown of function f_0 is the target range. Considering a one-dimensional target area in the range domain, the reflected signal is (Soumekh, 1999)

$$s(t) = \sum_n \sigma_n p(t - \frac{2x_n}{c}) , \quad (2.3)$$

where c is the speed of light and $p(t)$ is the transmitted signal. The echoed signal can also be written in the following form:

$$s(t) = f_0(\frac{ct}{2}) * p(t) . \quad (2.4)$$

The Fourier transform of the reflected signal is:

$$S(\omega) = P(\omega) \sum_n \exp(-j\omega \frac{2x_n}{c}) , \quad (2.5)$$

where $P(\omega)$ is the Fourier transform of the transmitted pulse. The ideal target function can be formed dividing $S(\omega)$ by $P(\omega)$ and applying the inverse Fourier transform to the result:

$$f_0(\frac{ct}{2}) = \sum_n \sigma_n \delta(t - \frac{2x_n}{c}) = \mathcal{F}^{-1}(\omega) [\frac{S(\omega)}{P(\omega)}] . \quad (2.6)$$

The situation described is impractical due to the system limitations, among which we find its finite bandwidth and sampling rate.

The reconstruction via matched filtering consists precisely in multiplying the complex conjugate of the reference transmitted signal $P^*(\omega)$, also called the matched filter, and the retrieved signal in the frequency domain:

$$s_M(t) = \mathcal{F}^{-1}(\omega) [S(\omega) P^*(\omega)] = \mathcal{F}^{-1}(\omega) [\sum_n \sigma_n |P(\omega)|^2 \exp(-j\omega \frac{2x_n}{c})] , \quad (2.7)$$

$$\omega \in [\omega_c - \omega_0, \omega_c + \omega_0] .$$

The point spread function is built upon the reference transmitted signal (see Eq. 2.8). It defines the resolvability of a target in the range domain:

$$psf_t(t) = \mathcal{F}^{-1}(\omega) [|P(\omega)|^2] . \quad (2.8)$$

The signal bandwidth will define the system ability to resolve a target in the range domain, as it is shown in the following equation (Soumekh, 1999):

$$psf_t(t) = \exp(j\omega_c t) \text{sinc}(\frac{\omega_0 t}{\pi}) , \quad (2.9)$$

where ω_c and ω_0 are the carrier and baseband bandwidths of the emitted radar signal. Obviously, the larger the signal bandwidth the sharper the point spread function in the range

or time domain. If the baseband support of $P(\omega)$ is $2\omega_0$ in the frequency domain, the support of the point spread function in the time domain is approximately π/ω_0 . Multiplying the time resolution by the speed of light and dividing by 2 to obtain a distance, we finally arrive to the following expression for the range resolution:

$$\Delta_x = \frac{c}{2} \frac{\pi}{\omega_0} . \quad (2.10)$$

We have seen that both the bandwidth and the sampling rate of the signal are limiting factors for the radar systems. The bandwidth influence on the resolution has already been described. The signal sampling rate defines whether the reconstructed signal is aliased or not. The Nyquist sampling criterion applies here and, considering the bandwidth of the signal ω_0 , the temporal interval of sampling is defined by

$$\Delta_t \leq \frac{\pi}{\omega_0} . \quad (2.11)$$

2.3 Cross-range Imaging

Cross-range imaging consists in calculating the target cross-range position and reflectivity using a system model. In this section, the problem will be simplified assuming that all targets lie at the same range distance from the radar antenna. We will also assume, as opposed to what we considered in the range imaging problem, that the transmitted radar signal is a constant single-frequency signal. As we did in the previous section, we define an *ideal* target function:

$$f_0(y) = \sum_n \sigma_n \delta(y - y_n) . \quad (2.12)$$

The distance from the target to the radar is given by

$$\sqrt{x_n^2 + (y_n - u)^2} . \quad (2.13)$$

The signal retrieved by the radar antenna is given by

$$s(t, u) = \sum_n \sigma_n p\left[t - \frac{2\sqrt{x_n^2 + (y_n - u)^2}}{c}\right] = \exp(j\omega t) \sum_n \sigma_n \exp[-j2k\sqrt{x_n^2 + (y_n - u)^2}] , \quad (2.14)$$

where u is the synthetic aperture coordinate of the antenna and $k = \omega/c$ is the wavenumber. After fast-time baseband conversion, the echoed signal is expressed as

$$s(\omega, u) = s(t, u) \exp(-j\omega t) = \sum_n \sigma_n \exp[-j2k\sqrt{x_n^2 + (y_n - u)^2}] . \quad (2.15)$$

We define a reference signal which represents a baseband signal for a unit reflector at the centre of the broadside target area:

$$s_0(\omega, u) = \exp(-j2k\sqrt{X_c^2 + u^2}) . \quad (2.16)$$

Therefore, the recorded signal can be written as a function of $f_0(y)$ and $s_0(\omega, u)$:

$$s(\omega, u) = f_0(u) * s_0(\omega, u) . \quad (2.17)$$

The signal can be expressed as the sum of the reflected signals from the targets in the scene (Soumekh, 1999):

$$s_n(\omega, u) = \sigma_n \exp[-j2k\sqrt{x_n^2 + (y_n - u)^2}] , \quad (2.18)$$

$$s(\omega, u) = \sum_n s_n(\omega, u) . \quad (2.19)$$

The function on the right-hand side of Equation (2.18) is a phase-modulated (PM) signal, also known as spherical PM signal. This function is a nonlinear function of both the target and radar coordinates.

The Fourier transform of the spherical PM signal $s_n(\omega, u)$ with respect to the radar position u (synthetic aperture) leads to the following expression:

$$S_n(\omega, k_u) = \sigma_n \exp(-j\sqrt{4k^2 - k_u^2}x_n - jk_u y_n) , \quad (2.20)$$

$$k_x = \sqrt{4k^2 - k_u^2} ,$$

where k_u is the slow-time frequency or Doppler domain, and $k_u \in [-2k, 2k]$. Therefore, $S_n(\omega, k_u)$ has a finite-support band. The term σ_n in Equation (2.20) includes a slowly-fluctuating amplitude function not relevant in this analysis. On the other hand, the phase term in the Equation is key to cross-range image formation. Equation (2.20) has a linear phase function of (x_n, y_n) , while Equation (2.15) contains a nonlinear phase function of the same variables.

We assume that all targets are at the same range distance to the radar, X_c . Therefore, Equation (2.20) can be written as

$$S(\omega, k_u) = \exp(-j\sqrt{4k^2 - k_u^2}X_c) \sum_n \sigma_n \exp(-jk_u y_n) , \quad (2.21)$$

$$x_n = X_c .$$

The reference signal of a unit reflector at the center of the broadside area and the ideal target function in the frequency domain are:

$$S_0(\omega, k_u) = \exp(-j\sqrt{4k^2 - k_u^2}X_c) , \quad (2.22)$$

$$F_0(k_y) = \sum_n \sigma_n \exp(-jk_y y_n) . \quad (2.23)$$

Equation (2.21) can be rewritten as

$$S(\omega, k_u) = S_0(\omega, k_u)F_0(k_u) . \quad (2.24)$$

The reconstruction of the ideal target function can be achieved using matched filtering in the following manner:

$$F(k_y) = S(\omega, k_u)S_0^*(\omega, k_u) = \sum_n \sigma_n \exp(-jk_y y_n) . \quad (2.25)$$

The inverse Fourier transform of $F(k_y)$ with respect to the spatial frequency k_y gives the cross-range reconstructed image:

$$f(y) = \sum_n \sigma_n \text{sinc}\left[\frac{k}{\pi}(y - y_n)\right] . \quad (2.26)$$

The cross-range resolution of the retrieved signal depends on the bandwidth of the observability of the targets in the k_u domain, which is $k_y \in [-2k, 2k]$. Therefore,

$$\Delta_y = \frac{\pi}{2k} = \frac{\lambda}{4} , \quad (2.27)$$

where $\lambda = 2\pi/k$ is the radar wavelength. The alias-free sampling of the signal is achieved satisfying the Nyquist criterion. Considering the support band of the echoed signal $S(\omega, k_u)$ in the slow-time frequency domain, the sampling rate is

$$\Delta_u \leq \frac{\pi}{2k} . \quad (2.28)$$

2.4 Generic Synthetic Aperture Radar

This section will cover the generic SAR system description considering a simple system model with an omnidirectional emitting antenna. A basic image reconstruction method will also be described. The system model is described by the following equation:

$$s(t, u) = \sum_n \sigma_n p\left[t - \frac{2\sqrt{x_n^2 + (y_n - u)^2}}{c}\right]. \quad (2.29)$$

Applying the Fourier transform with respect to the fast-time, we obtain

$$s(\omega, u) = P(\omega) \sum_n \sigma_n \exp[-j2k\sqrt{x_n^2 + (y_n - u)^2}]. \quad (2.30)$$

We further apply the Fourier transform with respect to the slow-time, considering an infinite support $u \in (-\infty, \infty)$:

$$S(\omega, k_u) = P(\omega) \sum_n \sigma_n \exp(-j\sqrt{4k^2 - k_u^2}x_n - jk_u y_n). \quad (2.31)$$

In the latter section, we saw that the phase function in Equation (2.31) is a linear function of (x_n, y_n) . This property will be fundamental for formulating the SAR reconstruction algorithms (Curlander and McDonough, 1991; Soumekh, 1999).

We define two functions that map the radar frequency and the slow-time frequency domain (ω, k_u) into spatial frequency variables (k_x, k_y) . The mapping is done using the following formulae:

$$\begin{aligned} k_x(\omega, k_u) &= \sqrt{4k^2 - k_u^2}, \\ k_y(\omega, k_u) &= k_u. \end{aligned} \quad (2.32)$$

We can see that the mapping into the k_y is evenly spaced, since $k_y = k_u$. Equation (2.31) can be formulated as

$$S(\omega, k_u) = P(\omega) \sum_n \sigma_n \exp[-jk_x(\omega, k_u)x_n - jk_y(\omega, k_u)y_n]. \quad (2.33)$$

Defining the ideal target function and its two-dimensional Fourier transform as

$$\begin{aligned} f_0(x, y) &= \sum_n \sigma_n \delta(x - x_n, y - y_n), \\ F_0(k_x, k_y) &= \sum_n \sigma_n \exp(-jk_x x_n - jk_y y_n), \end{aligned} \quad (2.34)$$

the retrieved signal can be also expressed in the frequency domain as

$$S(\omega, k_u) = P(\omega) F_0[k_x(\omega, k_u), k_y(\omega, k_u)]. \quad (2.35)$$

Considering that the signal has the support in $k_u \in [-2k, 2k]$ and $\omega \in [\omega_c - \omega_0, \omega_c + \omega_0]$, the reconstruction can be done using fast-time matched filtering as

$$F[k_x(\omega, k_u), k_y(\omega, k_u)] = P^*(\omega)S(\omega, k_u) = |P(\omega)|^2 \sum_n \sigma_n \exp(-jk_x x_n - jk_y y_n) . \quad (2.36)$$

The SAR system measures the echoed signal $S(\omega, k_u)$ with evenly-spaced samples in the (ω, k_u) domain. However, the two-dimensional mapping from the $S(\omega, k_u)$ to the (k_x, k_y) domain is nonlinear. Therefore, the resulting mapped function $F(k_x, k_y)$ is unevenly spaced (Soumekh, 1999). The function $F(k_x, k_y)$ needs to be defined on a uniform rectangular grid if we want to obtain the function $f(x, y)$ using the two-dimensional Fast Fourier Transform (FFT).

The sample spacing of the measured signal $S(\omega, k_u)$ in the slow-time Doppler domain is defined by

$$\Delta_{k_u} = \frac{2\pi}{M\Delta_u} , \quad (2.37)$$

where M is the number of samples and Δ_u is the sample spacing in the Synthetic Aperture domain. The following expression describes the mapping between domains:

$$k_{ymn} = k_{um} = m\Delta_{k_u} . \quad (2.38)$$

This mapping does not present any problem. However, the problem lies in the interpolation in the k_x from the unevenly spaced samples given by the formulae:

$$k_{xmn} = \sqrt{4k_n^2 - k_{um}^2} , \quad (2.39)$$

with

$$k_n = n\Delta_k = n\frac{\Delta\omega}{c} , \quad (2.40)$$

where $\Delta\omega$ is the sample spacing of the echoed SAR signal in the fast-time frequency domain. Before proceeding with the aforementioned interpolation step, we need to perform a baseband conversion to the retrieved signal. This is done to transform the signal to a lowpass signal. The echoed signal $F(k_x, k_y)$ is a bandpass signal since its inverse Fourier transform is supported by the region $x \in [X_c - X_0, X_c + X_0]$ and is centred at $x = X_c$. To achieve this, we apply a baseband conversion:

$$F_b(k_x, k_y) = F(k_x, k_y) \exp(jk_x X_c) = P^*(\omega) \exp(j\sqrt{4k^2 - k_u^2} X_c) S(\omega, k_u) . \quad (2.41)$$

The generic SAR signal of unit reflector located at the broadside and its Fourier transform are:

$$s_0(t, u) = p\left[t - \frac{2\sqrt{X_c^2 + u^2}}{c}\right], \quad (2.42)$$

$$S_0(\omega, k_u) = P(\omega) \exp(-j\sqrt{4k^2 - k_u^2} X_c). \quad (2.43)$$

Therefore:

$$F_b(k_x(\omega, k_u), k_y(\omega, k_u)) = S(\omega, k_u) S_0^*(\omega, k_u). \quad (2.44)$$

The baseband target function $F_b(k_x, k_y)$ is retrieved using a two-dimensional matched filtering of the measured radar echoed signal with the defined generic SAR signal.

2.5 Terrain Observation by Progressive Scans

The most common SAR acquisition schemes are STRIPMAP and SPOTLIGHT. SPOTLIGHT acquisition is based on the antenna focusing on the same target area during the whole acquisition period, therefore increasing the bandwidth in the slow-time Doppler domain and increasing the resolution of the image. This SAR imaging modality uses either a mechanical or an electronic beam steering of the physical radar antenna. In STRIPMAP SAR imaging, as opposed to SPOTLIGHT, the antenna is not steered and keeps its broadside radiation pattern during the whole acquisition. This modality implies that the illuminated cross-range area varies with the displacement of the radar antenna. The bandwidth in the slow-time Doppler domain decreases and the resolution in the same direction also diminishes.

Nowadays, space agencies are trying to improve the SAR-mission capabilities by reducing the revisit times using increased wide-range coverage. The first attempt to increase the SAR range coverage was done using scanning synthetic aperture radar acquisition (henceforth, ScanSAR). This imaging modality consists in sampling the target area in several range subswaths by periodically changing the antenna elevation beam (Moore et al., 1981). The antenna elevation steering forces the sensor to retrieve a limited series of echoes within a dwell time from the same subswath, which is called burst. The limited amount of time dedicated to sample a subswath, as compared to the STRIPMAP acquisition sampling limitation (the antenna beamwidth in the azimuth direction), decreases the signal bandwidth in the slow-time frequency domain decreasing the azimuth resolution. Therefore, the larger

the range coverage (more subswaths and less dwell time per burst) the lower the azimuth resolution (lower amount of samples in azimuth and smaller bandwidth).

The main limitations of the ScanSAR technique, apart from the decreased azimuth resolution, are, firstly, the space-varying modulation of the amplitude called scalloping, caused by the different slices of the Azimuth Antenna Pattern (AAP) retrieved for each target. The second limitation is the azimuth-varying Doppler spectrum, since interferometric applications from two acquisitions need the same spectrum from the same targets (Zan and Guarnieri, 2006).

Zan and Guarnieri (2006) designed a new SAR acquisition scheme which is very similar to ScanSAR imaging. The main difference is the rotation of the radar antenna in the opposite direction of SPOTLIGHT acquisitions. This antenna rotation has the effect that all targets are affected by the same AAP weighting. Therefore, ambiguities become stationary in azimuth, azimuth resolution stabilises and scalloping disappears (Zan and Guarnieri, 2006).

Sentinel-1 satellite SAR sensor acquires imagery using TOPS imaging modality. The number of subswaths and bursts per image are 3 and 9, respectively. The main advantage of Sentinel-1 TOPS imagery is the large ground coverage with a swath width of 250 km, with an acceptable resolution of 5 and 20 metres in the range and cross-range directions. The main disadvantages are the restrictive approaches both for coregistration (Grandin, 2015a) and for interferometric processing (Scheiber et al., 2015).

2.5.1 TOPS SAR coregistration

TOPS Single Look Complex (SLC) coregistration requires very high accuracy (a few thousands of a pixel) in the azimuth direction for avoiding phase jumps between subsequent bursts; otherwise, interferometry would not be applicable (Wegmüller et al., 2016). To achieve this level of coregistration accuracy, preliminary information on the target area topography is needed. A first calculation of a lookup table is done using the highly-accurate Sentinel-1 orbit parameters and a preliminary DEM. This lookup table is later on refined using several methods such as matching procedures or the spectral diversity method, which considers the interferometric phase of the burst overlap regions (Geudtner et al., 2014a; Grandin, 2015a). This type of SAR acquisitions imply the antenna steering from aft to fore during burst acquisition leading to an azimuth variant Doppler centroid (Scheiber et al., 2015), which is corrected with a deramping operation of the Doppler spectrum.

Once the phase jumps between consecutive bursts have been erased using the mentioned coregistration methodology, we can normally apply the known interferometric processing to the corrected and perfectly coregistered SLC Sentinel-1 images.

2.5.2 TOPS SAR cross-correlation

The cross-correlation methodology between two SAR images has been already described in Chapter 1. However, applying cross-correlation to TOPS SAR imagery requires several processing steps of the original acquired TOPS image. Firstly, the SAR images need to be coregistered in the way described in the previous section. Then, considering that image oversampling has to be applied for improving the quality calculated offsets (Wegmüller et al., 2016), one needs to first deramp the reference SAR image, storing the phase ramps used for deramping and applying the same phase ramp to the second coregistered SAR image.

Once these processing steps are performed over the SAR images, the remaining offset tracking processing steps (geocoding, conversion of displacements to metres and quality control) are the same as for a normal STRIPMAP SAR image.

Chapter 3

Glacier surface velocity retrieval using D-InSAR and offset tracking techniques applied to ascending and descending passes of Sentinel-1 data for southern Ellesmere ice caps, Canadian Arctic¹

3.1 Introduction

Remote sensing is becoming an increasingly important tool in climate research (Cracknell and Varotsos, 2011), with Synthetic Aperture Radar (SAR) being one of the techniques that has experienced a faster growth. The European Space Agency (ESA) has long supported various satellite Earth Observation missions with SAR sensors onboard such as the European Remote Sensing (ERS-1 and ERS-2) and ENVISAT satellites. Continuing this long-term policy of providing continuous and consistent observational data, Sentinel-1A was launched in 2014, with its twin satellite, Sentinel-1b, launched in 2016. Both are equipped with a C-Band sensor (Berger et al., 2012). The Sentinel mission is comprised of a set of three different observational platforms intended to provide measurements that will contribute to better understanding the Earth System (Malenovský et al., 2012). The Sentinel mission is part of the Copernicus program (formerly known as Global Monitoring for Environment and Security, GMES), which comprises three components: space, in-situ and services. The latter component is in charge of defining the Sentinel-1 observation scenario (Potin et al.,

¹(Sánchez-Gómez and Navarro, 2017)

Table 3.1 Sentinel-1 Beam modes with incidence angles, nominal resolution and scene size.

Beam Mode	Incidence Angle (degrees)	Resolution (m)	Swath Width (km)
Stripmap	20-45	5 x 5 m	80 km
Interferometric Wide swath (IW)	29-46	5 x 20 m	250 km
Extra Wide swath (EW)	19-57	20 x 40 m	400 km
Wave	22-35/35-38	5 x 5 m	20 x 20 km

2014). The space component is intended to provide free of charge easily accessible data for registered users (Aschbacher and Milagro-Pérez, 2012). The Sentinel-1 mission maintains continuity with key characteristics of former GMES missions (e.g. ERS and ENVISAT) including stable and accurate, well-calibrated products while at the same time enhancing other key mission aspects such as data reliability, revisit time and geographic coverage (Torres et al., 2012). A major improvement of the Sentinel-1 mission has been the reduction of the revisit time of the sensor, which has dropped from 35 days for ENVISAT to 12 days for Sentinel-1A, and 6 days when considering the full Sentinel-1 constellation. Furthermore, the coverage has increased from 80 km (for image mode in ERS and ENVISAT) to 250 km in the case of Sentinel-1 with its interferometric wide-swath (IW) mode (Table 3.1).

The Terrain Observation by Progressive Scans (TOPS) acquisition mode of Sentinel-1 allows for a substantial increase in geographical coverage at the expense of the increasing difficulty associated with co-registering overlapping bursts. A small co-registration error can produce azimuth phase ramps. This problem is solved by applying a spectral diversity method over the overlapping areas of contiguous bursts (Geudtner et al., 2014b; Grandin, 2015b). This type of SAR acquisitions imply the antenna steering from aft to fore during burst acquisition leading to an azimuth variant Doppler centroid (Scheiber et al., 2015) which is corrected with a deramping operation of the Doppler spectrum.

There are several examples in the literature that apply the interferometric technique to TOPS acquisitions (Marotti et al., 2011; Prats-Iraola et al., 2012). The application of interferometry to ascending and descending passes is also well documented for glacier surface velocity estimations (Joughin et al., 1998; Kumar et al., 2011; Mohr et al., 1998). There are plenty of articles dealing with glacier velocity retrieval using the offset tracking technique (Fallourd et al., 2010; Pritchard, 2005; Strozzi et al., 2008, 2002; Werner et al., 2005), including the particular case of Sentinel-1 TOPS acquisitions (Dall et al., 2015; Nagler et al., 2015), the latter reference providing a comprehensive velocity field for the entire Greenland Ice Sheet. Furthermore, the combination of both techniques has also been covered with different approaches (e.g. merging of results or using least squares (Gray, 2005; Gudmundsson et al., 2002; Joughin, 2002; Liu et al., 2007)) and also for entire ice sheets, such as the study (Mouginot et al., 2012) for the Antarctic Ice Sheet. Finally, Hu et al. (2014)

give a holistic approach on the state of the art of SAR techniques for surface deformation retrieval.

The aim of the paper is to show the performance of the Sentinel-1 SAR Single Look Complex (SLC) Interferometric Wide-swath level-1 product (with a nominal resolution of 5 and 20 m in the across- and along-track directions respectively) for retrieving glacier surface velocities using offset tracking and differential interferometry (D-InSAR) techniques. We estimate the surface velocity using differential interferometry where this technique proves to be more useful, namely land-terminating glaciers and ice caps characterised by a low amount of deformation, which prevents decorrelation. The offset-tracking technique, in turn, proves to be more useful for fast-flowing glaciers. We apply the latter technique to Sentinel-1 imagery from ascending and descending passes (Fallourd et al., 2010), with the particularity of avoiding the use of azimuth offsets for such velocity estimates. The reason behind this is that Sentinel-1 TOPS IW acquisitions have their worse resolution in the azimuth direction. Additionally we avoid the ionospheric effect strongly affecting this direction (Gray et al., 2000; Wegmüller et al., 2006). By using this approach, we try to tailor the existing techniques to the particular characteristics of Sentinel-1, while avoiding the cumbersome corrections of ionospheric disturbances. Furthermore, Sentinel-1 acquisition plan often includes ascending and descending passes from the same glaciated areas (Nunavut, Alaska, Himalayas, etc.), therefore making this approach suitable to be readily used in those cases.

3.2 Study Area and Data Sources

Ellesmere Island is located in the Canadian High Arctic, neighbouring the western coast of Greenland (Figure 3.1). Ellesmere is part of the Northern Canadian Arctic Archipelago (NCAA) whose glacierized area in 2000 was around 104 000 km² (Sharp et al., 2014) and contains one-third of the global volume of land ice outside Greenland and Antarctica (Radić and Hock, 2010). The climate characteristics of this location are low amounts of precipitation during the winter time (300-600 mm yr⁻¹) and increasing surface temperatures linked with a shift to a dominantly easterly positioned July vortex associated with an increased frequency of tropospheric ridging over the Canadian High Arctic (Alt, 1987; Gardner and Sharp, 2007). This implies that the surface mass balance (SMB) of the NCAA is governed by precipitation, in the form of snow, and increasing meltwater runoff accompanied by a shrinking capability of the snowpack and the firn layer to buffer mass loss through refreezing of percolating meltwater (Lenaerts et al., 2013).

Surface mass budgets in this region are currently showing an increasing negative trend (Koerner, 2005). Between 2005 and 2009, CAA ice caps suffered the most negative trends

Fig. 3.1 (a) Major ice masses of Ellesmere Islands within the Arctic (Wessel and Smith, 1996). The glacier outlines are from the RGI version 5.0 (Pfeffer et al., 2014). (b) Zoom of red rectangle in panel a, showing our three main study areas (I, II and III) in Southern Ellesmere. (c) Northern velocity component for Ekblaw Glacier obtained using intensity offset tracking. The zone shown corresponds to the red rectangle labelled I in panel b. The background images in this panel c were taken between 10th and 17th February 2016. (d) Schematics of the vectors and angles involved in the non-orthonormal estimation of the ice flow velocity. All maps in this paper use the Universal Transverse Mercator (UTM) projection of the zone 17 north, and the reference ellipsoid is WGS84.

in their observational records (Gardner et al., 2011). These trends show that the SMB in the NCAA is moving from a mass loss predominantly dominated by calving during the 20th century to a contemporary mass loss driven by meltwater runoff (Gardner et al., 2011; Lenaerts et al., 2013; Millan et al., 2017; van Wychen et al., 2014).

Studies focusing on the regional glacier dynamics revealed that surge-type glaciers, with a relatively long surge phase, are common in the Canadian High Arctic. Many of them are tidewater glaciers located in the high-precipitation coastal areas (Copland et al., 2003a; van Wychen et al., 2014, 2016). Independent of this surge behaviour, glacier ice velocities during the summer are typically an order of magnitude larger than the mean annual velocities, and summer calving rates 2-8 times larger than the annual average (van Wychen et al., 2016; Williamson et al., 2008). Other studies point to a high interannual variability of the velocities of the large outlet glaciers (Short and Gray, 2005). van Wychen et al. (2014) showed that velocity changes associated with the termination or initiation of surges can rapidly change the rate of ice discharge to the oceans. Finally, many research papers indicate a clear increase in ice mass losses and associated contribution of Canadian Arctic glaciers to sea-level rise during recent years. This increase in mass loss is mostly caused by surface melt and runoff, and not by glacier dynamics (Gardner et al., 2012, 2011; Millan et al., 2017; van Wychen et al., 2014, 2016).

We frame our study within this climatic and geographical context. Specifically, we focus on glaciers located in the Prince of Wales (POW) and the Manson Icefields, and some eastern glaciers of Sydkap Ice Cap (Fig. 3.1). The glaciers in this area are both of land-terminating and tidewater type. The different nature and dynamics of these glaciers, together with the presence of wide ice-free areas (which allows performing a finer co-registration between acquisitions) makes this location optimal to test and apply different SAR surface velocity estimation methodologies. Furthermore, the low amount of precipitations in the area, together with the fact that all acquisitions correspond to northern hemisphere winter months, characterised by lower ice velocities, increases the chance of a successful application of various techniques.

The surface velocities on Ellesmere Island were obtained from a total of 14 Sentinel-1 SAR TOPS IW Level-1 Single Look Complex images taken during February and March 2016 (Table 3.2). These type of products provide a geo-referenced image (using accurate altitude and orbital information from the satellite), in slant-range geometry, processed in zero-Doppler. The image is normally composed of three sub-swaths with each sub-swath comprising normally 9 consecutive bursts, which overlap in azimuth. Burst synchronisation is needed for interferometry and for accurate offset tracking (Holzner and Bamler, 2002). The data were retrieved in two different tracks, with 6 and 9 scenes each, during ascending

Table 3.2 Sentinel-1 processed acquisitions.

Acquisition Date	Beam Mode and Number	Orbit Number	Slice Number	Pass Direction	Methodology
10 February 2016	IW1	9890	5	Ascending	D-InSAR
10 February 2016	IW1	9890	6	Ascending	D-InSAR
10 February 2016	IW1	9890	7	Ascending	D-InSAR
17 February 2016	IW1	9986	3	Descending	D-InSAR
17 February 2016	IW1	9986	4	Descending	D-InSAR
22 February 2016	IW1	10065	5	Ascending	Both
22 February 2016	IW1	10065	6	Ascending	Both
22 February 2016	IW1	10065	7	Ascending	Both
29 February 2016	IW1	10161	3	Descending	Both
29 February 2016	IW1	10161	4	Descending	Both
05 March 2016	IW1	10240	6	Ascending	Intensity Tracking
05 March 2016	IW1	10240	7	Ascending	Intensity Tracking
12 March 2016	IW1	10336	3	Descending	Intensity Tracking
12 March 2016	IW1	10336	4	Descending	Intensity Tracking

and descending passes respectively. The incidence angle, θ , was in all cases close to 33 degrees, while the α angle between ascending and descending tracks was around 40 degrees. The time span between corresponding acquisitions was 12 days, which halves to 6 days when considered the ascending and descending passes. We used the HH channel, which preserves better the amplitude features and has also a higher signal-to-noise ratio as compared to the HV channel, making the former better suited for glacier ice motion retrieval.

The TOPS acquisition mode improves the performance of already existing SAR imaging algorithms such as ScanSAR mode (Geudtner et al., 2014b; Zan and Guarnieri, 2006). During a TOPS mode acquisition the SAR antenna is steered in the azimuth direction from aft to fore with a constant rate. This mode of observation has several advantages; in particular, the measured targets are observed with the whole azimuth antenna pattern, which also reduces the scalloping effect (Zan and Guarnieri, 2006). The main disadvantage is that it imposes a more restrictive approach for the co-registration procedure and for the interferometric processing. The peculiarities of the co-registration with the spectral diversity method have been addressed by Grandin (Grandin, 2015b), while the procedure for obtaining interferometry from TOPS mode images has been dealt with by various authors (Grandin, 2015b; Prats-Iraola et al., 2012; Scheiber et al., 2015).

For the application of the differential interferometry to the SAR imagery, a fine digital elevation model (DEM) of the island was required. For this purpose, we used the freely accessible Canadian DEM (CDEM) designed by Natural Resources Canada (NRCAN) with a resolution of 0.75 and 3 arcsec in the south-north and east-west directions respectively (NRCAN, 2016). The DEM was used to produce the differential interferometry results and for geocoding and co-registering the imagery employed for the intensity offset tracking technique. This is the best currently available DEM from the area and it dates back to the 1960s. We want to highlight here the degrading effects that this could have specially on the results of the D-InSAR technique (Joughin et al., 1998).

3.3 Methods

3.3.1 GAMMA Remote Sensing Offset Tracking

We used GAMMA software (Wegmüller et al., 2016) for processing the SAR Sentinel-1 acquisitions. This software package is extensively used as a tool for retrieving ground deformations and surface displacements in many areas, including glacier-covered regions (Schellenberger et al., 2016; Strozzi et al., 2002). We exclusively applied the intensity offset tracking algorithm. The speckle tracking technique has its own peculiarities when it is applied to TOPS products, in particular when applied to non-stationary areas such as glaciers or ice caps (Dall et al., 2015) (e.g. TOPS azimuth antenna scan entails phase deramping and azimuth common band filtering (Scheiber et al., 2015)). Orbital accurate ephemerides are becoming a common resource in new space missions, as is the case for the Sentinel-1 (Guarnieri et al., 2015; Nagler et al., 2015) mission, which allows for a seamless generation of wide regional surface displacement products.

Sentinel-1 TOPS mode images first need to be co-registered before any offset tracking or interferometric algorithms are applied on them. This fine co-registration procedure requires the use of a fine DEM within the same area of the acquisition image extent (Wegmüller et al., 2016). The co-registration begins with obtaining the DEM in SAR coordinates followed by the matching algorithm and the spectral diversity method (which includes considering the interferometric phase) applied over the overlapping areas of the bursts. The co-registration requirements are quite stringent and a co-registration quality of 1/1000 of a pixel in the azimuth direction is required for the phase discontinuity between consecutive bursts edges to be less than 3 degrees (Scheiber et al., 2015). The co-registration performance can be improved when masking areas with expected azimuth displacements (e.g. glaciers or ice caps). This prevents fine co-registration methodologies to be applied to wide ice caps where no stable ground is found (Dall et al., 2015). This was a critical factor for choosing Ellesmere Island as a suitable area to apply a full co-registration procedure using the matching and the spectral diversity methods. When no interferometry is required and the only derived product is to be offset tracking displacements, the Sentinel-1 ephemerides are sufficient for comprehensive, region-wide results (Nagler et al., 2015).

After a full co-registration is achieved, a deramping of the SLC images for correcting the azimuth phase ramp is required to apply oversampling in the offset tracking procedures (Wegmüller et al., 2016). Once the above-mentioned steps are done, the offset tracking technique is the same as for normal stripmap mode scenes (Fallourd et al., 2010; Pritchard, 2005; Strozzi et al., 2008, 2002; Werner et al., 2005). Surface displacements can be obtained

in ground coordinates (e.g. slant range and azimuth directions), which are finally geocoded using a lookup table derived from the use of the DEM and the image parameter file.

We used a matching window of 320 by 64 pixels (1200 by 1280 m) in range and azimuth directions, with an oversampling factor of 2 for improving the tracking results. The sampling steps were of 40 by 8 pixels and the resolution of the final velocity map was 200 by 160 m in range and azimuth directions. The geocoding was completed using the Canadian Digital Elevation Model. We used a bicubic spline interpolation to generate the geocoded grid. This grid was transformed into a Tiff file using a functionality within GAMMA software. Finally, the determined velocities were manually checked for blunders and mis-matches.

3.3.2 GAMMA Remote Sensing D-InSAR

The application of interferometry and differential interferometry to TOPS mode data has been dealt with by various authors (Grandin, 2015b; Marotti et al., 2011; Prats-Iraola et al., 2012; Scheiber et al., 2015). The D-InSAR technique requires a full co-registration of the reference and slave scenes. For this purpose, we proceed as in the offset tracking case. Generation of wide area surface deformation products is possible thanks to highly accurate ephemerides products (Guarnieri et al., 2015). However, caution should be taken when considering multiple frames along track, making sure to account for the variation of azimuth co-registration error in this direction (Xu et al., 2016). The use of stationary areas is paramount for achieving a successful co-registration. Therefore, we foresee limitations for fine co-registration methodologies in the absence of such areas.

When all of the above-mentioned conditions are met, interferometry is applied in the same way as conventional interferometry to stripmap imagery (i.e. phase filtering, masking and unwrapping). We used the minimum cost-flow algorithm for unwrapping the filtered phase. When calculating the differential interferogram by using a DEM, the derived differential interferometry should be carefully checked for the presence of phase jumps between subsequent bursts (Scheiber et al., 2015; Wegmüller et al., 2016). For the computation of the differential interferometric product we used sampling steps of 10 by 2 pixels (50 by 40 m) in the range and azimuth directions respectively. The geocoding was completed using the Canadian Digital Elevation Model. We used a bicubic spline interpolation to generate the geocoded grid. This grid was transformed into a Tiff file using a functionality within GAMMA software.

3.3.3 Application to Ascending and Descending passes

Offset tracking case

In a study by other authors aimed at obtaining the full 3-D displacement vector from ascending and descending passes using offset tracking (Fallourd et al., 2010), all the offsets (both in range and azimuth) were used within a least-squares approach for calculating the velocity vectors.

We avoid using the azimuth offsets altogether, for two main reasons. First, the resolution of Sentinel-1 IW images in the azimuth direction is four times worse than in the range direction. Second, we avoid the ionospheric effect, which degrades substantially the performance of the offset tracking algorithm in the azimuth direction creating the so-called azimuth streaks (Gray et al., 2000; Wegmüller et al., 2006). The ionospheric effect will have a stronger influence on Sentinel-1 datasets due to the shorter repeat cycle of the constellation (which implies less amount of glacier movement in the retrieved signal). As in the case of D-InSAR, we will obtain a 2-D velocity vector by applying a transformation from a nonorthonormal basis to the corresponding geographical coordinate system (Joughin et al., 1998). By doing so, we tailored the existing methodologies to the special characteristics of Sentinel-1 TOPS product.

The horizontal velocity vector $\mathbf{v}_h$ can be expressed as a function of the ascending and descending across-track vectors (see Figure 3.1d)

$$\mathbf{v}_h = v_a \hat{\mathbf{a}} + v_d \hat{\mathbf{d}}. \quad (3.1)$$

We identify the range displacements from the ascending and descending passes as the first and second terms on the right hand side of Equation (3.1). At this stage, we used the intensity offset tracking range displacements obtained with GAMMA software and described in section 3.1. The main problem in the transformation is that the basis $(\hat{\mathbf{a}}, \hat{\mathbf{d}})$ is nonorthonormal. Therefore, a relation between the projections of the horizontal velocity vector and the ad coordinates needs to be established. These relations are given in (3.2) below, where the matrix form is also displayed.

$$v_a = \left(\frac{\mathbf{v}_h^T \hat{\mathbf{a}} - \mathbf{v}_h^T \hat{\mathbf{d}} \cos \alpha}{\sin^2 \alpha} \right) v_d = \left(\frac{\mathbf{v}_h^T \hat{\mathbf{d}} - \mathbf{v}_h^T \hat{\mathbf{a}} \cos \alpha}{\sin^2 \alpha} \right) \mathbf{B} = \left(\frac{1}{\sin^2 \alpha} \right) \begin{bmatrix} 1 & -\cos \alpha \\ -\cos \alpha & 1 \end{bmatrix}. \quad (3.2)$$

The first two expressions in (3.2) are inferred manipulating the equations that describe the two dimensional transformation between the ascending and descending coordinate systems

(Joughin et al., 1998). Next, we transform the projected velocity vector v_h from the ad coordinate system to xy coordinates (Eq. 3.3).

$$\begin{aligned} \hat{\mathbf{a}} &= \cos \beta \hat{\mathbf{x}} + \sin \beta \hat{\mathbf{y}} \\ \hat{\mathbf{d}} &= \cos(\alpha + \beta) \hat{\mathbf{x}} + \sin(\alpha + \beta) \hat{\mathbf{y}} \end{aligned} \quad \mathbf{A} = \begin{bmatrix} \cos \beta & \cos(\alpha + \beta) \\ \sin \beta & \sin(\alpha + \beta) \end{bmatrix} \quad \begin{bmatrix} v_x \\ v_y \end{bmatrix} = \mathbf{A} \begin{bmatrix} v_a \\ v_d \end{bmatrix}. \quad (3.3)$$

Finally, Equation (3.4) shows the application of the whole transformation from the nonorthonormal across-track vectors to the xy coordinate system

$$\begin{bmatrix} v_x \\ v_y \end{bmatrix} = \mathbf{AB} \begin{bmatrix} \mathbf{v}_h^T \hat{\mathbf{a}} \\ \mathbf{v}_h^T \hat{\mathbf{d}} \end{bmatrix}. \quad (3.4)$$

The α and β angles necessary to perform the transformation were obtained using the GAMMA software. The program provides a function to calculate the look-vector orientation producing an output grid with the SAR look-vector orientation as a function of position. This grid was geocoded back to the geographical coordinate system. Once the rasters of the look-vector direction from the ascending and descending images were obtained, we had all the necessary information to proceed with the computation of the velocity components. The raster algebra to transform the coordinate system ad to the geographical coordinate system xy was performed entirely using QuantumGIS software and an ad-hoc code to automatize the processing with Python. The histograms and plots were completed using Octave.

D-InSAR case

The retrieval of surface displacement from ascending and descending SAR pairs has been covered by several authors (Joughin et al., 1998; Kumar et al., 2011; Mohr et al., 1998). In all cases, for retrieving a 3-D full vector an assumption of surface parallel flow was made. The downside of this approach is that ice sometimes does not flow parallel to the surface. However, when considering only the interferometric phase difference due to the horizontal displacement, one can derive the horizontal 2-D velocity vector from ascending and descending passes. The important step in this procedure is the transformation from a nonorthonormal basis (the across-track ascending and descending vectors) to the geographical coordinate system (Joughin et al., 1998). We proceeded in the same way as for the intensity offset tracking case. The D-InSAR across-track ascending and descending horizontal displacements were obtained using GAMMA software. The angles α and β were obtained in the same way as for the intensity offset tracking case. The raster algebra was done with QuantumGIS software. By using this approach, we do not readily get the full three-dimensional velocity vector, but

the missing vertical component can be easily obtained afterwards using the surface gradient. Moreover, the vertical component is generally small.

3.4 Results and Discussion

3.4.1 Offset tracking Velocity Mapping of Southern Ellesmere's Ice Caps

In general, our results show that the performance of the amplitude matching algorithm is good, and the method is able to resolve the glacier velocities in most areas. There is a large variability in the observed velocities, due to the different dynamic behaviour of the various glaciers. Some glaciers present a surge-type behaviour (Copland et al., 2003a) or high seasonal and interannual variability of their velocities (Williamson et al., 2008). We can broadly group our studied glacier into three groups: 1) fast flowing with winter velocities greater than 600 m yr^{-1} ; 2) medium flowing with winter velocities within $50\text{-}200 \text{ m yr}^{-1}$; and 3) slow flowing with speeds less than 50 m yr^{-1} . Trinity and Wykeham glaciers (Fig. 3.2) belong to the fast flowing group with speeds up to 1200 m yr^{-1} for Trinity and 600 m yr^{-1} for Wykeham. To the medium flowing group belong glaciers such as Ekblaw Glacier (Fig 3.1c) or neighbouring glaciers such as Stygge or Cadogan (north and south of Ekblaw, respectively; not shown in figure), reaching winter velocities of 80 and 60 m yr^{-1} , or the southern glacier of Manson Icefield, with speeds up to 75 m yr^{-1} . This is currently the fastest glacier of the Manson Icefield because Mittie Glacier, which can have velocities greater than $1\,000 \text{ m yr}^{-1}$ when in full surge (Copland et al., 2003a), is currently in its quiescent phase (van Wychen et al., 2016). Most of the slow flowing glaciers are land-terminating glaciers located on the western side of the POW Icefield (Fig. 3.3) and on Sydcap Ice Cap. However, some of the land-terminating glaciers in the western POW Icefield have larger velocities, as happens with unnamed West 3, 4 and 5 (Supplementary Materials Fig. A1²), with velocities up to 75 , 150 and 100 m yr^{-1} respectively. These high velocities are due to the narrowing of the valleys confining the glacier flow. In the case of unnamed West 4, the velocity increase is more noteworthy because of its larger accumulation area.

However, the feature tracking algorithm is not able to resolve the surface velocity in slowest-moving zones of the ice cap's accumulation areas (Fig. A1) and in some zones of Trinity glacier (Fig. 3.2). In the case of the accumulation areas, this can be attributed to the absence of visible features due to the snow cover. In the case of Trinity Glacier, the underlying reason is the strong surface velocity gradients that deform the surface features

²Figures and Tables with a letter before the numbering are located in the appendices

Fig. 3.2 Ice velocities for Trinity and Wykeham glaciers obtained using intensity offset tracking applied to ascending and descending passes. The zone shown corresponds to the red rectangle labelled II in panel b of Figure 3.1. Velocities calculated from images acquired between 22nd February and 12th March 2016. Velocity map overlaid on top of Sentinel-1 SAR images acquired between 22nd and 29th February 2016.

(e.g. at the junction between the main glacier trunk and its principal tributaries, at the corners where the glacier changes direction or at the glacier margins.

The feature tracking algorithm is best suited for resolving areas of fast glacier movement. This is clear when comparing the feature tracking results with those from differential interferometry (Figures A1 and A2). However, there is also a limit on the amount of movement that the feature tracking algorithms are able to resolve. Nagler et al. (2015), using an in-house developed algorithm, were able to resolve velocities up to 12 m d^{-1} when using Sentinel-1 IW data (the same sensor that we are using). With increasing image resolution (smaller pixel size), larger velocities could be resolved. Anyway, this restriction is not relevant to us, because 12 m d^{-1} is equivalent to $4\,380 \text{ m yr}^{-1}$, a velocity well above the largest glacier velocities observed in the Canadian Arctic.

We turn now our attention to the performance of offset tracking using ascending and descending tracks without using the azimuth offsets. As expected, the areas of stable ground do not present the typical azimuth streaks that one would expect when using azimuth offsets

(Gray et al., 2000; Wegmüller et al., 2006). In the scenes that we used, the orientation of the ascending and descending across-track vectors was close to the horizontal (e.g. 12° below the west-east direction for the ascending case). This makes more challenging the retrieval of surface velocities for glaciers oriented in the north-south direction. Nevertheless, glaciers oriented in this direction (e.g. most of the tributaries of Ekblaw Glacier) still present homogeneous surface velocities (Fig. 3.1c). Furthermore, when analysing the surface velocity gradients of Ekblaw Glacier in the northern direction (Fig. 3.1c), it is noticeable the better performance of the our approach (not using azimuth offsets), since we can observe how the northern component of the velocity increases and decreases depending on the direction of glacier flow, while, when using the standard approach that uses azimuth offsets, we were unable to resolve the velocity gradients for this glacier. Also note that the standard approach requires the correction of the ionospheric effect, which increases the complexity when considering Sentinel-1 TOPS acquisition mode. This is because each sub-swath presents its own ionospheric-induced azimuth pattern. When no information on the total electron content of the atmosphere is available, we need to have available stable ground in different sub-swaths, a condition not easily fulfilled in many glacierized areas. Furthermore, when using a modelled trend of the ionospheric disturbance for subtracting the ionospheric effect, residues always remain that degrade the azimuth offset results.

3.4.2 D-InSAR Velocity Mapping of Southern Ellesmere's Ice Caps

Glacier surface velocity results using ascending and descending passes show good performance on all land-terminating glaciers, such as the western glaciers of POW Icefield (Figures 3.3a and A2) and the eastern glaciers of Sydcap Ice Cap. Moreover, there are two tidewater glaciers that also show good results when applying this technique, namely the South Margin glacier (POW Icefield) and Mittie East and West arms (Manson Icefield), as well as their tributary glaciers. All of them present surface velocities below 50 m yr^{-1} and smooth surface velocity fields.

D-InSAR performs better than offset tracking in areas with low glacier velocities. Furthermore, it is able to resolve small surface velocity gradients. The low amount of precipitation in this region (Alt, 1987; Gardner and Sharp, 2007) allows for a successful application of D-InSAR in the region's ice mass accumulation areas. This is the case of POW Icefield, where we see a fine and continuous ice velocity field. It is precisely this fine resolution what allows us to discern the ice divide between the eastern and western flowing glaciers (Fig. A2). We note that in some cases the Randolph Glacier Inventory (RGI) (Pfeffer et al., 2014) ice divides do not match well with our surface velocity field estimates. This fact is noticeable e.g. in Taggart Lake glacier and in the unnamed West 4 and 5 glaciers (Fig. A2). This fine

Fig. 3.3 (a) Ice velocities for the same zone obtained using D-InSAR. The zone shown corresponds to the red rectangle labelled III in panel b of Figure 3.1. (b) Centreline velocities for longitudinal section A-A' in panels a, c. (c) Ice velocities using intensity tracking from the western glaciers of Prince of Wales Icefield. (d) Centreline velocities for longitudinal section B-B' in panels a, c. The velocity maps are overlaid on top of Sentinel-1 SAR images taken between 10th and 17th February 2016.

resolution also allows us to recognize different glacier flows (tributaries) in some western land-terminating basins of the POW Icefield. Examples are Taggart Lake glacier, which shows three distinctive flows (one in its northern side and two in the southern one), Unnamed West 2 glacier and Unnamed West 4 glacier (Fig. 3.3a).

Regarding surface velocities for individual glaciers, Unnamed West 1 reaches 24 m yr^{-1} and Taggart Lake glacier reaches 33 m yr^{-1} in their main branches. None of the three different ice flows visible in Unnamed West 2 glacier exceeds 30 m yr^{-1} . The southernmost one shows the slowest velocities, of up to 15 m yr^{-1} , its intermediate branch 12 m yr^{-1} and the northern branch $\sim 21 \text{ m yr}^{-1}$.

3.4.3 Comparison of Offset Tracking and D-InSAR results

If we compare the results from offset tracking and differential interferometry (Fig.3.3), it is noticeable the more irregular velocity patterns produced by the offset tracking technique (e.g. there is some degree of variability on stable ground). Additionally, we perceive a lower resolution of the technique, either due to the size of the matching window or to the quality of the amplitude features. This lower resolution prevents the offset tracking technique from producing the continuous and clearly defined glacier surface flow fields produced when the differential interferometry technique is used. However, the offset-tracking technique allows us to derive region-wide results with less constraints as compared with other methodologies (e.g. resilience to de-correlation or fast surface velocities).

Differential interferometry produces a high-quality, continuous and smooth velocity field. The main shortcomings of this technique include the inability of the minimum cost flow algorithm to resolve strong surface gradients and its deterioration with longer time spans between acquisitions (e.g. increased de-correlation). The former situation would be exemplified by the case of the glacier tongues of glaciers Unnamed West 3, Unnamed West 4 and Taggart Lake north whose velocities cannot be retrieved with the D-InSAR algorithm (Fig. 3.3). Offset tracking, on the contrary, is capable of capturing these type of velocity gradients, making this methodology especially useful to retrieve region-wide surface velocity fields.

The areas where the differential interferometry performs best are those with subtle transitions between parallel or adjacent surface flows or in accumulation areas with lower velocities. In such areas we notice the difference between offset tracking and D-InSAR. While the former hardly resolves the velocity gradients (and, when it does, it produces a nonsmooth field; see the accumulation area of Taggart Lake glacier), D-InSAR results illustrate the capabilities of this technique in accumulation areas with low amount of precipitation, which helps to preserve the correlations between subsequent acquisitions.

3.4.4 Comparison with previous studies

We have produced regional velocity estimations for the Southern Ellesmere Island ice caps using intensity offset tracking and differential interferometry from Sentinel-1 acquisitions. What is special in our application of the intensity offset tracking technique is that we retrieved the surface velocity fields exclusively from range offsets using ascending and descending passes. We also applied differential interferometry for the same areas in order to establish a comparison on the performance of both methodologies in glaciers with different dynamic regimes (Copland et al., 2003a; Williamson et al., 2008). Sentinel-1 IW TOPS mode data was already tested with the offset tracking methodology by Nagler et al. (2015) giving successful region-wide results for the ice velocity field of Greenland. The application of the D-InSAR technique to TOPS mode acquisitions from ice caps is still a challenge due to the need for stable ground to achieve a fine co-registration with the spectral diversity method, aimed to avoid undesired azimuth phase ramps in the resulting products (Grandin, 2015b; Prats-Iraola et al., 2012; Scheiber et al., 2015).

The application of the intensity offset tracking approach to TOPS mode acquisitions was already dealt by Dall et al. (2015), who acknowledged the need for azimuth deramping and azimuth common band filtering when applying speckle tracking. These two latter steps are not required in the feature tracking that we apply. The application of the intensity offset tracking technique to ascending and descending passes was covered by Fallourd et al. (2010). However, their approach considered the use of all azimuth and range velocity components solving by least-squares, while our approach discards the azimuth offsets from the estimation to avoid undesired ionospheric effects and, simultaneously, to avoid Sentinel-1 shortcomings such as its lower azimuth resolution. We did this by transforming a nonorthonormal coordinate system (e.g. ascending and descending across-track vectors) to a geographical coordinate system (Joughin et al., 1998).

The results of the application of the proposed method show that there is a noticeable improvement in the resolution of the components of the surface velocity together with an increase of the final precision. We avoid the ionospheric effects and therefore, considering the improved accuracy of the ephemeris, the remaining error sources are restricted to the co-registration, the matching procedure and the DEM related geocoding error. Also, there are still some de-correlated areas in the accumulation zone devoid of results due to the lack of features for tracking. We would like to highlight that using azimuth offsets with Sentinel-1 TOPS data not only worsens the final velocity product but also increases the difficulty of correcting the azimuth streaks which present a different pattern for each sub-swath.

There are several articles devoted to the topic of interferometry using TOPS mode data (Marotti et al., 2011; Prats-Iraola et al., 2012). Furthermore, the use of ascending and

descending passes for retrieving ice surface velocities is also well covered in the literature (Joughin et al., 1998; Kumar et al., 2011; Mohr et al., 1998). The main obstacle that we faced when applying D-InSAR to TOPS mode data was to guarantee a successful fine co-registration using the spectral diversity method between contiguous bursts. The choice of Ellesmere Island as a test site was key for overcoming this limitation, because of the wide areas of stable ground that are found next to the ice masses.

Our results show that differential interferometry is optimal for areas of low movement with no (little) surface change (e.g. accumulation areas with low amount of precipitation, and land terminating glaciers). The method produced smooth velocity fields for most land terminating glaciers of Ellesmere southern ice caps. On the other hand, the method failed to work in areas with strong velocity gradients and areas of fast glacier movement (e.g. the minimum cost flow (MCF) algorithm did not resolve a continuous velocity field (Joughin et al., 1998)). The comparison between methodologies illustrates the difficulties of MCF to resolve the velocity field of a few tongues from the western glaciers of the POW Icefield.

Regarding the comparison with velocities presented in previous studies in the region, focusing on those based on scenes closest in time to ours, we note that the velocities are very similar, especially for the fastest glaciers. Our maximum winter (Feb-March 2016) velocities of 1200 m yr^{-1} near the terminus of Trinity Glacier or 600 m yr^{-1} for Wykeham Glacier are very similar to those of 1250 and 500 m yr^{-1} (respectively) reported by van Wychen et al. (2016) for winter 2014-2015 (from speckle tracking of RADARSAT-2 data), or those of 1200 and 650 m yr^{-1} (respectively) given by Millan et al. (2017) for winter 2015-2016 (from speckle tracking of Sentinel-1a data). Similarly, for Ekblaw Glacier both our results and those of van Wychen et al. (2016) show similar maximum velocities of $\sim 100 \text{ m yr}^{-1}$.

3.4.5 Error Estimates

Offset Tracking case

The three main sources of error in the offset tracking technique are: 1) the matching procedure (which is a function of the co-registration between images, template size, and the quality of the image features) (Nagler et al., 2015); 2) the ionospheric effect and its influence on the azimuth offsets in the form of azimuth streaks (Gray et al., 2000; Wegmüller et al., 2006); and 3) the geocoding error (with the high quality ephemerides of Sentinel-1 data this error is reduced to the quality of the DEM used for the topographic correction (Nagler et al., 2015)).

The first type of error is obviously unavoidable due to the intrinsic nature of the tracking algorithm. However, the ionospheric errors account for a big share of total velocity errors, providing a possible and effective way of reducing offset tracking errors. Sentinel-1 is

Fig. 3.4 Glacier surface velocity errors from traditional intensity offset tracking technique (average 14.7 m yr^{-1}) and from intensity offset tracking applying ascending and descending passes (average 3.3 m yr^{-1}).

equipped with a C-band sensor which is less affected by this disturbance than L-band sensors such as those on board of ALOS PALSAR (Gray et al., 2000). L-band sensors, however, have higher penetration depth in snow and ice, which improves correlation between scenes and therefore also interferometric results, as has been analyzed by Rignot et al. (2001). On the other hand, Sentinel-1 repetition cycle of 12 days has the side effect of increasing the share of signal corresponding to ionospheric effects. Various authors have acknowledged the influence of the ionospheric effect on the offset tracking results. Nagler et al. (2015) estimated this effect to account for $\sim 0.08 \text{ m d}^{-1}$ displacement when considering the low velocity areas of Greenland. This estimate dropped to 0.02 m d^{-1} when averaging over several acquisitions. Other authors (Gray et al., 2000; Mouginot et al., 2012), when evaluating RADARSAT-1 offset tracking data from Antarctica, estimated this error to be between 0.042 and 0.016 m d^{-1} .

We performed an analysis of our results in Ellesmere Island, focusing on the performance of the algorithm on stable ground (e.g. on ice-free areas on the western side of POW Icefield) where both matching and ionospheric components of the error are present and can be quantified. We calculated the root mean square error (RMSE) of the normal intensity offset tracking results (e.g. including azimuth offsets and the related ionospheric effect)

and also of our suggested methodology using exclusively range offsets from ascending and descending passes. The results show that the former approach casts a RMSE value of $\sim 0.051 \text{ m d}^{-1}$ (the mean error being $\sim 14.2 \text{ m yr}^{-1}$) while the latter drops to 0.012 m d^{-1} (the mean error being $\sim 3.3 \text{ m yr}^{-1}$) (Fig. C.3). If we consider that both sources of error are independent, the share of error that could be attributed to ionospheric disturbances in the first case amounts to $\sim 0.05 \text{ m d}^{-1}$, a value falling within those of Nagler et al. (2015) and Gray et al. (2000) estimates. We emphasize that, when using only range offsets extracted from ascending and descending passes with Sentinel-1 data, errors represent a fourth of the intensity offset tracking technique error budget. Furthermore, our technique still improves the multiple acquisition averaging, by halving the error of the latter.

Differential Interferometric case

The InSAR technique has several error sources. As for intensity offset tracking, the co-registration is a typical error source. For instance, the co-registration of the ascending and descending scenes using the DEM could have significant errors. These errors could be due to a low quality DEM, to shortcomings in the along-track timing or to inaccurate ephemeris (Joughin et al., 1998; Rosen et al., 2000). The largest share of the error corresponds to the interferometric baseline definition, whose influence can be observed as linearly varying errors. Sentinel-1 data is no longer affected by these two types of orbit-related errors because of its high quality orbital data (Guarnieri et al., 2015; Nagler et al., 2015). The phase unwrapping algorithm plays a role in a correct and continuous definition of surface velocity gradients.

In differential interferometry, a DEM is used to remove the topographic phase component from the interferogram. Therefore, a low quality DEM combined with long baseline acquisitions can also lead to worsened velocity estimates (Kumar et al., 2011).

Most glaciers of this region have directions approximately perpendicular to the along-track satellite direction (especially all the western POW Icefield glaciers). This optimizes the application of interferometry to these glaciers. All the above mentioned elements contribute with different shares to the total error budget. This makes it difficult to give a proper error estimate. Nonetheless, Luckman et al. (2007) gave an estimate of the error between 0.012 and 0.017 m d^{-1} for the specific case of ERS-1 satellite data applied to the Himalayas. We consider our performance to be better both because of Sentinel-1 improved orbital data and to the more gentle topography of Ellesmere Island and its positive impact on the DEM.

3.5 Summarizing Conclusions and Outlook

The proposed intensity offset tracking algorithm applied to ascending and descending passes, which disregards the azimuth offsets, has demonstrated its merits to resolve velocity gradients in the along-track direction, improving the resolution of the offset tracking and, at the same time, avoiding the ionospheric disturbances on the data. In this way, we have improved the accuracy of the derived surface velocity product, whose main source of error is reduced to the matching-related uncertainties. The D-InSAR interferometry has also shown a good performance in slowly moving areas, providing high resolution, smooth and continuous velocity fields for land-terminating glaciers.

In view of these results, we acknowledge the complementarity of both techniques (Hu et al., 2014; Luckman et al., 2007). Offset tracking can be applied on areas of faster glacier movement and velocity gradients, with some limitations, while differential interferometry is suited to areas with low velocities and small velocity gradients. Moreover, offset tracking is less affected by de-correlation than differential interferometry. We see here an opportunity to develop a hybrid velocity product combining DInSAR and offset tracking results in regions where either one method or the other, or both, perform best, following the lines suggested by Joughin (2002) and Liu et al. (2007). The recently deployed Sentinel-1B satellite will improve the interferometric capabilities of the Sentinel-1 constellation by reducing the de-correlation and the amount of movement between consecutive acquisitions (Joughin et al., 1998). Furthermore, we see a possible application of the ascending and descending passes to speckle tracking. This approach would allow to obtain surface velocity fields within featureless accumulation areas, while avoiding the ionospheric effects typical of Arctic and Antarctic areas (Nagler et al., 2015).

Chapter 4

Ice discharge error estimates using different cross-sectional area approaches: a case study for the Canadian High Arctic, 2016-2017¹

4.1 Introduction

Frontal ablation, that is, ice mass losses by calving, subaerial frontal melting and sublimation and subaqueous frontal melting (Cogley et al., 2011), is an important component of the mass balance of tidewater glaciers and marine-terminating ice caps. It has been reported to account for up to 30-40% of the total ablation of some Arctic glacierized archipelagos and ice caps (Błaszczyk et al., 2009; Dowdeswell et al., 2002, 2008) and up to 50% in some ice caps in the Antarctic periphery (Osmanoğlu et al., 2014). Because of the difficulty of calculating the components of frontal ablation separately, it is usually approximated by the ice discharge through flux gates close to the calving fronts, calculated as the product of ice velocity and cross-sectional area (volumetric flux). If the considered flux gate is not close to the calving front, the surface mass balance between them should be taken into account (Andersen et al., 2015; McNabb et al., 2015). Neglecting this effect will in nearly all cases lead to an overestimation of the frontal ablation. For Alaskan glaciers, McNabb et al. (2015) found an overestimate of 19% on average for individual glaciers (10% for the regional total). For Canadian Arctic glaciers the overestimate is expected to be larger, because of the strongly negative recent surface mass balance of this region (Gardner et al.,

¹(Sánchez-Gómez and Navarro, 2018)

2011). Possible front position changes should also be taken into consideration (Burgess and Sharp, 2004; Burgess et al., 2005; van Wychen et al., 2016; Williamson et al., 2008), though the terminus advance/retreat is in general expected to account for a smaller share of the frontal ablation estimates. For instance, McNabb et al. (2015) indicated an overestimation of 13% (2% of the regional total) by neglecting advance/retreat of the terminus for Alaskan glaciers. This effect is more difficult to quantify for the Canadian Arctic glaciers due to the pulsating behaviour of many of them (van Wychen et al., 2016, 2017; Williamson et al., 2008). Pulsating glaciers, as surging glaciers, show periods of speedup and slowdown. However, the velocity variability of pulsating glaciers is restricted to their lowermost terminal zone, which is grounded below sea level. Additional factors influencing frontal ablation estimates, such as cross-sectional ice-thickness variations and long-term thickness changes have been discussed e.g. by McNabb et al. (2015). The seasonality of the glacier velocity measurements can also have a noticeable impact on frontal ablation estimates, with seasonal velocity amplitudes averaging $\sim 50\%$ of the peak velocity for Alaskan glaciers (McNabb et al., 2015) and $\sim 45\%$ in Livingston Island, off the northwestern Antarctic Peninsula (Osmanoğlu et al., 2014). The changes for the Canadian Arctic glaciers seem to be more modest, with a seasonal variability of $\sim 10\text{--}20\%$ of the average values (van Wychen et al., 2012). The interannual variations of glacier velocity are also an important issue, particularly for the Canadian Arctic glaciers due to their pulsating nature. Estimates of total error in frontal ablation accounting for the various error sources described above are typically within $\sim 20\text{--}30\%$. For instance, values of 24% and 25% have been given for some Alaskan glaciers by McNabb et al. (2015) and Vijay and Braun (2017), respectively, while a value of 31% has been given by van Wychen et al. (2016) for the Canadian High Arctic.

In this paper, we will focus on analysing the errors in the ice discharge estimates through given flux gates, so we will refer to ice discharge rather than frontal ablation. A fundamental problem for estimating the ice discharge is that very often the cross-sectional area of tidewater glaciers is unknown. Normally, there is a lack of information regarding the thickness of glaciers. Typically, glacier thickness is only known along the central flowline, sometimes on the glacier's cross section and very rarely both of them are available for a specific glacier (Leuschen et al., 2010). This scarcity in ice-thickness data motivates the use of U-shaped cross-section profiles for ice discharge estimates when only ice-thickness data along the glacier flowline is available (van Wychen et al., 2012). Inverse modelling is used when we count on little ice-thickness data, or no data at all (Farinotti et al., 2009). Observed thickness data can be assimilated into inverse methods (Osmanoğlu et al., 2013, 2014).

Different approaches have been taken in the literature to estimate the error in ice discharge through flux gates. However, most of them only provide upper and lower bounds of the

error (Burgess et al., 2005; van Wychen et al., 2012), rather than statistical estimates of the expected error with a certain degree of uncertainty. In this paper we aim to contribute to fill this gap by analysing, using error propagation, the various error sources involved in the estimate of ice discharge, and to quantify their individual contributions to the total error in ice discharge.

When cross-sectional ice-thickness measurements are not available, ice-discharge estimates through given flux gates can be based on an increasingly complex approach, ranging from the so-called box-shaped approaches (Błaszczuk et al., 2009; Brown et al., 1982; Burgess et al., 2013; Rignot et al., 2008; Williamson et al., 2008) to those considering a varying cross-sectional depth (van Wychen et al., 2014, 2015, 2012, 2016, 2017). The box-shaped approach shows a tendency to overestimate the ice discharge, so we will not use it in this paper. Various U-shaped approaches can be taken, and a detailed analysis of their performance, on the basis of the comparison of their results with those for a large set of observed cross-sectional areas, is lacking in the literature. Filling this gap is the second aim of the present paper.

The two above analyses will be based on ice discharge estimates for the Canadian High Arctic using ice-thickness data and remotely sensed glacier surface-velocity data from 2012-2015 and 2016-2017 respectively. This will provide, as a by-product of this paper, updated ice-discharge estimates for glaciers of the Canadian High Arctic during 2016-2017.

4.2 Study site and data

The Queen Elizabeth Islands (Ellesmere, Axel Heiberg and Devon islands) are located in the Canadian Arctic, neighbouring the western coast of Greenland (Fig. 4.1a). This region is often referred to as Canadian High Arctic. The ice masses of Ellesmere Island contain one-third of the global volume of land ice outside Greenland and Antarctica (Radić and Hock, 2010) with a glacierized area in 2000 of ca. 104 000 km² (Sharp et al., 2014) while Devon Ice Cap covers approximately 14 000 km² and is one of the largest ice caps in the Arctic (Dowdeswell et al., 2004). The dynamics of Canadian Arctic glaciers and ice caps have been extensively studied (Burgess et al., 2005; Copland et al., 2003a; Dowdeswell et al., 2004; Gardner et al., 2011; Millan et al., 2017; Strozzi et al., 2017; van Wychen et al., 2014, 2015, 2012, 2016; Williamson et al., 2008). Copland et al. (2003a) made an analysis on the surge-type glaciers in this region and later studies by van Wychen et al. (2016, 2017) developed a classification scheme into three glacier types depending on their dynamics: surging, pulsing and consistent acceleration/deceleration. These investigations indicate that

Fig. 4.1 a) Main ice masses of Ellesmere, Axel Heiberg and Devon Islands, Canadian High Arctic (Wessel and Smith, 1996). The glacier outlines are from the Randolph Glacier Inventory (RGI) version 5.0 (Pfeffer et al., 2014). See more detail in Supplementary Materials figures B1, B2 and B3, and Table B1. b) Cross-sectional profile approaches overlaid on a real glacier cross section. All maps in this paper use the Universal Transverse Mercator (UTM) projection of the zone 17 north, and the reference ellipsoid is WGS84.

Table 4.1 Operation IceBridge airborne radar profiles used in this study. *Cross* means cross-sectional profiles and *Long* means longitudinal (along-flow, close to glacier centreline) profiles.

Data Location	Profile Type	Year
Prince of Wales Icefield	Cross(10)-Long(8)	2012/2014
Devon Ice Cap	Cross(8)-Long(6)	2012-2015
Northern Ellesmere Icefield	Cross(2)-Long(8)	2014
Agassiz Ice Cap	Long(8)	2014
Müller and Steacie Ice Caps	Long(2)	2014
Manson Icefield	Cross(2)	2012
Sydkap	Cross(1)-Long(1)	2012

the glaciers of the Canadian High Arctic show marked differences in dynamic behaviour, and large spatial and temporal variabilities.

The datasets used in this study encompass Synthetic Aperture Radar (SAR) images from the Sentinel-1 platform, Operation IceBridge airborne radar ice thickness (Gogineni, 2012; Leuschen et al., 2010) and the freely accessible Canadian DEM (CDEM) designed by Natural Resources Canada (NRCAN) with a resolution of 0.75 and 3 arc second in the south-north and west-east directions respectively (NRCAN, 2016). The DEM was used for geocoding and co-registering the imagery employed for the intensity offset tracking technique. ArcticDEM from WorldView satellite, from 2012 and 2014-2015, was used for estimating local ice-thickness changes from surface-elevation changes, assuming no glacier-bed elevation change.

The surface velocities on the Canadian High Arctic were obtained from SAR Terrain Observation by Progressive Scans (TOPS) Interferometric Wide (IW) Level-1 Single Look

Complex (SLC) images. These products provide a geo-referenced image (using accurate altitude and orbital information from the satellite), in slant-range geometry, processed in zero-Doppler (Geudtner et al., 2014b). The image is normally composed of three sub-swaths with each sub-swath comprising normally 9 consecutive bursts, which overlap in azimuth. Burst synchronisation is needed for interferometry and for accurate offset tracking (Holzner and Bamler, 2002). The resolution of Sentinel-1 SAR TOPS IW mode is of 5 and 20 m in the range and azimuth directions respectively. The SAR images used in this study were acquired during the winter of the year 2016 (beginning of February until mid March) and the winter of the following year (end of January to mid March 2017). See more detail in Supplementary Materials Table B2.

The ice-thickness dataset lumps a wide variety of data regarding both acquisition dates and geographical distribution (Table 4.1). In particular, we count on transverse ice-thickness profiles for 23 glaciers, and 20 additional glaciers for which only along-flow, close to centreline ice-thickness profiles are available, totalling 43 studied glaciers in the area. Ice-thickness data were measured using the Multichannel Coherent RADAR Depth Sounder (MCoRDS) (Leuschen et al., 2010). Specifically, the dataset used in this study is the post-processed L2, which contains information about time, latitude, longitude, elevation, glacier surface and bed elevation, and ice thickness, the latter one with an estimated uncertainty of ± 10 m (Gogineni, 2012).

4.3 Methodology

We calculate the ice discharge using a flux-gate approach for two cases: 1) the cross-sectional depth profile is known (Vijay and Braun, 2017), and 2) the cross section is estimated using an approximation to the depth profile (Cuffey and Paterson, 2010; Harbor, 1992). For the latter case, we will use three different approaches, a centred parabola for which the ice thickness at the glacier centreline is assumed to be known, a centred parabola with ice thickness known at an off-centred point and a quartic function with ice thickness known at an off-centred point. The availability of airborne radar cross-sectional profiles for 23 Canadian High Arctic glaciers, which will be taken as reference, will allow us to compare the accuracy of the various cross-sectional area approaches.

4.3.1 Ice discharge

Ice discharge is calculated as mass flux per unit time across a given surface S approximated per area bins as

$$\phi = \int_S \rho \mathbf{v} \cdot d\mathbf{S} = \sum_i \rho L_i H_i f v_i \cos \alpha_i, \quad (4.1)$$

where ρ is ice density, L_i and H_i are respectively the width and thickness of an area bin, f is the ratio of surface to depth-averaged velocity, $f \in [0.8, 1]$ (Andersen et al., 2015; Cuffey and Paterson, 2010; Thomas et al., 2000), v_i is the magnitude of surface velocity and α_i is the angle between the surface velocity vector and the direction normal to the local flux-gate for the bin under consideration. When airborne radar cross-sectional profiles are available, we define the bin widths and orientations as given between consecutive radar-measured points; when no cross-sectional airborne radar profiles are available, we use bins of fixed width and thickness estimated from the corresponding cross-sectional profile approach, and the velocity vector orientations are calculated with respect to the vector normal to the cross section.

The above-described approach for the calculation of ice discharge is valid for grounded sections of tidewater glaciers. The analysis of the available radar profiles, together with the flotation criterion, indicates that a few of the studied glaciers have floating tongues or are close to flotation, in agreement with Williamson et al. (2008). However, in all cases we have calculated the fluxes at locations where the glaciers are grounded.

4.3.2 Cross-sectional profile approaches

When airborne radar data is only available along a profile close to the glacier central flowline, the set of points to be used for approximating the cross-sectional thickness profile is limited to three points: the intersection of the longitudinal radar profile with the selected cross section and the intersection of the cross section with the glacier margins. In the two latter points, zero ice thickness is most often assumed.

We consider in our analysis three different approaches to the cross-sectional area of a tidewater glacier (Fig. 4.1b). The first approach is the one used by van Wychen et al. (2014, 2016), who used a parabola with axis at the glacier centreline, assuming that the radar-measured ice thickness H_m corresponds to the glacier centreline (Fig. 4.2 with $H_m = H_c$) and that the ice thickness at both margins, $x = \pm W$, is of 10 m; for short, we will refer to this approach as *centred parabolic*. The ice thickness H at a point situated at a distance x from the glacier centreline is then given by

$$H(x) = \frac{10 - H_m}{W^2} x^2 + H_m, \quad (4.2)$$

where we have renamed the variables and parameters used by van Wychen et al. (2014, 2016) as follows: $H_m = C$, $W = D_1$ and $x = D_2$.

Fig. 4.2 Geometry of the U-shaped cross-sectional approaches used in this study. The blue line represents the actual cross-profile of Glacier North 3 (Fig. B3) from NASA Operation Ice-Bridge data acquired the 4th of May 2012, while the red line represents its U-shaped approximation. H_m is the radar-measured ice thickness and H_c is the ice thickness at the glacier centreline. W is the glacier halfwidth.

The second approach, also of parabolic type and with axis located at the glacier centreline, differs in that H_m is not assigned to the centreline but is located at a distance d from it (Fig. 4.2). This is more realistic as it represents the possibility that the radar longitudinal profile could have an offset from the glacier centreline. In this case, we have assumed zero ice thickness at the margins. For short, we will refer to this approach as *off-centred parabolic*. The equation considered is of the type

$$H(x) = ax^2 + b, \quad (4.3)$$

which, upon applying the constrain that the parabola passes through the points $(W, 0)$ (or $(-W, 0)$) and (d, H) , where W represents the glacier half-width, becomes

$$H(x) = \frac{H_m}{W^2 - d^2}(W^2 - x^2). \quad (4.4)$$

When $d = 0$, i.e. the radar flight line coincides with the glacier centreline, the latter equation reduces to Equation (4.2), except for the addend 10.

The third approach is similar to this second one except that the function is now a quartic polynomial of the type

$$H(x) = ax^4 + b, \quad (4.5)$$

which, again, upon the constraint of passing through the points $(W, 0)$ and (d, H) , becomes

$$H(x) = \frac{H_m}{W^4 - d^4}(W^4 - x^4). \quad (4.6)$$

For short, we will refer to this approach as *off-centred quartic*.

4.3.3 Intensity offset-tracking velocities

We applied SAR offset tracking algorithm in *GAMMA* Remote Sensing software in order to produce ice-surface velocity fields in range and azimuth directions from Sentinel-1 TOPS IW SLC Level-1 image pairs (Wegmüller et al., 2016). The particularity of the procedure lies in the use of range offsets from ascending and descending passes, avoiding the use of azimuth offsets, since Sentinel-1 data shows a lower resolution in the azimuth direction. Simultaneously, we avoid the undesired ionospheric effect manifested in the data as azimuth streaks (Sánchez-Gómez and Navarro, 2017). We performed the error analysis for Ellesmere Island, focusing on the performance of the algorithm on stable ground (on ice-free areas on the western side of Prince of Wales Icefield). The area under consideration was $\sim 150 \text{ km}^2$, involving approximately 4700 velocity samples. The use of this novel approach allowed us to obtain an improved velocity field with a root-mean-square error (RMSE) of 0.012 m d^{-1} . Nevertheless, we decided to take a conservative error estimate of 0.021 m d^{-1} for the surface velocity field, resulting from one 20th of the range direction resolution as suggested by Strozzi et al. (2002) and Werner et al. (2005).

We used a matching window of 320 by 64 pixels (1200 by 1280 m) in range and azimuth directions, respectively, with an oversampling factor of 2 for improving the tracking results. The sampling steps were of 40 by 8 pixels and the resolution of the final velocity map was 200 by 160 m in range and azimuth directions. The geocoding was completed using the Canadian Digital Elevation Model. We used a bicubic spline interpolation to generate the geocoded grid. The determined velocities fields were manually checked for mismatches in all glacierized areas. These artefacts were removed from the dataset. We paid special attention to those areas where ice discharge was calculated.

4.3.4 Error Analysis

Two types of error estimates for ice discharge are considered in the literature. The first one uses upper and lower bounds for the input values (i.e. velocity and thickness), resulting in upper and lower bounds for the ice flux (Burgess et al., 2005; van Wychen et al., 2012; Williamson et al., 2008). The second approach consists of estimating the statistically expected error using error propagation from the individual error components (Andersen et al., 2015; Gardner et al., 2018; Vijay and Braun, 2017), and requires that all the considered errors are independent and uncorrelated. In our analysis we follow the latter approach.

Case 1: cross-sectional profiles of ice thickness are available

Applying error propagation to Equation (4.1) we get

$$\sigma_\phi = \sqrt{\sigma_{\phi_\rho}^2 + \sigma_{\phi_f}^2 + \sigma_{\phi_H}^2 + \sigma_{\phi_v}^2 + \sigma_{\phi_\alpha}^2}, \quad (4.7)$$

where the various terms represent the contribution to the error in ice discharge due to the uncertainties in density (σ_{ϕ_ρ}), ratio of surface to depth-averaged velocity (σ_{ϕ_f}), ice thickness (σ_{ϕ_H}), velocity modulus (σ_{ϕ_v}) and direction (σ_{ϕ_α}), and each of the terms is of the form (taking σ_{ϕ_H} as an example)

$$\sigma_{\phi_H} = \sqrt{\sum_i (\sigma_H \rho L_i f v_i \cos \alpha_i)^2}. \quad (4.8)$$

In Equation (4.7) we have omitted a term $\sigma_{\phi_L}^2$ because the bin width L_i is assumed to be error-free. As error estimates for the ice thickness and surface velocity we took $\sigma_H = 10$ m (Gogineni, 2012) and $\sigma_v = 0.021$ m d⁻¹ (Sánchez-Gómez and Navarro, 2017; Strozzi et al., 2002). Regarding the ratio of surface to depth-averaged velocity (f), it is considered in the literature that $f \in [0.8, 1]$ (Cuffey and Paterson, 2010). To give a good approximation for this parameter an analysis of the driving stresses present in a glacier and its derived flow regimes would be advisable (Burgess et al., 2005; Dowdeswell et al., 2004; van Wychen et al., 2017). However, simple observation of the glacier surface features could give a hint on the flow regime and therefore help to better constrain f . Normally, tidewater glacier velocity at the terminus is dominated by basal sliding which makes f close to unity. van Wychen et al. (2016), who used the error-bound approach, assumed lower and upper bounds for f of 0.8 and 1, respectively, while Andersen et al. (2015) and Vijay and Braun (2017), who used the statistical-error approach, took $f = 0.93 \pm 0.05$. We have also taken the latter value and error estimate for f . For ice density, we took $\rho = 900 \pm 17$ kg m⁻³. Finally, for calculating σ_α we used a moving window encompassing 10 velocity measurements along the cross section and calculated the standard deviation of their orientations with respect to the normal to the cross section.

The error in ice discharge given by Equation (4.7) assumes that the radar ice-thickness measurement and the velocity data are temporally coincident. This assumption is often not true. In particular, in our case study both data acquisitions are separated by 1-5 years, depending on the glacier, and the glaciers can be expected to have undergone an ice-thickness change during that period. We take this into account by calculating the thinning (or thickening) rate and estimating the change in flux implied by the assumption of simultaneity of airborne radar and SAR acquisitions. This amount constitutes a systematic error (a bias), which we will

represent as $\varepsilon_{\phi \frac{\partial h}{\partial t}}$ and that should be accounted for in our discharge calculations. This factor should not be disregarded, as van Wychen et al. (2016) have pointed out substantial thinning rates in this region during 1999-2015. These rates are highly variable depending on each particular glacier, and are important even when considering short periods of time because of the pulsating behaviour of many of these glaciers (van Wychen et al., 2016, 2017). The surface elevation change of a tidewater glacier does not always indicate ice thickness change, especially near the glacier fronts where flotation can occur. As discussed earlier, some of the studied glaciers have floating tongues or are close to flotation. However, in all cases we have calculated the fluxes at locations where the glaciers are grounded, so we can safely assume that surface elevation changes correspond to ice thickness changes.

We calculated the thinning rates as difference in surface elevation between the radar data points (Icebridge data collected within 2012-2015, depending on the glacier) and the corresponding points in a suitable DEM, for which we used the ArcticDEM from WorldView satellite, available for 2012 and 2014-2015. To make the computation interval of the thinning rate long enough, we used the pairs radar 2012 with DEM 2014-2015, and radar 2014-2015 with DEM 2012. The thinning rates were afterwards multiplied by the time interval between radar and SAR acquisitions to obtain the $\varepsilon_{\phi \frac{\partial h}{\partial t}}$ correction.

But, in addition to this systematic correction, the random nature of the thickness change (some glaciers thin while others thicken, the thinning rates change with time) requires to consider an error in thickness more conservative than that given by σ_{ϕ_H} . We do this by introducing an error σ_{ϕ_A} , calculated as an error *bound* for the cross-sectional area due to the uncertainty in ice thickness. We compute the magnitude of this error as the area of a band of 10-m thickness (the value of σ_H) all along the glacier cross section. The cross-sectional area incremented (decremented) by the area of this band will give us the upper (lower) bound for the cross-sectional area. If Equation (4.7) is used with σ_{ϕ_H} , we will denote the total error in flux as $\sigma_{\phi(H-based)}$; if, instead, it is calculated using σ_{ϕ_A} , then the total error in flux will be represented as $\sigma_{\phi(A-based)}$. Obviously, $\sigma_{\phi(A-based)} > \sigma_{\phi(H-based)}$, so $\sigma_{\phi(A-based)}$ provides a more conservative approach to the estimate of the discharge error. The use of $\sigma_{\phi(H-based)}$ is only recommended in cases where the radar and SAR acquisitions are simultaneous or very close in time; otherwise, the use of $\sigma_{\phi(A-based)}$ is recommended.

Case 2: only centreline profiles of ice thickness are available

When only radar profiles along the glacier centreline are available, we are forced to make an assumption on the cross-sectional area, and its associated uncertainty will become the dominant error source in the discharge calculation. We will estimate the error in cross-sectional area by comparing, for the 23 glaciers for which a radar cross-profile is available

(see Table 4.2), the areas calculated for the known cross section with those estimated using each of the three U-shape approaches described in Section 3.2. These approaches are based on an interpolation through the points of measured ice thickness and the intersection of the cross section with the glacier margins. Since the flight lines do not follow exactly the glacier centreline, we are interested in determining the error in cross-sectional area as a function of the distance d between the glacier centreline and the radar flight line. With this aim, and to be able to compute some statistics on the errors, the areas and lengths of the cross-sections of all 23 glaciers are normalized to unity, and each glacier's half-width is divided into 30 bins of equal length. Then, for each of the U-shape approaches and for each of the glaciers, a cross-profile is calculated passing through each measured radar data point plus the two points at the glacier margins (Fig. 4.2), and the areas of the resulting U-shape profiles are calculated. Their differences with the corresponding known cross-sectional areas are taken as errors in cross-sectional area. Then, for each of the 30 bins, we compute the mean and the standard deviation of the errors in cross-sectional area calculated for all the radar data points in the bin and for all 23 glaciers. The calculated mean represents the bias of the estimated cross-sectional area, and the calculated standard error will be taken as an error estimate of the cross-sectional area for the distance d under consideration (distance from the central point of the bin to the glacier centreline). Thus, the described procedure provides a bias and an error estimate for the cross-sectional area as a function of the distance d between the radar flight line and the glacier centreline, as shown in Figure 4.3 (we assume that the errors are equal at corresponding distances at each side of the centreline).

4.4 Results and Discussion

4.4.1 Computed ice discharge and estimated errors for glaciers with radar cross-sectional profiles

Table 4.2 shows the discharge results obtained, for the winters of 2016 and 2017, for all glaciers with available cross-sectional radar profiles, together with their corresponding error estimates and the detail of the various error components. The discharge values are given in Mt a^{-1} , and thus correspond to the extrapolation of winter estimates (typically, end of January to mid-March) to annual values. The annual values should not differ much from those given here, as the end-of-winter glacier velocities in the Canadian High Arctic are very close to their annual averages (e.g. 13.6% lower for Ellesmere and Axel Heiberg glaciers (van Wychen et al., 2016)), and the seasonal variability is typically not large (e.g. within 10-20% for Devon glaciers (van Wychen et al., 2012)). Moreover, the discharge values given

Table 4.2 Ice discharge using observed radar cross-sectional profiles

Glacier	Latitude	Longitude	Estimated Error Contributions (Mt a ⁻¹)								Discharge (Mt a ⁻¹)	
			σ_{ϕ_f}	σ_{ϕ_p}	σ_{ϕ_H}	σ_{ϕ_A}	σ_{ϕ_v}	$\sigma_{\phi(H-based)}$	$\sigma_{\phi(A-based)}$	$\epsilon_{\phi \frac{\partial h}{\partial t}}$	2016	2017
Prince of Wales Icefield												
North 1	78.94	-78.05	0.08	0.03	0.05	0.4	0.3	0.3	0.5	-1%	-	14
North 2	78.85	-78.24	0.07	0.02	0.05	0.4	0.2	0.2	0.4	-3%	18	12
Stygge	78.77	-78.24	0.04	0.02	0.1	0.7	2	2	2.1	-6%	8	8
Leffert	78.69	-74.92	0.05	0.02	0.1	0.9	0.2	0.2	0.9	-8%	17	12
Ekblaw	78.51	-76.71	1.4	0.5	0.8	2.6	2.8	3.3	4.1	+1%	-	112
Tanquary	78.46	-76.08	0.09	0.03	0.07	0.4	0.3	0.3	0.5	-2%	15	12
Cadogan	78.23	-76.94	0.13	0.05	0.17	1.7	0.7	0.7	1.9	-2%	49	32
Trinity	77.97	-78.57	3	1	3	40	6.5	7.8	41	-6%	1073	967
Wykeham	77.89	-78.61	1.6	0.6	1.6	20	6.8	7.2	21	-4%	493	419
South Margin	77.71	-77.88	0.1	0.04	0.2	4.7	11	11	12	-6%	67	53
Devon Ice Cap												
Sverdrup	75.72	-83.18	0.08	0.03	0.1	0.6	0.4	0.4	0.7	-1%	12	7
Eastern	75.79	-82.00	0.2	0.08	0.3	1.4	0.3	0.5	1.5	+1%	27	21
Belcher	75.67	-81.39	1.1	0.4	1.2	7.3	4	4.3	8.4	-6%	176	174
Fitzroy	75.45	-80.46	0.4	0.2	0.5	3.5	1.1	1.3	3.6	+2%	84	75
East	75.07	-80.41	0.1	0.03	0.2	1.3	0.3	0.4	1.3	+8%	17	9
South East 1-2	74.98	-80.44	0.2	0.07	0.2	1.4	1.2	1.2	1.8	-6%	34	64
South Crocker	74.85	-83.20	0.2	0.08	0.3	1.7	0.4	0.5	1.8	0%	31	47
North Crocker	74.91	-83.62	0.03	0.01	0.05	0.5	0.3	0.3	0.6	-3%	6	8
Northern Ellesmere Icefield												
Marine	82.24	-81.74	0.01	0	0.02	0.1	0.2	0.2	0.2	-4%	2	2
Marine North	82.41	-82.56	0.02	0.01	0.03	0.2	0.2	0.2	0.3	-3%	6	6
Manson Icefield												
Mittie West Arm	76.90	-79.53	0.02	0.01	0.02	0.1	0.4	0.4	0.4	0%	4	6
Mittie East Arm	76.87	-79.12	0.02	0.01	0.02	0.2	0.5	0.5	0.5	0%	6	5
Sydkap Ice Cap												
Sydkap	76.62	-85.11	0.2	0.08	0.3	1.4	2.4	2.4	2.8	+5%	29	23

by e.g. van Wychen et al. (2016) and Millan et al. (2017) also correspond to approximately the same period in the year, which makes the comparison of results simpler.

We note that the estimated error in discharge due to the errors in the angle between the velocity vector and the vector normal to the cross section, σ_{ϕ_α} , was negligible as compared to the rest of error components and therefore it has been excluded from the table. Discharge values for North 1 and Ekblaw Glaciers in 2016 are not given because of unavailability of proper SAR data.

As described in the Methods section, we give two different estimates of the total error in ice discharge: $\sigma_{\phi(H-based)}$, to be used in the case of temporarily coincident radar measurements and SAR acquisitions, and a more conservative error estimate $\sigma_{\phi(A-based)}$, to be used otherwise. For some glaciers this distinction does not imply a significant difference in the error estimates. However, in other cases $\sigma_{\phi(A-based)}$ is up to three to five times larger than $\sigma_{\phi(H-based)}$, as happens for the largest glaciers (Wykeham and Trinity). It is also three times larger for some medium-size glaciers (Fitzroy, South Crocker, Eastern) and three to five times larger for some small-size glaciers (East, Leffert). On average, $\sigma_{\phi(H-based)}$ is 5% of the discharge and $\sigma_{\phi(A-based)}$ is 8%, though individual values can be as high as 21% (South Margin) to 25% (Stygge) for $\sigma_{\phi(H-based)}$, and 23% (South Margin) to 26% (Stygge) for $\sigma_{\phi(A-based)}$.

The percentage values given under column $\epsilon_{\phi \frac{\partial h}{\partial t}}$ represent the change in discharge implied by the consideration of glacier thinning (negative values) or thickening (positive values) between the radar and SAR data acquisitions. We note that the discharge values given in the

last two columns have *not* been corrected for this systematic error. There is a mixture of thinning and thickening glaciers, with total ice-thickness changes between the radar and SAR data acquisitions ranging from the 8% thinning of Leffert Glacier and the 8% thickening of East Glacier. The average of the absolute values of the ice-thickness changes is of $\sim 3\%$. There is a clear predominance of thinning in the Prince of Wales Icefield and the Northern Ellesmere Icefield, no changes in the Manson Icefield, thickening in the Sydkap Ice Cap and no clear trend in the Devon Ice Cap.

Regarding the individual error components, as most error sources have been assigned constant values (σ_ρ , σ_H , σ_v , σ_f), their individual contributions to the total error depend in a systematic way on the characteristics of each glacier (geometry, velocity field). In general, the contributions to the error in flux of the errors in ρ and f are small. The main contributors to the total error are the uncertainties in thickness and in velocity, as has been acknowledged by other authors (Burgess et al., 2005; McNabb et al., 2015). Their relative contributions to the total error in ice discharge will depend on the glacier under consideration, so that the thinner and slower the glaciers (e.g. $<100 \text{ Mt a}^{-1}$ and $<100 \text{ m a}^{-1}$), the larger the shares of thickness and velocity to the total error, and conversely. The above is true when considering the total error as given by $\sigma_{\phi(H\text{-based})}$. If, instead, we use $\sigma_{\phi(A\text{-based})}$ as total error estimate then the interpretation on how the glacier characteristics affect the error estimate changes. In this case, the velocity field is the largest contributor to the total error for most of the small glaciers (discharge $<100 \text{ Mt a}^{-1}$) with low velocities ($<100 \text{ m a}^{-1}$) (Stygge, Marine, Mittie West Arm, Mittie East Arm) and some medium-size glaciers with low velocities (South Margin, Sydkap). This situation gradually changes with the glacier dimensions, so the cross-sectional area uncertainty becomes the largest contributor to the total error for the largest glaciers (Trinity, Wykeham).

4.4.2 Computed ice discharge and estimated errors for glaciers with radar centreline profiles

Error estimates for the various U-shaped approaches

The bias and the standard deviation of the cross-sectional area calculated using each of the U-shape approaches discussed in Section 3.2 are shown, as a function of the distance d between the radar flight line and the glacier centreline, in Figure 4.3. The bias and standard error are calculated as described under Case 2 of Section 3.4.

For coincident radar profile and glacier centreline, the centred parabolic approach of van Wychen et al. (2014) shows zero offset and around 20% standard error in area, and the offset steadily decreases with increasing distance radar profile-glacier centreline, without

Fig. 4.3 Normalized cross-sectional area errors for the three different U-shape approaches, as a function of the normalized distance between the radar flight line and the glacier centreline. The vertical bars represent the standard deviation, and the distance from the centre of each bar to the zero line represents the corresponding bias. The continuous lines indicate the variation of the bias with the normalized distance. The blue bars/lines correspond to the off-centred parabolic approach, the green ones to the off-centred quartic approach, and the red ones to the centred parabolic approach of van Wychen et al. (2014).

significant increase in standard error. By contrast, the off-centred parabolic approach shows nearly zero offset for distances between the radar profile and the glacier centreline up to $\sim 65\%$ of the glacier halfwidth, but then it increases exponentially. The percent standard error in this case increases steadily from values close to 20% for distances radar-centreline close to zero to around 50% for the mentioned distance of $\sim 65\%$ of the glacier halfwidth. Finally, the off-centred quartic approach starts with a similar standard deviation but with a rather large positive offset for distances radar-centreline close to zero. This offset decreases steadily until a distance of about 75% of the glacier halfwidth (during its decrease, it becomes negative at distances of about 40% of the glacier halfwidth). Then it starts to increase exponentially, although more slowly than the off-centred parabolic approach. The standard error of the quartic approach increases with distance more slowly than that of the off-centred parabolic approach. The exponential growth of the bias for the off-centred approaches, for large distances between the radar profile and the glacier centreline (i.e. when the radar flight line approaches the glacier margins) is a consequence of the fact that the denominator in equations (4.4) and (4.6) approaches zero in such cases, while no such impact is visible in Equation (4.2) (note, however, that the latter equation is based on the ideal assumption that the radar flight line and the glacier centreline are coincident).

The negative offset of the centred parabolic approach is consistent with the results by van Wychen et al. (2014), who noted that their approach underestimated the cross-sectional area by $\sim 12\%$. The assumption that the longitudinal radar profile coincides with the glacier centreline is fairly good for most of the airborne radar data of NASA's Operation IceBridge. We analysed the distances (in absolute value) between the radar flight lines and designated glacier centrelines for all of the glaciers in this region, obtaining a mean distance of $\sim 22\%$ of the glacier halfwidth. For this average distance from the centreline, we see from Figure 4.3 that our calculated underestimation of the area for the centred parabolic approach is $\sim 10\%$, which compares well with the results by van Wychen et al. (2014). To correct for this negative bias, van Wychen et al. (2014) increased their discharge values by 12%. However, the fact that the bias steadily increases in absolute value with increasing distances radar profile-centreline suggest that, rather than applying a single correction factor for all glaciers as done by van Wychen et al. (2014), it would be advisable to apply a correction factor for the cross-sectional area in a case by case basis, depending on the distance profile-centreline for each particular glacier.

Based on the fact that the distance between radar profile and glacier centreline is usually not large in this region, we decided to choose the off-centred parabolic approach for our discharge calculations, as this approach shows nearly zero offset and admissible standard deviations under such conditions. In fact, the distances from the profile to the glacier

Table 4.3 Comparison of ice discharges calculated using observed and estimated cross-sectional profiles.

Glacier	Latitude	Longitude	Discharge using observed thickness data (Mt a ⁻¹)		Discharge using estimated thickness data (Mt a ⁻¹)		$\Delta\phi(\%)$
			2017		2017		
			ϕ_{obs}	$\sigma_{\phi_{obs}}$	ϕ_{estim}	$\sigma_{\phi_{estim}}$	
North 1	78.94	-78.05	14	0.5	15	5.6	-7
North 2	78.85	-78.24	12	0.4	20	5	-67
Ekblaw	78.51	-76.71	112	4	107	20	4
Trinity	77.97	-78.57	967	41	928	160	4
Wykeham	77.89	-78.61	419	21	380	106	9
Cadogan	78.23	-76.94	32	1.9	30	11	6
Sverdrup	75.72	-83.18	7	0.7	8	1.1	-14
Sydkap	76.62	-85.11	23	2.8	22	4.2	4
South East 1	74.98	-80.44	62	1.8	63	12	-2
South Crocker	74.85	-83.20	47	1.7	42	8.5	11

centreline for which the off-centred parabolic approach starts to show an undesired behaviour (large offset and standard deviation) are rarely reached in the study area (Leuschen et al., 2010).

Comparison of observed and estimated cross-sectional fluxes

To check the performance of the off-centred parabolic approach, we present in Table 4.3 the discharge results for some glaciers for which a cross-sectional radar profile is available, obtained using: 1) the radar-measured cross-sectional area, and 2) its approximation by the off-centred parabolic approach. We remark that, for each glacier, both calculations are made at the location of the available cross-sectional radar profile (i.e. without considering the criteria for optimal location of the approximated cross section described in the next section), hence the slight differences with the discharge values shown later in Table 4.4 for the coincident glaciers.

The relative percent differences between observed and estimated discharges, $\Delta\phi$, show that in some cases there is an underestimation (positive $\Delta\phi$) and in other cases an overestimation (negative $\Delta\phi$) of the discharge. If we exclude North 2 Glacier, for which there is a large underestimation, the average of the absolute values of the deviations is $\sim 7\%$. The discharge error estimates are on average 6 times larger when using the cross-sectional area estimated using the U-shape approach. In spite of these differences, both sets of discharge estimates have comparable values, in the sense that, for each glacier, the calculated discharges by both methods are within error bounds (even for North 2 Glacier it falls within the 95% confidence interval or $\pm 1.96\sigma_{\phi_{estim}}$). Whereas for glaciers with low to moderate discharge (below 100 Mt a^{-1}) the parabolic approach shows no bias, for larger glaciers (e.g. Trinity and Wykeham) the ice discharge is underestimated using both parabolic and quartic approaches (the latter one not shown in the table). The underestimation in the parabolic case ($\sim 10\text{-}30\%$) is larger than that of the quartic case ($\sim 5\text{-}15\%$). This stresses the need for observing the cross-sectional profiles of large glaciers in order to better constrain their estimated ice fluxes.

Criteria for choice of flux-gate position

When radar ice-thickness profiles are only available along (or close to) the glacier centreline, an important decision to take is where to locate the cross section whose area is to be approximated using the U-shape approach. It might seem obvious that it should be located as close as possible to the glacier calving front, as we aim to estimate the ice discharge to the ocean. However, in this section we will see that other factors should be taken into consideration when choosing the flux-gate location. We will illustrate these criteria using the sample case of Vanier Glacier shown in Figure 4.4. In this figure we show the radar flight line (panel a) and the location of 4 glacier cross-sections (A through D), and in panel b we show the variations along the radar profile of the main parameters whose values will be of help to determine a suitable location for the flux gate. These include the ice discharge, the cross-sectional area, the distance between the radar profile and the glacier centreline (expressed in the figure as position of the radar profile with respect to the glacier margins), the west-east and south-north components of the velocity field and the along-flow changes of these velocity components (i.e. the components of the velocity gradient along the radar profile); all of them are shown in panel b.

We see in Figure 4.4b that the ice discharge steadily decreases as we approach the glacier terminus. This is mostly due to the fact that this calculation of ice discharge does not include a correction for surface mass balance between the chosen flux gate and the glacier calving front (Gardner et al., 2011), and illustrates the importance of such a correction for flux gates distant from the glacier terminus. It is also obvious that the chosen flux gate should be closer to the terminus than the confluence of any significant tributary glacier (e.g. cross section D would not be suitable, while C would be admissible, because a tributary glacier coming from the north-east joins the main trunk slightly upglacier from cross section C). As shown in the previous section (illustrated by Figure 4.3), the choice of a cross section for which the radar profile and the glacier centreline are close to each other is critical to prevent both an undesirable large standard deviation and large bias in the estimated cross-sectional area. In particular, in the case of the off-centred parabolic approach, using any cross section for which the distance from the profile to the glacier centreline is larger than 65% for the glacier halfwidth would render the calculated ice discharges useless.

It is equally important, when choosing the flux-gate location, to avoid zones with large variations of either the cross-sectional area or the glacier velocity. This is because these rapid spatial variations often correspond to marked short-scale heterogeneities of the glacier bedrock (e.g. bedrock bumps) which imply locally anomalous cross-sectional areas, as well as noise in the velocity field signal, both of which should be avoided whenever possible to

Fig. 4.4 Spatial variations along the radar longitudinal profile of Vanier Glacier of the main parameters to be considered for a suitable choice of the flux-gate location. In panel b, the relative position of the radar profile is indicated (lefts axis) as the percentage over the total glacier cross-sectional length; therefore a 50% value indicates that the radar profile is located exactly at the glacier centreline.

minimize the error in the discharge estimate. These undesired effects can also be minimized by averaging the ice discharges calculated for several closely-spaced flux gates:

$$\begin{aligned}\phi_{avg} &= \frac{\sum_{j=1}^n \phi_{A-based_j}}{n}, \\ \sigma_{\phi_{avg}} &= \frac{\sigma_{\phi(A-based)}}{\sqrt{n}},\end{aligned}\tag{4.9}$$

where n is the number of considered flux gates. We see in Equation (4.9) that the error decreases with the inverse of the square root of n . We suggest a limit of 11 flux gates (ca. 150 m in our case study) for the averaging (the calculated discharge value would be assigned to the central flux gate), given the underlying assumption of Equation (4.9) that we are averaging distinct measurements of the same quantity so the fluxes should not differ significantly. The area of each cross section would be calculated using the off-centred parabolic approach based on a different radar ice-thickness point for each cross section.

Summarizing the above, our choice for the flux-gate location should be guided by:

1. Flux-gate close to the glacier terminus and downglacier with respect to any significant tributary glacier.
2. radar profile close to the glacier centreline.
3. Spatial stability of the cross-sectional area calculated with the U-shaped approach.
4. Spatial stability of the ice surface velocity field.

In Table 4.4 we show the ice discharges calculated using the off-centred parabolic approach for glaciers with only one radar longitudinal profile, close to the glacier centreline (no cross-sectional profiles available). The location of the approximated cross section has been selected following the criteria discussed above. The U-shape approximation has also been applied to three glaciers (marked with an asterisk) for which there is an available cross-sectional radar profile. The reason is that, for these glaciers, the available radar cross section is located far from the calving front (especially for North 1 and North 2). This implies a large difference between the discharge estimates calculated using the observed cross section (in Table 4.2 and in column ϕ_{obs} of Table 4.3) and the estimated discharge shown in Table 4.4, which is based on an estimated cross section situated much closer to the glacier terminus. Consequently, the latter does not need a correction for surface-mass balance between the flux-gate position and the calving front, while the former would require it to provide a fair estimate of the ice discharge to the ocean. Therefore, the estimate of ice discharge given in Table 4.4 for these particular glaciers is the recommended one. As shown in Table 4.4,

Table 4.4 Ice discharge values calculated using estimated cross-sectional areas by means of the off-centred parabolic approach.

Glacier	Latitude	Longitude	Estimated Errors (Mt a ⁻¹)		Discharge (Mt a ⁻¹)	
			$\sigma_{\phi(A-based)}$	$\sigma_{\phi_{avg}}$	2016	2017
Prince of Wales Icefield						
*North 1	78.94	-78.05	5.6	1.8	10	3
*North 2	78.85	-78.24	1.5	0.5	4	5
*Ekblaw	78.51	-76.71	20	6.3	212	129
South	77.33	-79.59	2.8	0.9	32	28
Palisade	77.39	-80.99	2.5	0.8	7	6
Northern Ellesmere Icefield						
Disraeli North	82.84	-70.79	1.3	0.4	6	5
Disraeli	82.67	-72.50	13	4.1	-	10
M'Clintock	82.43	-76.15	4	1.3	1	3
Milne	82.44	-80.22	6.9	2.2	30	45
Vanier	82.14	-80.75	2	0.6	4	8
DeVries	82.01	-79.60	0.4	0.1	1	1
Yelverton	81.84	-79.43	10	3.1	74	79
Otto	81.30	-84.70	0.4	0.1	1	0
Agassiz Ice Cap						
Tuborg	80.89	-76.14	8.7	2.7	30	33
Antoinette	80.81	-76.30	4	1.3	2	20
d'Iberville	80.56	-77.92	1.3	0.4	5	5
Cañon	79.68	-79.64	14	4.4	68	80
Sawyer Bay	79.36	-78.05	0.6	0.2	3	2
Parrish	79.57	-77.18	0.3	0.1	2	1
Eugenie	79.82	-74.93	4	1.3	2	21
Unnamed 4	80.07	-72.39	2.6	0.8	10	10
Müller and Steacie Ice Caps						
Iceberg	79.43	-92.37	1.4	0.4	9	8
Good Friday Bay	78.55	-91.76	20	6.3	9	11

* The glaciers marked with an asterisk have an available radar cross section.

errors are reduced when averaging the ice discharge using several closely-spaced flux gates as described in Equation (4.9) (compare $\sigma_{\phi(A-based)}$ with $\sigma_{\phi avg}$).

4.4.3 Comparison of calculated ice discharge with other studies

Several recent studies have dealt with the estimation of ice discharge from tidewater glaciers of the Canadian High Arctic (Millan et al., 2017; van Wychen et al., 2016, 2017) and a further study has analysed the glacier surface velocity changes in the circum-Arctic region (Strozzi et al., 2017). We compare their results with those presented in this paper with the support of Table 4.5. We note that the measurements from all studies correspond to approximately the same period of the year, the end of the winter, so differences should not be attributed to seasonality. We see that, in most cases, their results are comparable and consistent with those presented here.

When comparing our results with those of van Wychen et al. (2016), we acknowledge an increase of ice discharge from the main glaciers of the Prince of Wales Icefield (Trinity and Wykeham) in 2016, in line with the trend indicated by van Wychen et al. (2016) for 2013-2015. This increase is followed by a decrease in 2017, which is consistent with the decrease in surface velocities of the main Canadian Arctic tidewater glaciers pointed out by Strozzi et al. (2017). In fact, the only large tidewater glacier keeping a similar rate of ice discharge for both 2016 and 2017 is Belcher Glacier in Devon Ice Cap. However, the mentioned increase in discharge for Trinity and Wykeham from 2015 to 2016 becomes a substantial decrease (also for Belcher) if we compare our results with those of Millan et al.

(2017). If the comparison of our results were limited to those obtained by the latter authors for 2015, we could think that the differences in discharge estimates for Trinity, Wykeham and Belcher, 27, 28 and 20% lower (respectively) for 2016 in our estimate, could be partly attributed to interannual variations in ice discharge, even if these show typical values of $\sim 10\text{--}15\%$ for Canadian High Arctic glaciers, obtained from comparison of differential Global Positioning System data recorded during the summer and winter months (van Wychen et al., 2014). However, we note that there is an inconsistency between the estimates by van Wychen et al. (2016, 2017) and by Millan et al. (2017) for 2015, with discharge estimates by the latter authors substantially higher for the largest glaciers, by 27, 29 and 29% for Trinity, Wykeham and Belcher, respectively. In all three cases, our own estimates for 2016 are much closer to those given by van Wychen et al. (2016, 2017).

Comparing our results with those of van Wychen et al. (2016, 2017) and Millan et al. (2017), we note a substantial increase in discharge of Ekblaw Glacier in 2016, followed by a decrease in 2017, though maintaining a level higher than that of 2015. Though our estimate for 2016 can seem very high, it is comparable to the values given for 2011–2012 by van Wychen et al. (2016) (not shown in table). This is no surprise, considering the pulsating behaviour of Ekblaw Glacier noted by van Wychen et al. (2016), and are within the range of random variations during the period 1999–2015 analysed by these authors. We also found differences in ice discharge for Tanquary Glacier, whose discharges for 2016 and 2017 are lower than that observed by van Wychen et al. (2016) in 2015, but similar to that given by Millan et al. (2017) also for 2015. We found that the surface velocity field of this glacier presents a singular behaviour, with its southern part showing no signs of movement. The ice flowing in this area comes from four tributary glaciers located to the south of Tanquary, which are nearly stagnant. The difference in discharge estimates could be due to a distinct weight given to the differing velocities of the northern and southern parts of the cross section, although an alternative explanation is that the glacier could be entering into a quiescent phase. For Sydkap and Cadogan Glaciers we also find a decrease in discharge in 2016, and a further decrease in 2017, from the figures reported by van Wychen et al. (2016) for 2015. The pattern is similar for Sydkap when compared with the results by Millan et al. (2017), though different for Cadogan due to the 33% lower estimate by Millan et al. (2017) as compared with van Wychen et al. (2016). In the Devon ice cap, Fitzroy Glacier shows no sign of change when compared with van Wychen et al. (2017) results, but Millan et al. (2017) gave a discharge estimate 33% lower than that of van Wychen et al. (2017). We note that there is a gap in ice-thickness data for a certain area of Fitzroy Glacier, which could be the reason for at least part of the observed differences in ice discharge estimates.

Table 4.5 Comparison of ice discharge values between studies.

Glacier	van Wychen et al. (2016)	Millan et al. (2017)	This study	
			Using observed profiles	Using estimated profiles
Trinity	1.02 (2015)	1.30 (2015)	1.07 (2016) 0.97 (2017)	-
Wykeham	0.41 (2015)	0.53 (2015)	0.49 (2016) 0.42 (2017)	-
Belcher	0.17 (2015)	0.22 (2015)	0.18 (2016) 0.17 (2017)	-
Ekblaw	0.08 (2015)	0.06 (2015)	0.11 (2017)	0.21 (2016)
Tanquary	0.05 (2015)	0.01 (2015)	0.02 (2016) 0.01 (2017)	-
Sydkap	0.04 (2015)	0.05 (2015)	0.03 (2016) 0.02 (2017)	-
Cadogan	0.06 (2015)	0.04 (2015)	0.05 (2016) 0.03 (2017)	-
Yelverton	0.08 (2015)	0.11 (2015)	-	0.07 (2016) 0.08 (2017)
Fitzroy	0.08 (2015)	0.05 (2015)	0.08 (2016) 0.08 (2017)	-
Tuborg	0.03 (2015)	0.03 (2015)	-	0.03 (2016) 0.03 (2017)
Cañon	0.07 (2015)	0.07 (2015)	-	0.07 (2016) 0.08 (2017)
Milne	0.06 (2015)	0.04 (2015)	-	0.03 (2016) 0.05 (2017)
Good Friday Bay	0.05 (2015)	0.08 (2015)	-	0.01 (2016) 0.01 (2017)

All values are given in Gt a^{-1} .

The main glaciers of Aggasiz Ice Cap, Tuborg and Cañon, show similar discharge values for the present study and those for 2015 by van Wychen et al. (2016) and Millan et al. (2017). For Yelverton Glacier, the main active glacier of Northern Ellesmere Icefield (Otto Glacier is in its quiescence phase), our estimate is closer to that of van Wychen et al. (2016), while for Milne Glacier our estimate is closer to that of Millan et al. (2017). Both Glaciers, Yelverton and Milne, show a similar pattern of decrease in discharge from 2015 to 2016, followed by an increase in 2017. Good Friday Bay Glacier, located in the Steacie Ice Cap, presents a different discharge for all three studies. We believe that this disparity should not be attributed to seasonality or interannual variability, nor to its surging behaviour (Copland et al., 2003a; van Wychen et al., 2016). Rather, we attribute this difference to the location of the flux gate in our study. The terminus of this glacier advanced ~ 2 km during 2000-2015 (van Wychen et al., 2016). This advance implies an increase of the glacier area in its lowest part. Therefore, the effect of the surface mass balance losses on the increased glacier area could easily account for the differences in estimated ice discharge. Indeed, when calculating the ice discharge using radar observations close to the terminus we obtain an ice flux of $\sim 10 \text{ Mt a}^{-1}$, while when calculating it for flux-gates 10 km upglacier from the terminus (avoiding the area where a Nunatak is present), our estimated discharge value becomes similar to that of van Wychen et al. (2016).

4.5 Conclusions and outlook

We have analysed the contributions of the various error components involved in the estimation of ice discharge through predefined flux gates, distinguishing two cases: 1) ice-thickness data is available for glacier cross-sections close to the glacier terminus, and 2) ice-thickness data is only available along the glacier centreline. In the latter case, we have analysed the performance of three different U-shaped cross-sectional approaches, and given hints for the

choice of a suitable location of the flux-gate. The following conclusions can be drawn from our study:

Regarding the relative contribution from the various error components:

1. The velocity field is the dominant source of error for small and medium-size glaciers (discharge $< 100 \text{ Mt a}^{-1}$ with low velocities ($< 100 \text{ m a}^{-1}$)).
2. For large glaciers (discharge $> 100 \text{ Mt a}^{-1}$) with high velocities ($> 100 \text{ m a}^{-1}$) the error in cross-sectional area becomes the main contributor to the total error. This stresses the need of measuring radar cross-sectional profiles for the largest glaciers.
3. The bias (systematic error) implied by glacier thinning/thickening between the radar and SAR acquisitions is variable according to subregions, oscillating between 8% and -8% of the discharge value over the period of 2-5 years between acquisitions, with an average of the absolute values of $\sim 3\%$. Temporally coincident radar and SAR acquisitions are recommended to reduce the effect of the bias, especially for the largest glaciers (Trinity, Wykeham, Belcher, Ekblaw), which contribute to most of the ice discharge in the region.

Concerning the performance of the U-shaped cross-sectional approaches:

1. If the radar flight line is not too far from the glacier centreline, the off-centred parabolic approach shows the lowest bias and acceptable standard deviation, so it is the recommended approach.
2. The centred parabolic approach shows nearly constant standard deviation, but is more strongly biased than the off-centred approach for common distances flight line-centreline.
3. The off-centred quartic approach shows large variable bias and its standard deviation, though nearly constant, is large.

Finally, regarding the comparison of the ice discharge results presented here for the winters of 2016 and 2017 with the results for 2015 presented by van Wychen et al. (2016, 2017) and Millan et al. (2017), in general we see comparable results with only small differences. When there is a difference between the results by these authors (usually a larger estimate by Millan et al. (2017)), our data in general agree better with those by van Wychen et al. (2016, 2017), so we will comment on the comparison with the latter to gain some understanding on the interannual changes. There is an increase of ice discharge from the main

glaciers (Trinity and Wykeham) of the Prince of Wales Icefield from 2015 to 2016, by 5% and 20%, respectively, but this is followed by a decrease in 2017, by 10% and 15% respectively, consistent with the reduction in surface velocities of the main Canadian Arctic tidewater glaciers pointed out by Strozzi et al. (2017). Among the largest glaciers, only Belcher Glacier, in the Devon Ice Cap, maintains similar discharges during the period 2015-2017. Two small glaciers show significant decreases in ice discharge. Tanquary, part of the Prince of Wales Icefield, changes by 70% from 2015 to 2016, and a further 20% from 2016 to 2017. Good Friday Bay, part of the Steacie Ice Cap, decreases by 80% from 2015 to 2016, remaining stable in 2017.

In our view, most of the work remaining to be done does not correspond to the analysis of error estimates of ice discharge through given flux gates and at a given time, as done in this paper, but to the approximation of the frontal ablation of tidewater glaciers by the ice discharge calculated at flux gates close to the calving front. Aside from seasonality and interannual variability considerations, we believe that two critical aspects deserve further investigation for Canadian High Arctic glaciers: 1) the surface mass balance between the flux gate location and the calving front, and 2) the front position changes. If the surface mass balance effects are ignored, the overestimate of frontal ablation is expected to be large for Canadian High Arctic glaciers, because of the strongly negative recent surface mass balance of the Canadian Arctic, with values between -1.5 and -2.0 m w.e. a^{-1} at the lowermost part of the tidewater glaciers during 2003-2009 (Gardner et al., 2011). Our own preliminary estimates for the studied glaciers suggest typical overestimates by $\sim 30\%$ of the calculated ice discharge, reaching up to 50% for individual glaciers. Regarding the effect of the terminus advance/retreat, it is difficult to quantify for Canadian High Arctic glaciers, due to the pulsating behaviour of many of them. As noted by van Wychen et al. (2016), pulse-type and surge-type glaciers have some common characteristics, such as periods of speedup and slowdown, and terminus advance coincident with acceleration, but their key difference is that all of the velocity variability of the pulse-type glaciers appears to be restricted to their lowermost terminal region, which is grounded below sea level. Very little interannual variability is observed upglacier from this area. This poses difficulties in the approximation of frontal ablation by ice discharge through flux-gates, as these estimates are heavily dependent on flux-gate location for the pulsating glaciers. This stresses the importance of monitoring the terminus advance/retreat for the glaciers in this region, as has been done during the last decades (Burgess and Sharp, 2004; Burgess et al., 2005; van Wychen et al., 2016; Williamson et al., 2008).

Chapter 5

Intra- and inter-annual variability in dynamic discharge from the Academy of Sciences Ice Cap, Severnaya Zemlya, Russian Arctic, and its role in modulating mass balance¹

5.1 Introduction

Calving is an important mechanism of mass loss for marine-terminating Arctic glaciers, including those of the Russian Arctic (Dowdeswell et al., 2008; Huss and Hock, 2015). The Russian Arctic, which comprises the archipelagos of Novaya Zemlya, Severnaya Zemlya and Franz Josef Land (Dowdeswell et al., 2010) (Fig. 5.1a), had a total glacierized area of 51,592 km² around 2000-2010, of which 65% corresponded to tidewater glaciers (Pfeffer et al., 2014). Its total ice volume is estimated to range between 15,000 and 18,000 km³ (Kotlyakov et al., 2010). In spite of recent climate warming over the Arctic region (Hartmann et al., 2013), the glacier ice mass losses from the Russian Arctic have been moderate, of $\sim 11 \pm 4$ Gt a⁻¹ over 2003-2009 (Gardner et al., 2013). This is far behind other Arctic regions such as the Canadian Arctic or peripheral Greenland, or Alaska, even if considered per unit area (Gardner et al., 2013). However, the mass losses from the Russian Arctic to the end of the 21st century have been projected to increase substantially, with an expected contribution between 9.5 ± 4.6 and 18.1 ± 5.5 mm in sea-level equivalent (SLE) over 2010-2100, depending on

¹(Sánchez-Gómez et al., 2018)

emission pathway (Huss and Hock, 2015). These projected losses are similar to those of the other regions, in spite of the substantially lower current glacierized area of the Russian Arctic. This projected sea-level rise contribution by glacier wastage from the Russian Arctic is equivalent to an average rate of ice mass loss between 38 and 73 Gt a^{-1} over the period 2010-2100.

There are, however, substantial differences among the various estimates of current mass losses, not only among those obtained using different techniques, such as ICESat altimetry versus GRACE gravimetry, but also among those obtained using a common technique. For instance, Moholdt et al. (2012b) find mass changes of $-9.8 \pm 1.9 \text{ Gt a}^{-1}$ using ICESat data and of $-7.1 \pm 5.5 \text{ Gt a}^{-1}$ using GRACE data in the Russian High Arctic, both for the same period October 2003-October 2009. Regarding additional GRACE estimates, Moholdt et al. (2012b) find mass changes of $-4.6 \pm 5.4 \text{ Gt a}^{-1}$ over a longer period April 2003-March 2011, while Jacob et al. (2012) obtain changes of $-5 \pm 3 \text{ Gt a}^{-1}$ over January 2003-December 2010, and Matsuo and Heki (2013) report values of $-15.4 \pm 11.9 \text{ Gt a}^{-1}$ over February 2004-January 2008 and $-6.9 \pm 7.4 \text{ Gt a}^{-1}$ over February 2004-January 2012. Although some of the differences can be attributed to the non-overlapping periods, part of the discrepancies likely stem from the uncertainty in the glacial isostatic adjustment (GIA) correction and is known to be poorly constrained in this region (Svendsen et al., 2004). We recall that the GIA has limited effects on the estimates derived from ICESat, while a strong effect on those derived from GRACE (Hanna et al., 2013).

Within the Russian Arctic, most of the current ice mass losses ($\sim 80\%$) correspond to Novaya Zemlya, with Severnaya Zemlya and Franz Josef Land contributing the remaining $\sim 20\%$ (Moholdt et al., 2012b). For this reason, several recent studies have focused on determining which is the main driver of the large ice mass losses from Novaya Zemlya. While Moholdt et al. (2012b) suggest that climate (in particular, increased melt due to higher temperatures) is a more important factor in driving mass loss at Novaya Zemlya than glacier dynamics, Carr et al. (2014) claim that calving flux may be a more important driver of mass loss than suggested by Moholdt et al. (2012b). Melkonian et al. (2016), while acknowledging the important role of climate in driving the recent mass losses of land-terminating glaciers, indicate that calving flux is an important contributor to mass losses at the marine-terminating glaciers along the Barents Sea coast, suggesting that retreat at these glaciers reflects dynamic processes. Some recent events have raised the interest on Severnaya Zemlya, in spite of its lower current rate of mass loss. One of them was the collapse of the Matusевич Ice Shelf, October Revolution Island (Fig. 5.1b) in 2012, with subsequent accelerated thinning of the glaciers feeding the ice shelf (Willis et al., 2015). Another relevant event is the ‘slow’ surge of the Vavilov Ice Cap, also on October Revolution Island, initiated around 2015, which has

involved both a large increase in ice surface velocity and a remarkable advance in frontal extent (Glazovsky et al., 2015; Strozzi et al., 2017).

Satellite remotely-sensed glacier surface velocity estimates for the Russian Arctic have been very scarce until recent years, and even scarcer those focused on the use of such velocity fields to calculate ice discharge to the ocean. Among the earlier studies are those of Sharov and Tyukavina (2009) and Dowdeswell et al. (2002), both based on interferometry using ERS-1/ERS-2 Synthetic Aperture Radar (SAR) images of September-December 1995, and Moholdt et al. (2012a), based on image matching between repeat-pass Landsat optical imagery of 2000-2002. The velocity analyses of all three studies focused on the Academy of Sciences Ice Cap, Komsomolets Island (Fig. 5.1b and c), which at 5,575 km² is the largest ice cap of Severnaya Zemlya and is the study site of the present paper. While the former study used the SAR data to determine the overall velocity field and help to define the ice divides, the two latter studies used the velocity fields to calculate the calving flux of the main marine-terminating basins of the ice cap. Recent availability of larger amounts of remotely-sensed SAR data from platforms such as TerraSARX, PALSAR-1 and Sentinel-1 has allowed further studies such as that by Melkonian et al. (2016), which used TerraSARX imagery of 2014 (together with WorldView, Landsat and ASTER optical imagery) to investigate the velocity fields and constrain the calving flux at the two fastest-retreating glaciers of Novaya Zemlya. Another example is the investigation by Strozzi et al. (2017) of the velocity fields of all glacierized Arctic archipelagos, including the Russian Arctic, where they analysed the variability of glacier dynamics using SAR data from various sources: JERS-1 data of 1998, ALOS-1 PALSAR-1 data of 2008 to 2010, and Sentinel-1 data of 2016-2017 (plus some additional sources in the case of the Vavilov Ice Cap). Their analysis of the Russian Arctic focused on Novaya Zemlya, where the availability of SAR imagery was largest, while for Franz Josef Land and Severnaya Zemlya the analysis was more limited, with the exception of the Vavilov Ice Cap on Severnaya Zemlya.

Focusing on the Academy of Sciences Ice Cap the various estimates available for different periods from 1988 to 2009 (Dowdeswell et al., 2002; Moholdt et al., 2012a) differ substantially, indicating large interannual and decadal variations, but also suggesting some possible under- or overestimations due to the limitations of the available data. For instance, Moholdt et al. (2012a) suggest that the estimates by Dowdeswell et al. (2002) for 1995 could be underestimated because of the unresolved glacier movement perpendicular to the SAR look angle. On the other hand, there are no available studies analysing the intra-annual or seasonal variations in calving flux.

All of the above discussion motivates the present study, aimed at the calculation of updated estimates of calving flux for the Academy of Sciences Ice Cap and of its intra-annual

and seasonal variations. This is undertaken through the analysis of 54 pairs of weekly Sentinel-1 SAR Terrain Observation by Progressive Scans (TOPS) acquisition mode images acquired from November 2016 to November 2017 with a 12-day period between the images in each pair. Here we use the terms *calving flux* and *ice discharge* interchangeably because over our studied period the front position changes have been negligible (see more details in the subsection Dynamic ice discharge). The study of the calving flux regime is of particular relevance for the Academy of Sciences Ice Cap because a large fraction ($\sim 42\%$) of its margin is marine and $\sim 50\%$ of its bed is below sea level (Dowdeswell et al., 2002). We aim to understand, and analyse the consistency, of the long-term variations in calving flux observed in recent decades by other authors. We are also interested in determining whether the intra-annual and seasonal velocity variations in calving flux could partly explain the comparatively low calving flux reported by Dowdeswell et al. (2002), which is the one based on the shortest temporal baseline. We also present an analysis of the main flow regimes present in the Academy of Sciences Ice Cap, based on the relationship between the ratio of surface velocity to ice thickness and the driving stress. We also report the initiation of ice stream flow in the southern part of the ice cap. Finally, we calculate the geodetic mass balance of the ice cap, infer its climatic mass balance and estimate the contribution to total ablation of surface ablation and frontal ablation.

5.2 Study site

The Academy of Sciences Ice Cap is located on Komsomolets Island, Severnaya Zemlya, and is one of the largest Arctic ice caps, with an area of $\sim 5575 \text{ km}^2$ and an estimated volume of $\sim 2184 \text{ km}^3$. Its dome reaches 787 m a.s.l. (ArcticDEM) and the ice cap has a maximum ice thickness of $\sim 819 \text{ m}$ (Dowdeswell et al., 2002). The latter authors, together with Moholdt et al. (2012a), are responsible for the two main observationally-based studies of the the dynamics and mass balance of the Academy of Sciences Ice Cap. Dowdeswell et al. (2002) undertook radar flights to determine the ice thickness of the ice cap and SAR interferometry from ERS tandem phase scenes of 1995 to infer the ice velocities. Combining these data, they estimated the calving flux from the ice cap. In their analysis of the form and flow of the ice cap, which included the analysis of the driving stresses, they indicated that there is no evidence of past surge activity within the residence time of the ice (Dowdeswell and Williams, 1997). In particular, they noted that there are no evidences of any deformation of either large-scale ice structures or medial moraines. Assuming zero climatic mass balance (based on observations at the neighbouring Vavilov Ice Cap by Barkov et al. (1992)) and a total accumulation rate based on an ice cap drilled at the ice cap summit (Fritzsche et al., 2005,

2002; Opel et al., 2009) plus an assumption on the change of accumulation with altitude (Bryazgin and Yunak, 1988), they also estimated the surface ablation of the ice cap and thus the shares of total ablation by surface ablation and calving, showing that iceberg calving contributes to $\sim 40\%$ of the total mass losses. Moholdt et al. (2012a), in turn, used ICESat altimetry, together with older DEMs and velocities from Landsat imagery, to calculate the geodetic mass balance and the calving flux for various periods during the last three decades, showing that variable ice-stream dynamics dominates the mass balance of the ice cap. Some studies of ice-flow modelling and physical parameter inversion have also become available for the Academy of Sciences Ice Cap (Konovalov, 2012; Konovalov and Nagornov, 2017).

The climate of Severnaya Zemlya is classified as a polar desert with both low temperatures and precipitations (Moholdt et al., 2012b). The atmospheric circulation is dominated by high-pressure areas over Siberia and the Arctic Ocean and low-pressures over Barents Sea and Kara Sea (Alexandrov et al., 2000; Bolshiyarov and Makeyev, 1995). There exists a gradient in precipitation between the south and the north with the Kara Sea as a probable moisture source (Bolshiyarov and Makeyev, 1995; Zhao et al., 2014). This precipitation gradient manifests in the south-north increase of the ELA ranging from ~ 600 m for Vavilov Ice Cap, ~ 400 m for the Academy of Sciences Ice Cap and ~ 200 m for Schmidt Island (Bassford et al., 2006a; Dowdeswell et al., 2002).

Two Russian permanent weather stations, Golomyanny and Fedorova, located in the nearby Sredniy Island and on the northern tip of Taymyr Peninsula, provide meteorological records from the 1930s (Alexandrov et al., 2000; Dowdeswell et al., 1997). These datasets show a mean annual surface air temperature of -14.7 °C and -15 °C respectively, with Fedorova registering a maximum of 1.5 °C in July. Mean annual precipitation is also similar for both weather stations, ~ 0.19 m for Golomyanny and ~ 0.2 m for Fedorova (Alexandrov et al., 2000; Dowdeswell et al., 1997), whereas the ice caps receive a higher amount of precipitation of ~ 0.4 m (Bolshiyarov and Makeyev, 1995).

However, Zhao et al. (2014) have shown that NCEP-NCAR reanalysis summer temperatures over Severnaya Zemlya have weak correlations with Golomyanny Island station-measured summer mean temperatures. They note that Golomyanny Island station is located within the Severnaya Zemlya archipelago 130 km away from the Ice Cap to the southwest into the Kara Sea, with only 7 m a.s.l., and is strongly influenced by the cold ocean environment due to melting sea ice in summer. On the other hand, Opel et al. (2009) found no correspondence between the amount of melt layers in the ice core drilled at the Academy of Sciences Ice Cap summit and the Golomyanny station summer surface air temperatures. Furthermore, Zhao et al. (2014) found that the total melt days at Severnaya Zemlya were strongly correlated with NCEP-NCAR reanalysis summer temperatures. All of these evi-

Fig. 5.1 a) Location of Severnaya Zemlya within the Russian Arctic. b) Main ice masses of Severnaya Zemlya (Wessel and Smith, 1996). The glacier outlines are from the Randolph Glacier Inventory (RGI) version 5.0 (Pfeffer et al., 2014). c) Surface topography of the Academy of Sciences Ice Cap in Severnaya Zemlya. The outlines defining the various drainage basins of the ice cap are from the RGI. We have named basins A, B, C and D following Dowdeswell et al. (2002) and Moholdt et al. (2012a). We have denoted the basin between basins B and C as Basin BC. We further introduced the names West Basin, South Basin and Southeast Basin. The ellipsoid used in this study is WGS84. The polar stereographic projection is used for map a). The rest of the maps in this study are projected in UTM 47 North.

dence question the use of Golomyanny Station temperature data, which we will avoid in our analysis.

An automatic weather station installed on top of the Academy of Sciences Ice Cap between May 1999 and May 2000 provided temperature information of the air and the snow (Kuhn, 2000). The mean annual air temperature was -15.7°C , while the average temperature of uppermost 10 metres of snow/firn was -10.2°C , because of the latent heat released by the refreezing of percolating surface meltwater. During the summer months of July and August temperatures are commonly above freezing point causing snowmelt and decrease of the snow height (Kuhn, 2000).

The temperature record of the Academy of Sciences Ice Cap during the last 275 years, inferred from $\delta^{18}\text{O}$ ice core concentrations, shows a minimum in ~ 1790 followed by an increasing overall trend up to present but with an absolute double maximum in the first half of the 20th century (Fritzsche et al., 2005; Opel et al., 2013). This increasing temperature trend helps explaining the role of the Kara Sea as a moisture source in the area (Opel et al., 2009; Zhao et al., 2014) as does the increase in sea salt content in glacier low altitudes, specially during warm summers (Opel et al., 2013). The increase in moisture in the region was also triggered by the decreasing trend of sea ice cover in the Arctic beginning in the 1980s (Stroeve et al., 2011). The overall picture of temperature change in the last decades is specially critical for the Arctic region with a tipping point at the beginning of the 1980s (Hansen et al., 2010).

5.3 Data

5.3.1 Synthetic aperture radar velocity data

We obtained surface velocities on the Academy of Sciences Ice Cap from Sentinel-1B SAR Terrain Observation by Progressive Scans (TOPS) Interferometric Wide (IW) Level-1 Single Look Complex (SLC) images (Zan and Guarnieri, 2006). This type of product provides a geo-referenced image (using accurate altitude and orbital information from the satellite) in slant-range geometry, processed in zero-Doppler. The image is normally composed of three sub-swaths with each sub-swath comprising normally 9 consecutive bursts, which overlap in azimuth. Burst synchronisation is needed for interferometry and for accurate offset tracking (Holzner and Bamler, 2002). The resolution of Sentinel-1 SAR TOPS IW mode is of 5 and 20 m in the range and azimuth directions respectively. We used the vertical transmit and vertical receive (VV) channel, which preserves best the amplitude features and also has a

higher signal-to-noise ratio as compared with the horizontal transmit and vertical receive (VH) channel, making the former better suited for retrieval of glacier ice motion.

The TOPS acquisition mode improves the performance of already-existing SAR imaging algorithms such as ScanSAR mode (Zan and Guarnieri, 2006). During a TOPS mode acquisition, the SAR antenna is steered in the azimuth direction from aft to fore with a constant rate. This mode of observation has several advantages; in particular, the measured targets are observed with the whole azimuth antenna pattern, which also reduces the scalloping effect (Zan and Guarnieri, 2006). The main disadvantage is that it imposes a more restrictive approach for the co-registration procedure and for the interferometric processing. The peculiarities of the co-registration with the spectral diversity method have been addressed by Grandin (2015b). The SAR images used in this study were acquired from November 2016 to November 2017. See more detail in Supplementary Materials Table C1.

5.3.2 Surface elevation data

We used ICESat elevation data from version 34 of the GLAH06 altimetry product (Zwally et al., 2014). This dataset was acquired by the Geoscience Laser Altimeter System (GLAS) onboard ICESat (Zwally et al., 2002). The period of activity of this sensor spanned 6 years, from October 2003 to October 2009. The satellite was operated in campaign mode, being able to retrieve data from the same ground tracks for 17 periods of ~ 33 days each. ICESat altimetry is very accurate (~ 15 cm) when gentle sloping topography is considered (Zwally et al., 2002). The majority of the tracks used in this study date from Spring 2004. See more detail in Supplementary Materials Table C2.

We also used the ArcticDEM derived from high-resolution sub-meter satellite imagery from the WorldView satellite constellation. The Surface Extraction with TIN-based Search-space Minimization (SETSM) algorithm allows a fully automated retrieval of surface heights. The resulting DEMs are finally adjusted in position using ICESat-derived altimetry as a reference (Noh and Howat, 2015; Noh et al., 2016). The horizontal resolution of the strip DEM product is $2\text{ m} \times 2\text{ m}$ while the tile DEM product is of 5 m. The vertical accuracy of these datasets depends on the use of ground control points as a final step for DEM vertical position refinement. Thus, when no ground control is available, the DEM accuracy relies on the sensor's rational polynomial coefficients accuracy. Typically the uncertainty figures range from 4 m when no ground control is used and sub-metre accuracy with ground control (Noh and Howat, 2015; Noh et al., 2016). See more detail on the set of ArcticDEM used in this study in Supplementary Materials Table C3.

5.3.3 Ice thickness from airborne radar, and other data

We used ice thickness derived from airborne radar measurements, using a 100 MHz radar, made during the 1997 campaign on Severnaya Zemlya (Dowdeswell et al., 2002). The dataset was assessed to have a standard deviation of ~ 4.9 m at the crossing points of the flight tracks.

We used several additional sources of ancillary data to support our investigation. These include Sea Surface Temperatures derived from the Terra Thermal Infrared sensor at 9 km resolution (OBPG, 2015a,b), sea-ice concentration (Fetterer F. and Knowles K. and Meier W. and Savoie M. and Windnagel A. K., 2017) and NCEP/NCAR Reanalysis 1 air surface temperatures (Kalnay et al., 1996).

5.4 Methodology

5.4.1 Surface velocities

SAR data processing

We used GAMMA software (Wegmüller et al., 2016) (GAMMA Remote Sensing AG, Gümlingen, Switzerland) for processing the SAR Sentinel-1 acquisitions (Schellenberger et al., 2016; Strozzi et al., 2002). We exclusively applied the intensity offset tracking algorithm.

Sentinel-1 TOPS mode images need to be co-registered before any offset tracking or interferometric algorithms are applied on them. This fine co-registration procedure requires the use of a fine DEM within the same area of the acquisition image extent (Wegmüller et al., 2016). The co-registration begins with obtaining the DEM in SAR coordinates followed by the application of the matching algorithm and the spectral diversity method (which uses the interferometric phase) applied over the overlapping areas of the bursts. The co-registration requirements are quite stringent and a co-registration quality of 1/1000 of a pixel in the azimuth direction is required for the phase discontinuity between consecutive bursts edges to be less than three degrees (Scheiber et al., 2015).

After a full co-registration is achieved, a deramping of the SLC images for correcting the azimuth phase ramp is required to apply oversampling in the offset tracking procedures (Wegmüller et al., 2016). Once the above-mentioned steps are done, the offset tracking technique is the same as for normal stripmap mode scenes (Strozzi et al., 2002; Werner et al., 2005). Surface displacements can be obtained in ground coordinates (e.g. slant range and azimuth directions), which are finally geocoded using a lookup table derived from the use of the DEM and the image parameter file.

We used a matching window of 320 by 64 pixels (1200 by 1280 m) in range and azimuth directions respectively, with an oversampling factor of two for improving the tracking results. The sampling steps were of 40 by 8 pixels and the resolution of the final velocity map was 200 by 160 m in range and azimuth directions. The geocoding was completed using the ArcticDEM. We used a bicubic spline interpolation to generate the geocoded grid. This grid was transformed into a Tiff file using a functionality within GAMMA software. Finally, the determined velocities were manually checked for blunders and mis-matches, which were then removed from the dataset.

Surface velocity error estimates

We estimated the error in surface velocity by analysing the performance of the algorithm on stable ground in Komsomolets Island (e.g. on ice-free areas to the north of the Academy of Sciences Ice Cap) under the hypothesis that the error of the offset tracking technique on these areas should be close to zero. We calculated the root-mean-square error (RMSE) of the normal intensity offset tracking results (i.e. including azimuth offsets). The results of our analysis cast an RMSE value of $\sim 0.013 \text{ m day}^{-1}$ ($\sim 4.5 \text{ m a}^{-1}$) for the range offsets and $\sim 0.021 \text{ m day}^{-1}$ (7.5 m a^{-1}) for the azimuth offsets. The combined error in the magnitude of the ice surface velocity is of $\sim 0.024 \text{ m day}^{-1}$ ($\sim 8.75 \text{ m a}^{-1}$). These figures are similar to those presented in other studies (Short and Gray, 2004; van Wychen et al., 2017).

These error estimates vary between individual SAR image pairs, mentioned figures are sufficiently representative of the actual error of the whole set of images. The main factors that have an impact on the error budget in these image pairs are: 1) the matching procedure (which is a function of the co-registration between images, template size, and the quality of the image features), and 2) the geocoding error (given the high-quality ephemerides data of Sentinel-1, this error is reduced to the quality of the DEM used for the topographic correction (Nagler et al., 2015)). The relatively short time interval between the images of each pair of Sentinel-1 images (12 days) could have an adverse effect on the estimates of error in velocity for slow-moving glaciers. However, all considered marine-terminating glaciers have high surface speeds so this situation does not apply.

5.4.2 Surface elevation change rates and associated mass changes

We have used two different datasets to determine the surface elevation changes of the Academy of Sciences Ice Cap during recent decades: ICESat altimetry and ArcticDEM. Following a methodology similar to that used by Melkonian et al. (2016), who estimated $\partial h / \partial t$ by applying a weighted linear regression to stacked DEMs on a pixel-by-pixel basis.

Extrapolation was used in the areas where no ice was found in the latest DEM acquisitions. The decadal-scale surface elevation changes were estimated by differencing ICESat altimetry and WorldView DEMs, which provided the $\partial h/\partial t$ averages over the period 2004–2016. The elevation change rates were split into 25-m height bins using a hypsometry calculated from the WorldView DEMs. The mean $\partial h/\partial t$ were calculated basin-wise and hypsometry-wise. Finally, we converted the volume change rates

$$\Delta V \approx \sum_{i=1}^n A_i d\bar{h}_i \quad (5.1)$$

to mass loss rates multiplying by an ice density of 900 kg m^{-3} . This assumes Sorge's law (Bader, 1954), i.e. that there is no changing firm thickness or density through time and that all volume changes are of glacier ice. We calculated the elevation change rates at the basin level to be able to distinguish the particular signal of each basin (see Fig. 5.4). We considered two error sources, the error derived from the differencing of the two datasets and the extrapolation error. The error of the differencing was calculated as the square root of the sum of the squares of the measurement errors of ICESat altimetry (Moholdt et al., 2010; Zwally et al., 2002) and the WorldView DEMs (Noh and Howat, 2015). The extrapolation error was estimated from the difference, within the same height bins, of the calculated point-wise elevation change rates from ICESat altimetry and the mean elevation change rate obtained from WorldView DEMs.

The short-term changes in surface elevation were calculated by differencing pairs of Worldview-derived DEMs. The pairs of ArcticDEM strips that were compared correspond to 2012–2013 and 2016. The DEMs were vertically coregistered to the ICESat dataset. We used DEMs from similar periods of the year (May–July), with a 3–4 year difference. This temporal gap allowed minimizing the influence of seasonal variability in the retrieved elevation change (e.g. an episode of snow precipitation could conceal the signal). The errors in elevation change rate were estimated by comparing two WorldView DEMs on ice-free areas. This analysis provided a RMSE value of 0.91 m for the height differences. Finally, the error for the basin-wide mass change rates was calculated using error propagation.

5.4.3 Flow regime mapping

We calculated a set of four ice-flow regimes following Burgess et al. (2005) and van Wychen et al. (2017). The flow regimes are classified according to the relationship between the ratio of the surface velocity to the ice thickness (v/h) and the driving stress (τ_d) along flowlines across the ice cap. When basal motion is by internal deformation only, the quotient v/h represents the mean shear strain rate in a vertical column, while, when basal motion is

important, v/h is a measure of the effective viscosity of the glacier ice. The driving stress is derived from

$$\tau_d = \rho g h \sin \alpha, \quad (5.2)$$

where ρ is ice density (900 kg m^{-3}), g gravity (9.81 m s^{-2}), h the ice thickness and α the glacier surface slope. The local surface slope was averaged using a moving window with a size of 10 times the glacier ice thickness in order to reduce the effect of local variations due to longitudinal stress gradients. Burgess et al. (2005) based their classification on the increasing role of basal sliding in glacier motion, compared with internal deformation, as the considered region gets closer to the terminus of an outlet tidewater glacier. The four main flow regimes are defined as follows (Burgess et al., 2005):

1. Flow regime 1: values of $v/h < 0.075 \text{ a}^{-1}$ and with a high positive correlation with τ_d . This flow regime describes ice motion solely by deformation with few to no contribution of basal sliding. These regions are characterised by convex upward surface profiles.
2. Flow regime 2: values of v/h between 0.075 and 0.28 a^{-1} . The glacier regions with this type of flow present a decrease in effective viscosity and flow resistance, with a larger contribution of basal motion to the surface velocity. This flow regime appears at the head of the outlet glaciers where convergent flow and the presence of flow stripes on the ice surface is common. The surface profile transitions from convex-up to concave-up.
3. Flow regime 3: values of $v/h > 0.28 \text{ a}^{-1}$ and $\tau_d > 0.075 \text{ MPa}$. The areas with this flow regime show a reduction in viscosity and probably also a larger weight of basal motion to the total surface velocity. These regions are characterised by the presence of well-developed flow stripes at the glacier surface.
4. Flow regime 4: values of $v/h > 0.28 \text{ a}^{-1}$ and $\tau_d < 0.075 \text{ MPa}$. The flow in these areas is characterised by low basal friction, so basal motion is the major component of the surface velocity. Sediment deformation may contribute to basal motion.

We note that the surface-slope averaging involved in the calculation of driving stress could have an influence on the separation between flow regimes 3 and 4 near the glacier termini, because the averaging window size decreases as the glacier margins are approached.

5.4.4 Dynamic ice discharge

In this paper we use the term *calving flux* as defined in Cogley et al. (2011), denoting the *ice discharge* calculated through a flux gate close to the calving front minus the mass flux involved in front position changes. In our case study, spanning the period November 2016–November 2017, the front position changes have been negligible (in some basins, small advances in localized zones are compensated by small retreats in other zones), so calving flux and ice discharge are equivalent. On the other hand, *frontal ablation*, that is, ice mass losses by calving, subaerial frontal melting and sublimation and subaqueous frontal melting at the nearly-vertical calving fronts (Cogley et al., 2011), can also be considered in our study as nearly equivalent to calving flux or ice discharge. The reason is that subaerial frontal melting and sublimation are usually very small, and submarine melting is assumed here to be small because no substantial retreat has been observed along the more stagnant ice fronts around the ice cap (Moholdt et al., 2012a). Moreover, if ice discharge is calculated from remote sensing velocities assuming a grounded terminus and vertical front geometry, and the flux gate used is close to the calving front, then the calculated ice discharge includes both the calving losses and the submarine melting. Consequently, in our case study the mass balance of marine-terminating basins will be simply the result of climatic mass balance at the glacier surface (disregarding basal mass changes) minus calving mass losses (estimated through ice discharge). For the land-terminating basins, mass balance will be simply the climatic mass balance at the glacier surface (again, disregarding basal balance).

Calculation of ice discharge

There is some evidence from both radar data collected in 1997 and earlier investigations by Russian scientists that parts of the ice cap margin at the seaward end of the ice streams may be floating (Dowdeswell et al., 2002). However, we calculated ice discharge at flux gates located within $\sim 1.5 - 3$ km from the calving front, where ice is grounded. Ice discharge is thus calculated as mass flux per unit time across a given surface S , approximated using area bins as

$$\phi = \int_S \rho \mathbf{v} \cdot d\mathbf{S} \approx \sum_i \rho L_i H_i f v_i \cos \gamma_i, \quad (5.3)$$

where ρ is ice density, L_i and H_i are respectively the width and thickness of an area bin, f is the ratio of surface to depth-averaged velocity, v_i is the magnitude of surface velocity and γ_i is the angle between the surface velocity vector and the direction normal to the local flux-gate for the bin under consideration. In the literature it is assumed that $f \in [0.8, 1]$ (Cuffey and Paterson, 2010). To give a good approximation for this parameter an analysis of the driving

stresses present in a glacier and its derived flow regimes would be advisable (Burgess et al., 2005; van Wyken et al., 2017). However, simple observation of the glacier surface features could give a hint on the flow regime and therefore help to better constrain f . Normally, tidewater glacier velocity at the terminus is dominated by basal sliding which makes f close to unity. Following Vijay and Braun (2017), we took $f = 0.93 \pm 0.05$, assuming that all tidewater glaciers in the Ice Cap have a large component of basal motion. For ice density, we took $\rho = 900 \pm 17 \text{ kg m}^{-3}$.

The flux gates that we used cover the whole frontal area of all marine-terminating glacier basins. Once the flux-gate is defined for a glacier, it is divided in small bins of the same length. The ice thickness of each bin was calculated interpolating the ice thickness field provided by Dowdeswell et al. (2002). This preliminary ice-thickness was then adjusted using the surface elevation change estimated from the comparison of the ICESat and ArcticDEM elevation datasets. The velocity vector orientations were calculated with respect to the vector normal to each flux-gate bin.

Estimate of error in ice discharge

We estimate the error in ice discharge following Sánchez-Gómez and Navarro (2018). Applying error propagation to Equation (5.3) we get

$$\sigma_\phi = \sqrt{\sigma_{\phi_\rho}^2 + \sigma_{\phi_f}^2 + \sigma_{\phi_H}^2 + \sigma_{\phi_v}^2 + \sigma_{\phi_\gamma}^2}, \quad (5.4)$$

where each of the terms is of the form (taking σ_{ϕ_H} as an example)

$$\sigma_{\phi_H} = \sqrt{\sum_i (\sigma_H \rho L_i f v_i \cos \gamma_i)^2}. \quad (5.5)$$

In Equation (5.4) we have omitted the term $\sigma_{\phi_L}^2$ because the bin width L_i is assumed to be error-free. As error estimates for the ice thickness and surface velocity we took $\sigma_H = 10 \text{ m}$ (Dowdeswell et al., 2002) and $\sigma_v = 0.024 \text{ m d}^{-1}$ (Strozzi et al., 2002). For the errors in density and in ratio of surface to depth-averaged velocity we took 0.17 kg m^{-3} and 0.05 respectively, as discussed earlier in this section.

5.4.5 Climatic mass balance

Neglecting basal balance, the mass balance rate $\dot{M}$ for a given basin is calculated as

$$\dot{M} = \dot{B} + \dot{D} = \int_S \dot{b} dS + \int_P \dot{d} dp, \quad (5.6)$$

where $\dot{B}$ is the climatic mass balance rate (surface mass balance plus internal balance) and $\dot{D}$ is the calving flux (ice discharge minus mass flux associated to front position changes), calculated as a surface integral over the area S of the basin and a line integral along the perimeter P of its marine-terminating margin, respectively (Cogley et al., 2011). We note that the calving flux term $\dot{D}$ is always negative, as it represents a rate of mass loss. If we know the calving flux and the mass balance rate derived from the surface elevation changes, then we can use Equation (5.6) to estimate the climatic mass balance. In principle, Equation (5.6) also assumes that submarine melting at the glacier front is negligible. This is probably a fair assumption in our case. As Moholdt et al. (2012a) have pointed out, no substantial retreat rates have been observed along the more stagnant ice fronts around the ice cap, indicating that submarine melting is probably small. But, even if it were not negligible, when ice discharge is calculated across vertical flux gates defined at grounded ice close to the calving front, as we do, then submarine melting is in fact included in the ice discharge term rather than neglected. Using this approach we calculated the climatic mass balances for each basin and thus the partition of mass balance into climatic mass balance and frontal ablation (here equal to calving flux).

5.5 Results

5.5.1 Glacier surface velocities and their intra-annual variability

We present in Figure 5.2 the glacier surface velocities inferred from intensity offset tracking of Sentinel-1B IW SAR images. We can easily recognise 5 large-scale marine-terminating drainage basins with ice-stream-like zones of high velocity. In contrast with Dowdeswell et al. (2002) and Moholdt et al. (2012a), we see a further ice stream located on the south of the ice cap (Basin BC). We can also identify two smaller fast-flowing zones on both sides of Ice Stream BC. Several other small fast-flowing zones are located on the Southeast Basin.

All major ice streams present a similar pattern of surface velocity field. The surface velocity starts to increase at the stream heads, where ice flow converges from the accumulation areas. The surface velocity increases up to the glacier fronts where it normally reaches a maximum (Ice Stream A is an exception). The mean surface velocities of all flux gates are shown in Table 5.1. We note that the averaged velocities presented in Figure 5.3 do not match with the ones given in Table 5.1 because the latter are calculated over the entire flux-gate length and the former within a smaller window. Maximum surface velocities are ~ 80 , 1200, 1200, 1100 and 750 m a^{-1} for drainage basins A, B, BC, C and D respectively.

Fig. 5.2 Surface velocities for the various drainage basins of the Academy of Sciences Ice Cap.

Table 5.1 Principal characteristics and yearly-averaged (November 2016–November 2017) ice-discharge rates for the marine-terminating drainage basins of the Academy of Sciences Ice Cap. Velocity and thickness are average values over each basin's flux gate.

Drainage basin	Basin area (km ²)	Flux gate length (m)	FG mean thickness (m)	FG mean surface velocity (m a ⁻¹)	Ice discharge (Gt a ⁻¹)
West	1033	62821	174	6	0.06±0.03
A	707	7274	251	19	0.03±0.01
B	413	5788	83	441	0.18±0.03
South	47	14107	121	28	0.04±0.02
BC	276	6820	184	384	0.41±0.05
Southeast	359	37040	164	15	0.08±0.04
C	829	10594	223	344	0.69±0.07
D	475	10820	171	280	0.44±0.05
Total	4139	155264	166	88	1.93±0.12

Figure 5.3 shows the intra-annual variability of surface velocity averaged along the fastest (central) part of the main ice streams. Each graph consists of 54 velocity values that represent 12-day averages (the repeat period between passes of Sentinel-1), so shorter-term variations are smoothed out. For each of the time series we did a least-square fit to a function sum of a straight line plus a sinusoidal signal. The amplitude of the latter (shown in Table 5.2) gives the seasonal variability, while the slope of the straight line gives the linear trend (also shown in Table 5.2), which represents a deceleration in all cases. Note that the seasonal variability of ice discharge given in Table 5.2 corresponds to the amplitude of the sinusoidal variation, i.e. half of the range of seasonal variability. Ice Stream A is not represented because we did not identify either a significant seasonal signal or a linear trend of its velocity. We also calculated the mean of the absolute values of the deviations of the velocities with respect to the yearly-averaged values (shown in Table 5.2), which gives an idea of the intra-annual variability irrespective of a seasonal signal. As absolute values with respect to the mean have been used, this intra-annual variability represents half of the full range of variability.

Ice streams C and D show a marked seasonal velocity signal and a small decreasing trend. Ice Stream BC shows a marked decreasing velocity trend and a small seasonal signal, and Ice Stream B a mixed decreasing trend-seasonal signal. Excluding Basin BC, which has a weak seasonality, the seasonal signal is minimum in late March and April, and maximum around mid-September, but slightly earlier (mid-August) for the fastest ice stream (Ice Stream B). It should be noted that the minimum/maximum observed velocities do not always match exactly to those of the seasonal fits. For Ice Stream B, the *minimum* observed velocity occurs one month later than the minimum of the seasonal fit, and the *maximum* observed velocities for ice streams C and D happen in July-August, $\sim 1 - 2$ months earlier than the maximum of their seasonal fits, which takes place in mid-September. In all cases (ice streams B, C, D) the maximum observed velocities correspond to an event of sustained high velocities lasting for about two months in July-August.

All ice streams in Figure 5.3 also exhibit short-term, non-seasonal intra-annual variations along the year, though these are much weaker for Ice Stream BC. These velocity variations, with typical ranges within $50 - 100 \text{ m a}^{-1}$ but often exceeding 100 m a^{-1} for Ice Stream D, are all significant, since our average error in surface velocity is of 8.75 m a^{-1} . Ice Stream D is the slowest (yearly-averaged velocity of 491 m a^{-1}) but shows the largest velocity variations, above 200 m a^{-1} during the event of sustained high velocities starting in early July (whose peak-to-peak variation represents a $\sim 40\%$ change over the yearly-averaged velocity). By contrast, Ice Stream BC, whose average velocity is rather high (703 m a^{-1}) shows the smallest intra-annual velocity variations (in addition to the weakest seasonal signal), but has the largest deceleration (-118 m a^{-2}). Ice Stream B is the fastest (average velocity of 912

Fig. 5.3 Surface velocity variability for the largest ice streams of the Academy of Sciences Ice Cap: a) Ice Stream B, b) Ice Stream BC, c) Ice Stream C and d) Ice Stream D. The red line represents the seasonal signal together with the linear trend, and the dashed blue line is the monthly-averaged air temperature over the ice cap (average value for the three cells of NCEP/NCAR Reanalysis 1 over Northern Komsomolets Island). All figures have a common vertical scaling. The monthly axes ticks indicate the start of the corresponding month.

Table 5.2 Average velocities (November 2016–November 2017) over the fastest central part of ice streams B, BC, C and D, (whose temporal variations are shown in Fig. 5.3), linear trend in velocity variation along the year, and intra-annual and seasonal variabilities of velocity. The variability in intra-annual velocity represents the mean of the absolute values of the deviations with respect to the yearly-averaged value (the maximum deviation is also given), while seasonality gives the amplitude of the sinusoidal fit function. The percentages shown are calculated with respect to the mean velocities.

Drainage basin	Mean surface velocity (m a^{-1})	Linear trend (m a^{-2})	Seasonality (a^{-1} , %)	(Gt	Intra-annual variability	
					Mean (Gt a^{-1} , %)	Max (Gt a^{-1} , %)
B	912.4	-100	0.01 6%		0.02 11%	0.03 17%
BC	703.0	-118	0.01 3%		0.06 15%	0.12 30%
C	796.2	-74	0.05 7%		0.10 14%	0.15 21%
D	490.7	-56	0.04 10%		0.07 16%	0.14 32%

m a^{-1}) and has medium-to-large velocity variations. Finally, Ice Stream C has rather high average velocity (796 m a^{-1}) and medium-range velocity variations, some lower than those of Ice Stream B.

Also shown in Figure 5.3 is the monthly-averaged air temperature over the Academy of Sciences Ice Cap. Note that the temperature seasonal signal is $\sim 1 - 2$ months ahead of the velocity seasonal signal, and that the maximum temperatures match in time with the maximum observed velocities, corresponding to the sustained high-velocity event.

5.5.2 Ice discharge

The ice discharge for each marine-terminating drainage basin, averaged over the period November 2016–November 2017, is displayed in Table 5.1. The total ice discharge from the ice cap over the period November 2016–November 2017 is $1.93 \pm 0.12 \text{ Gt a}^{-1}$. The ice discharges for the main drainage basins with well-developed ice streams (A through D) vary between 0.03 and 0.69 Gt a^{-1} . The additional basins with less-developed ice streams (Southeast, South and West basins) contribute with individual ice discharges between 0.04 and 0.08 Gt a^{-1} . Drainage Basin A has the lowest discharge ($0.03 \pm 0.01 \text{ Gt a}^{-1}$), in spite of the large ice thickness of its flux gate, due to its low velocity near the terminus (19 m a^{-1}). Drainage Basin C has the highest discharge ($0.69 \pm 0.07 \text{ Gt a}^{-1}$) as a result of the length and thickness of its flux gate, together with a rather high ice velocity. The highest velocity (441 m a^{-1}) corresponds to Drainage Basin B, but, due to the low average thickness and limited length of its flux gate, results in a moderate ice discharge of $0.18 \pm 0.03 \text{ Gt a}^{-1}$. Drainage basins BC and D have similar and rather large ice discharges (0.41 ± 0.05 and $0.44 \pm 0.05 \text{ Gt a}^{-1}$ respectively) through a combination of high velocity and limited length of its flux gate, and lower velocity and longer flux gate, respectively, at similar average ice thickness at their flux gates. We note that Drainage Basin BC, which in our case is the third contributor to ice discharge from the ice cap, was not found in previous studies to be a significant source of ice discharge (Dowdeswell et al., 2002; Moholdt et al., 2012a). Our data thus indicates that fast ice-stream flow was initiated in this basin, likely during the last decade.

Table 5.3 Surface elevation change rates and mass-change rates for the main marine-terminating drainage basins of the Academy of Sciences Ice Cap. Values are calculated from both ICESat-WorldView and WorldView-WorldView DEM differencing, which represent decadal (2004-2016) and recent, shorter-term (2012/13-2016) average values respectively. The rates for some basins during 2012/13-2016 are not given because of insufficient coverage of the WorldView images in 2012/13.

Drainage Basin	Surface elevation change rates		Mass-change rate	
	ICESat-WV (m a^{-1})	WV-WV (m a^{-1})	ICESat-WV (Gt a^{-1})	WV-WV (Gt a^{-1})
	2004-2016	2012/13-2016	2004-2016	2012/13-2016
North	-0.05 ± 0.10	-	-0.05 ± 0.12	-
West	0.06 ± 0.07	-	0.05 ± 0.06	-
A	-0.10 ± 0.10	-0.12 ± 0.11	-0.06 ± 0.07	-0.07 ± 0.07
B	-0.28 ± 0.11	-0.58 ± 0.18	-0.10 ± 0.04	-0.21 ± 0.08
South	-0.20 ± 0.13	-	-0.02 ± 0.01	-
BC	-1.31 ± 0.33	-1.21 ± 0.24	-0.33 ± 0.08	-0.30 ± 0.06
Southeast	-0.14 ± 0.08	-	-0.05 ± 0.03	-
C	-1.00 ± 0.14	-0.95 ± 0.26	-0.75 ± 0.11	-0.71 ± 0.17
D	-1.02 ± 0.13	-0.84 ± 0.21	-0.44 ± 0.06	-0.36 ± 0.10

The intra-annual and seasonal variations in ice velocity, which imply corresponding and proportional variations in ice discharge, are shown in Figure 5.3. The intra-annual variations other than the seasonal oscillate between 11 and 16% of their yearly-averaged values, but maximum deviations reach up to 32%. The seasonal signal amplitude varies between 3 and 10%. In both cases, the values given represent half of the full range of variation. In terms of ice discharge, the maximum intra-annual variations at a basin scale reach 0.15 Gt a^{-1} and correspond to the event of sustained high velocities in July-August shown in Figure 5.3. The extent of the intra-annual variations illustrates the magnitude of the error that would be undergone when extrapolating to the whole year the ice discharge calculated at a particular snapshot in time.

5.5.3 Surface elevation changes and associated mass changes

The results for surface elevation change rates for the main marine-terminating drainage basins are shown in Table 5.3 and figures 5.4 and 5.5. Table 5.3 also includes the associated mass-change rates calculated assuming, in the volume to mass conversion, an ice density of 900 kg m^{-3} .

The surface elevation changes at the decadal scale (2004-2016) and at the recent, short-term scale (2012/2013-2016) show some differences, but are all within error bounds, except for Basin B, which doubles its thinning rate between both periods. All drainage basins, except Basin West, are thinning, more markedly the basins with ice stream flow draining to the southeast and east (basins BC, C and D). Drainage Basin A, which has the slowest ice-stream flow, shows a small average thinning, but with thinning at the upper part and thickening at the lower part. The thinning pattern is similar for all fast-flowing basins. The highest thinning rates take place where flow converges from the accumulation areas at the head of the major ice streams (see e.g. detail for Basin BC in Figure 5.5). The overall surface elevation change rate map displayed in Figure 5.4 presents a general thinning pattern on all

Fig. 5.4 Results from ICESat-WorldView DEM differencing for the Academy of Sciences Ice Cap in terms of surface elevation change rates ($\partial h/\partial t$). The background image of the ice cap is the WorldView DEM for 2016.

marine-terminating basins and a state close to balance on the land-terminating northern and marine-terminating western drainage basins.

Analysing by individual basins, the land-terminating Basin North low average thinning rate, and the mostly marine-terminating Basin West has low average thickening rates, both with thickening in the upper elevations and thinning in the lower parts, though Basin West has also a zone with thickening at low elevations. Basin A, though has a small average thinning, has a clear thickening signal at low elevations, while thinning at the upper ones. Basin B has apparently doubled its average thinning rate between the two periods analysed. But a note of caution is needed here. The ICESat tracks have a limited coverage of Basin B and do not capture the zone with the largest thinning rates in 2012/13-2016 shown in Fig. C1. Consequently, the thinning rate for the earlier period could be underestimated. Regarding the spatial distribution of surface elevation changes, Basin B shows in 2004-2016 widespread thinning at the upper elevations, and some thickening close to the margin (Fig. 5.4). This thickening is also visible during 2012/13-2016 (Fig. C1), when it reaches values in the order of 1 m a^{-1} . However, the thickening is not marked in either period and, moreover, the surface elevation change signal is less credible in these highly crevassed zones. The small Basin South shows moderate thinning with a complex spatial distribution. Basin BC has the largest average thinning rates, with an average surface elevation change rate of $-1.31 \pm 0.33 \text{ m a}^{-1}$ in 2004-2016. The largest thinning rates in 2004-2016 are found in the zone of stream flow (Fig. 5.4), and in 2012/13-2016 they are remarkably large in the zone of onset of stream flow (surface elevation change rates up to $-8 \pm 0.33 \text{ m a}^{-1}$; Fig. 5.5). Basin Southeast

Fig. 5.5 Results from WorldView-WorldView DEM differencing for Drainage Basin BC in terms of surface elevation change rates ($\partial h/\partial t$).

has small average thinning but a clear thickening in the land-terminating zone. Both basins C and D show marked average thinning rates, with surface elevation change rates in the order of -1 m a^{-1} . We can see in Figure 5.4 a wide zone of thinning that spans most of the basin's area, though the thinning is more moderate at the uppermost and in the vicinity of the margins, where it shows locally some thickening. Interestingly, the slight thinning at the upper elevations over the period 2004-2016 turns into a more recent thickening in the period 2012/2013-2016 for both basins (Figs. C2 and C3).

Focusing on the basins with the largest thinning rates (basins BC, C and D), although absolute thinning is largest for Basin BC, when analysed in terms of mass losses those of Basin C are largest ($\sim -0.75 \text{ Gt a}^{-1}$), due to its larger area, followed by basins D ($\sim -0.44 \text{ Gt a}^{-1}$) and BC ($\sim -0.33 \text{ Gt a}^{-1}$).

5.5.4 Climatic mass balance

Using Equation 5.6 we calculated the climatic mass balance (B) from the available total mass balance (M) and frontal ablation (here equal to calving flux, D). The total mass balance was obtained using the geodetic method from the surface elevation changes, and corresponds to the mass change rates given in Table 5.3 but expressed here also in m w.e. a^{-1} . As we aimed at obtaining a climatic mass balance estimate for a period as close as possible to present,

Table 5.4 Partition of mass balance into climatic mass balance and frontal ablation for the drainage basins of the Academy of Sciences Ice Cap. The climatic mass balance has been derived from WorldView-WorldView DEM differencing for 2012/13-2016, except for the basins marked with an asterisk, for which ICESat-WorldView DEM differencing for 2004-2016 has been used. The frontal ablation data correspond to the period November 2016-November 2017.

Drainage Basin	M		B		D	
	Gt a ⁻¹	m w.e. a ⁻¹	Gt a ⁻¹	m w.e. a ⁻¹	Gt a ⁻¹	m w.e. a ⁻¹
North*	-0.05	-0.04	-0.05	-0.04	0	0
West*	0.05	0.05	0.11	0.11	-0.06	-0.06
A	-0.07	-0.11	-0.04	-0.04	-0.03	-0.06
B	-0.21	-0.52	-0.03	-0.07	-0.18	-0.44
South*	-0.02	-0.18	0.02	0.23	-0.04	-0.46
BC	-0.30	-1.09	0.11	0.4	-0.41	-1.49
Southeast*	-0.05	-0.13	0.03	0.08	-0.08	-0.21
C	-0.71	-0.86	-0.02	-0.02	-0.69	-0.83
D	-0.36	-0.76	0.08	0.17	-0.44	-0.93
Ice Cap total	-1.72	-0.31	0.21	0.04	-1.93	-0.35

we took the geodetic mass balance for the period 2012/13-2016. However, as this was not available for certain basins (North, West, South, Southeast) because of lack of coverage by WorldView images, we used for these basins the geodetic mass balance for the period 2004-2016. The difference, however, is negligible. We calculated a total geodetic mass balance of $-1.72 \pm 0.67 \text{ Gt a}^{-1}$ ($-0.31 \pm 0.12 \text{ m w.e. a}^{-1}$). If we had taken instead the 2004-2016 rates for all basins, we would have got $-1.74 \pm 0.67 \text{ Gt a}^{-1}$ ($-0.31 \pm 0.12 \text{ m w.e. a}^{-1}$). The frontal ablation was approximated by the ice discharge given in Table 5.1, again expressed in m w.e. a^{-1} . The results are shown in Table 5.4, which illustrates the separation of the total mass balance into its two main components, climatic mass balance and frontal mass balance. We note that this does *not* represent the partition of the total mass *losses* into surface ablation and frontal ablation, because the climatic mass balance at the glacier surface is the *net* result of surface accumulation and ablation (corrected for internal balance). We see that the climatic balance of the individual basins is in all cases nearly zero (in fact, not significantly different from zero), and the climatic mass balance for the whole ice cap is $0.04 \pm 0.12 \text{ m w.e. a}^{-1}$. This means that, currently, the *net* losses are due to frontal ablation, since accumulation and ablation at the ice cap surface balance to each other.

5.5.5 Ice Cap flow regimes

We provide a map of the driving stress on the Academy of Sciences Ice Cap (Fig. 5.6) that shows noticeable changes (and more detail) with respect to that given by Dowdeswell et al. (2002). The areas of largest driving stress are readily observed in the regions of convergent flow at the heads of the ice streams. The most remarkable examples are ice streams D and BC, whose driving stresses reach values of up to 150 kPa at their heads. While ice streams A and B show a similar driving stress distribution, their maximum values are lower, of about 100 kPa. There are also areas of high driving stress along the Basin Southeast, the Basin North and the land-terminating margin between Basin A and Basin West, with values of 125, 100 and 100 kPa respectively.

Fig. 5.6 Driving stress field of the Academy of Sciences Ice Cap.

The map of flow regimes, derived from the surface velocity, the ice thickness and the driving stress fields, is presented in Figure 5.7. The map shows an overall dominance of flow regime 1. This flow regime dominates in the accumulation areas of the Academy of Sciences Ice Cap, and also the land-terminating Basin North, except when reaching the terminal parts, where flow regime 2 and ultimately flow regime 4 can be seen, indicating an ever decreasing viscosity while approaching to the glacier terminus, limited to a thin band of a few km from the glacier margin.

Flow regime 2 is present in the zones of transition from the accumulation areas to the high-velocity areas. It coincides with convergent flow at the head of the ice streams (Burgess et al., 2005). Also, as described by Burgess et al. (2005), this area is normally a transition from a convex-up profile typical of flow regime 1 to a concave-up surface profile.

Interestingly, flow regime 3 is only present at the head of Ice Stream BC and more subtly at Ice Stream D. For the rest of the major drainage basins, in general a fast transition from flow regime 2 directly to flow regime 4 occurs. This suggests a rapid transition from zones of high viscosity and flow resistance to zones of low friction at the glacier bed (e.g. linked to the presence of deformable marine sediments (Dowdeswell et al., 2002)).

5.6 Discussion

5.6.1 Intra-annual and seasonal ice velocity variations

When analysing the intra-annual and seasonal velocity variations shown in Figure 5.3, we noted that the seasonal signal was lagging around 1 – 2 months with respect to the air

Fig. 5.7 Flow regimes of the Academy of Sciences Ice Cap.

temperature signal. This supports the interpretation that the summer speed-up is triggered by the drainage of surface meltwater to the glacier bed, enhancing lubrication and basal sliding (Zwally, 2002), as has been observed in many Arctic regions (e.g. Dunse et al. (2012) and references therein). As explained by Copland et al. (2003b), glacier acceleration usually lags the onset of summer melt. Meltwater refreezes in the snowpack until the cold-content of the snow and firn decreases, thereby causing delayed runoff. Basal lubrication does not occur until a connection between the supraglacial and the englacial/subglacial drainage systems is established.

The maximum observed velocities correspond to the sustained high-velocity event starting in late June-early July and lasting for almost two months (three months for the fastest ice stream). The sudden drop in velocities following this event, while the air temperatures remain still high during 1 – 2 additional months (and sustained melt is thus expected), suggests that the abrupt slow-down could be due to the transition from a hydraulically inefficient distributed drainage system to an efficient channelized system (Schoof, 2010).

Recall from Figure 5.3 that the maximum of the seasonal fit to the observed velocities occurred around mid-September (mid-August in the case of the fastest ice stream). This is nearly coincident with the period (in this particular year 2017) when the coasts of northern Severnaya Zemlya passed from being nearly sea-ice free to being fully sea-ice-covered (Figure C4). This is of interest because Zhao et al. (2014) have shown, after ruling out the regional temperature influence using partial correlation analysis, that the total melt days on both Novaya Zemlya and Severnaya Zemlya is statistically anti-correlated with regional late summer sea ice extent, linking in this way land ice snowmelt dynamics to regional sea-ice extent variations.

The relationship between higher air temperatures, increased surface melt, and enhanced basal lubrication and sliding can explain both the seasonal velocity variations and the shorter-term intra-annual velocity changes occurring during the summer. However, shorter-term intra-annual velocity variations take place along the whole year. This is a common feature of the observed velocities of many outlet glaciers in Greenland and other polar regions (Dunse et al., 2012; Howat et al., 2010) but its occurrence during the winter is seldom analysed in the literature. Trying to understand the reasons for the observed intra-annual oscillations in surface velocity during the non-melting period, we did a Fourier analysis of the residual signal resulting from subtracting the linear trend and the sinusoidal fit from the velocity time series. However, the resulting Fourier power spectra did not show any significant peaks, so we can discard periodic processes such as e.g. monthly tides (note that, because of the 12-day averaging and 7-day sampling period, we cannot detect any shorter-period tidal component). Other possible periodic signals, such as a hypothetical regular passage of cyclones, cannot be detected because typical cyclone frequencies over northern Severnaya Zemlya are $0.075\text{--}0.1\text{ d}^{-1}$ (10–13 d period) during the winter and $0.125\text{--}0.15\text{ d}^{-1}$ (7–8 d period) during the summer (Zahn et al., 2018). The most thorough analysis of the physical basis of the intra-annual glacier velocity variations, which explains the glacier velocity variations during the winter, is that of Willis (1995). If the Academy of Sciences ice Cap were hard-bedded, then a possible explanation for the observed velocity variations would be as follows. As explained by Willis (1995), during the winter a low subglacial water flux will tend to be routed via a thin film. Subglacial water pressures will be high and sliding might occur by the drowning of small roughness features on the bed. If during the winter water is also trapped at high pressure in subglacial cavities, sliding might be expected to be even higher as a result of the drowning of larger bed-roughness features. However, a soft bed has been suggested for the Academy of Sciences ice Cap (Dowdeswell et al., 2002). As noted by these authors, internal deformation alone would account for a just a few meters per year of movement, so fast flow within the ice streams is likely to be through basal motion. As $\sim 50\%$ of the ice cap bed lies below present sea level and terminates in the adjacent seas, it is likely that much of the bed is made up of deformable marine sediments. If we thus rule out the explanations for hard-bedded glaciers, a theory at hand that explains velocity variations during winter, again following Willis (1995), is that a low subglacial water flux will tend to be routed via Darcian flow through the till. Subglacial water pressures will be high and stable constant bed deformation might occur. However, this can explain stable constant bed deformation, but not oscillatory deformations such as those observed in our records. The same applies to the theories indicating that an inefficient drainage system, such as a layer of subglacial sediments, might have retained a significant fraction of water throughout

the winter months, thereby facilitating continuously high, but diminishing deformation of water-saturated sediments (Bougamont et al., 2011; Fischer and Clarke, 2001; Tulaczyk et al., 2000). An alternative explanation suggested by Willis (1995) is that periods of accelerating unstable bed deformation may occur temporarily during till failure by liquefaction. Finally, we note that oscillations in fast flow could be caused by changes in driving stresses associated e.g. to changing surface slope due to heterogeneous cumulative snow accumulation during the winter. All of this, however, still remains as speculation because of lack of observational evidences, though opens the door to further investigations.

5.6.2 Initiation of stream flow at Basin BC and suspected current slow-down

The most striking difference in ice cap dynamics between our study and those of Dowdeswell et al. (2002) and Moholdt et al. (2012a) has been the initiation of stream flow in Basin BC. This flow accounts for nearly all of the difference in total ice discharge from the ice cap between our own estimate (1.93 Gt a^{-1}) and that of Moholdt et al. (2012a) (1.4 Gt a^{-1}). Ice stream flow was clearly not visible in the interferograms in Figure 9 of Dowdeswell et al. (2002) and in Figure 5a of Sharov and Tyukavina (2009), both based on the SAR acquisitions of 1995. In fact, Dowdeswell et al. (2002) explicitly mention that 'A fifth basin with rough topography, on the south side of the ice cap, shows only limited evidence of fast flow'. The decrease in driving stress between the study by Dowdeswell et al. (2002) (between 75 and 100 kPa; Figure 12a of Dowdeswell et al. (2002)) and our analysis (25-50 kPa; Fig. 5.6), reflects the large drop in surface slope associated to the initiation of stream flow. Additionally, the analysis of the flow regimes shown for Ice Stream BC in Figure 5.7 reveals a peculiar behaviour for this ice stream as compared to others: the zone of flow regime 3 is the largest of all ice streams, and also the size of the zone of combined flow regimes 2 and 3 as compared with that of flow regime 4. This suggests a later start of ice stream flow in this basin.

The ice velocity analysis by Moholdt et al. (2012a) using Landsat 2000-2002 images did not cover this area. But the comparison of the Landsat-7 image of July 2002 with the Sentinel-2 image of March 2016 shown in Figure 5.8 clearly shows the initiation of fast flow between acquisitions. Note the similarity of the flow characteristics of Basin South in both images, in clear contrast with the marked dissimilarity for those of Basin BC, with stream flow evident in the 2016 image.

Regarding the 2003-2009 $\partial h / \partial t$ data of Moholdt et al. (2012a), the ICESat repeat tracks do not cover the area of fastest flow of Ice Stream BC in our 2016-2017 velocity data, but the track on the zone above that of onset of fast flow shows a clear increase in thinning rate

Fig. 5.8 Comparison of Landsat-7 and Sentinel-2 images, acquired on 09/07/2002 and 29/03/2016 respectively. The crevassing of Glacier BC is noticeable on the second image.

(by $0.25 - 0.50 \text{ m a}^{-1}$) from 1988-2006 to 2003-2009 (figures 4b and 4c of Moholdt et al. (2012a)). In our analysis of ICESat-WorldView (2004-2016) surface elevation changes, the ICESat track do not cover the zone of fastest flow, but shows a marked thinning (surface elevation change lower than -0.7 m a^{-1}) just above the zone of onset of fast flow of Ice Stream BC (Fig. 5.4). WorldView-WorldView (2012/13-2016) surface elevation changes, on the other hand, show $\partial h / \partial t$ values up to -8 m a^{-1} in the zone of onset of fast flow, but much lower, even some slight thickening, near the margin, though the signal is very noisy in this area.

All of the above indicates that stream flow in Basin BC was initiated clearly after 2002, and likely during the period 2003-2009, though the fastest flow was probably reached some time later. But there are also some indications of a possible current slow-down of Ice Stream BC. One of them is the recent slight thickening observed near the margin of Ice Stream BC, though, as mentioned, the signal is very noisy in this area so this indication is weak. Second, the clear decreasing trend in velocities observed, during our studied period November 2016-November 2017, in the central part of the ice stream (Fig. 5.3b and Table 5.2). The deceleration during this period has been of -118 m a^{-2} . Although the rest of ice streams have also shown a decreasing trend in velocities, the trend for ice stream BC has doubled those of ice streams C and D. However, the trend for ice stream B has been nearly as large as that of ice stream B. Moreover, the period analysed for short-term velocity variations is limited to one year. Consequently, the slowdown of Ice Stream BC could be part of the regional response to some driver of the velocity field, and perhaps be temporary. Finally, the slowdown of Ice Stream BC does not have to be understood as an indication of cessation of stream flow but probably just of stabilisation of its fast flow in a regime closer to steady state.

5.6.3 Comparison of thinning rates with those of previous studies

Moholdt et al. (2012a) calculated, from DEM differencing for various periods (1956-1997, 1988-2006 and 2003-2009), the surface elevation changes of the Academy of Sciences Ice Cap. They focused their analysis on the two later periods, because of limitations in accuracy and resolution of the 1956 and 1997 DEMs. The DEM for 1956 was in fact built from a combination of Russian topographic maps of 1956 and 1988, and the DEM for 1997 was based on surface elevation profiles from airborne radio-echo sounding in 1997, constructed from pressure-altimeter data. We here analyse, basin by basin, our observed surface elevation change rates for the period 2004-2016 (which have been presented in Section 5.3) in the context of the earlier analysis by Moholdt et al. (2012a), focusing on the main changes observed along the three periods. Recall from Section 5.3 that the changes between our two studied periods (2004-2016 and 2012/13-2016) were negligible, with the exception of Basin B.

Basins North (land-terminating) and West (marine-terminating but with slow flow) are grouped together under 'North' in Moholdt et al. (2012a) study, and have remained fairly stable along the entire sequence of periods analysed, with average slight thickening in 2003-2009 for the grouping, and slight thinning/thickening for basins North/West in 2004-2016, respectively, without significant change of the space distribution of the surface elevation change rate (SECR).

Basin A in 2004-2016 shows thinning at the upper elevations and thickening at the lower elevations, as observed during both periods analysed by Moholdt et al. (2012a), who noted a surge-like elevation change pattern supported by the velocity field of the 1995 InSAR data of Dowdeswell et al. (2002). Moholdt et al. (2012a) also pointed out that dynamic mass flux at the equilibrium line must be larger than the balance flux. This instability was much larger in the period 1988-2006 than during 2003-2009, indicating glacier deceleration. This trend has continued during 2004-2016, with differences in surface elevation change rates between the upper and lower parts larger than 0.8 m a^{-1} (Fig. 5.4), suggesting continued deceleration. For Basin A, as for Basin North, there has been a transition from slight average thickening (SECR of $0.16 \pm 0.10 \text{ m a}^{-1}$) in 2003-2009 to slight average thinning (SECR of $-0.10 \pm 0.10 \text{ m a}^{-1}$) in 2004-2016.

The upper elevations of Basin B have been thinning in both periods analysed by Moholdt et al. (2012a), though at a slower rate during 2003-2009. Thinning at the upper parts continues in our data, with surface elevation change rates between -0.3 and -0.7 m a^{-1} . In the steeper lower parts, while Moholdt et al. (2012a) report no clear elevation-change patterns in either period due to the noisy signal, our data show thickening close to the margin both in 2004-2016 (Fig. 5.4) and in 2012/13-2016, in the order of $\sim 1 \text{ m a}^{-1}$ during the latter period.

Recall from Table 5.3 that basin B nearly doubled its average thinning from 2004-2016 to 2012/13-2016 (SECRs of -0.28 ± 0.11 and $-0.58 \pm 0.18 \text{ m a}^{-1}$, respectively), though we pointed out that the thinning rate for 2004-2016 could have been underestimated due to the limited coverage of ICESat tracks for this basin. Our average thinning for 2004-2016 is very close to that of Moholdt et al. (2012a) for 2003-2009 (SECR of $-0.26 \pm 0.12 \text{ m a}^{-1}$), but the later is also based on ICESat data (the surface elevation data for the end of their period is nearly coincident with that of the start of our period). But both of them are remarkably lower than the estimate of Moholdt et al. (2012a) for 1988-2006, of $-1.26 \pm 0.31 \text{ m a}^{-1}$.

Our basins South, BC and Southeast are reported jointly under 'Others' in the study by Moholdt et al. (2012a). They show a decrease in average thinning rate, with surface elevation change rates of $-0.32 \pm 0.26 \text{ m a}^{-1}$ in 1988-2006 and a near-balance value of $-0.02 \pm 0.10 \text{ m a}^{-1}$ in 2003-2006. The transition to our studied period 2004-2016 is marked, with a large increase in average thinning rate (to a SECR of $-0.59 \pm 0.17 \text{ m a}^{-1}$). However, if only Basin BC were considered, its surface elevation change rate in 2004-2016 would be much larger, of $-1.31 \pm 0.33 \text{ m a}^{-1}$, due to the initiation of ice stream flow. The latter is thus responsible for the recent observed changes.

Basin C has shown widespread thinning during all periods, but much more marked during the earliest one, with average SECR of $-2.56 \pm 0.26 \text{ m a}^{-1}$ in 1988-2006, changing to $\sim -1 \text{ m a}^{-1}$ in the two most recent periods. Moholdt et al. (2012a) note that Basin C must have slowed down markedly sometime before 2003. The most rapid thinning happens in the middle part of the basin in all periods. In 2004-2016, a slight thickening is observed at particular locations of its lower parts, suggesting a slight slowdown, which also agrees with the slowdown observed in our SAR-derived velocity data between November 2016 and November 2017 (Fig. 5.2c and Table 5.2).

Basin D has also shown widespread thinning in all periods, but decreasing slowly in magnitude. This indicates sustained fast flow and large cumulative thinning. The thinning is also largest in the central part of the basin. Similarly to Basin C, certain locations of its lower part show slight thickening, indicative of slowdown, as is observed between November 2016 and November 2017 (Fig. 5.2c and Table 5.2)

Because of the above-mentioned changes in surface elevation rate change, there has been a northward migration of the zero surface elevation change contour level with respect to that determined by Moholdt et al. (2012a), which indicates a larger dynamic thinning and a larger contribution to mass loss by the marine-terminating southern and eastern drainage basins. The largest thinning at the zones of onset of ice stream flow is an additional indication of dynamic instability.

5.6.4 Comparison of calving fluxes with those of previous studies

Calving flux estimates using different techniques are available for various periods (some of them overlapping) within the last ~ 30 years. Dowdeswell et al. (2002) estimated the calving flux for September/December 1995 using SAR interferometry. Moholdt et al. (2012a) presented estimates for the period June 2000–August 2002 using image matching of Landsat scenes. For the periods 1988–2006 and 2003–2009, Moholdt et al. (2012a) estimated the ice discharge in an indirect way, subtracting from the geodetic mass balance (calculated from DEM differencing assuming Sorge’s law) an assumption on the climatic mass balance. The assumption consisted in considering Basin North (basins North and West in our study) as an analogue for the climatic mass balance of the whole ice cap (Moholdt et al., 2012a). This was justified because the northern part of the ice cap is land-terminating (and thus climatic mass balance and geodetic mass balance are equal) and its western part, though marine-terminating, seems to be dynamically inactive with no significant calving losses.

The results from these studies, together with ours, are shown in Table 5.5. We see large variations in calving flux along the whole period analysed. The lowest estimate of calving flux, of only 0.6 Gt a^{-1} , is that of Dowdeswell et al. (2002) for 1995. In spite of the large variability of calving flux during recent decades, this value seems low, though not infeasible. In the study by Dowdeswell et al. (2002), ice velocity could only be detected in the satellite look direction (northeastward), and a DEM had to be used to decompose look-angle vectors into assumed downslope movement. As Moholdt et al. (2012a) have pointed out, this is only feasible for angles within $\sim 70^\circ$ of the look direction, so potential perpendicular movements (i.e. southeast-northwest) could not be resolved. Moreover, the DEM available to Dowdeswell et al. (2002) was of limited accuracy, derived from an airborne pressure-based altimeter. This DEM was additionally used to remove the effects of ice surface topography by subtracting a synthetic interferogram generated from the DEM (Dowdeswell et al., 2002). All of this suggests large uncertainties in the estimated calving flux. Finally, we note that in the study by Dowdeswell et al. (2002) ice velocity for ice streams B, C, D is not resolved (phase coherence breaks) near the margins, so the closest point of velocity evaluation is $\sim 5 - 8 \text{ km}$ from the calving front. This implies that the highest velocities further downglacier are not used in the discharge estimate, leading to a likely underestimate. We note that a similar distance flux gate–calving front was used in the estimate by Moholdt et al. (2012a) for 2000–2002. In our study, this distance is of $\sim 1.5 - 3 \text{ km}$, just enough to avoid the possible floating tongues. Another source of uncertainty in the estimate by Moholdt et al. (2012a) for 2000–2002 is that Landsat-derived velocities were only available for the crevassed zones of ice streams B, C and D, and these velocities had to be extrapolated to the flux gates used by (Dowdeswell et al., 2002). The calving flux obtained by Moholdt et al.

Table 5.5 Estimated calving flux for the different basins of the Academy of Sciences Ice Cap for various periods, as reported in different studies. Basin North here groups our basins North and West, and 'Others' groups our basins South, BC and Southeast. We have used these names for compatibility with Moholdt et al. (2012a).

Drainage Basin	Dowdeswell et al. (2002)		Moholdt et al. (2012a)		This study
	1995 (Gt a ⁻¹)	1988-2006 (Gt a ⁻¹)	2000-2002 (Gt a ⁻¹)	2003-2009 (Gt a ⁻¹)	2016/2017 (Gt a ⁻¹)
Basin North	~ 0	~ 0	~ 0	~ 0	0.06±0.03
A	~ 0	~ 0	~ 0	~ 0	0.03±0.01
B	0.03	0.5	0.3	0.1	0.18±0.03
C	0.37	1.9	1.9	0.7	0.69±0.07
D	0.12	0.7	~ 0.7	0.5	0.44±0.05
Others	~ 0.1	~ 0.1	~ 0.1	~ 0.1	0.53±0.07
Ice Cap total	0.6	3.2	~ 3.0	1.4	1.93±0.12

(2012a) for 2000-2002, however, was very close to that obtained by the same authors for the longer period 1988-2006 using an independent method.

Regarding the indirect estimates for 1988-2006 and 2003-2009, their potential main shortcomings are the assumption of Sorge's law in the conversion from volume changes to mass changes, and the assumption made on considering the climatic mass balance of Basin North as an analogue for that of the whole ice cap. The latter one seems to be a reasonable one, in light of the various evidences about a sustained nearly zero mass balance during the last decades that will be discussed later. The use of Sorge's law involves the assumption that there is no changing firn thickness or density through time and that all volume changes are of glacier ice. Modelling experiments combined with field observations have suggested that, on the neighbouring Vavilov Ice Cap in October Revolution Island, Severnaya Zemlya, superimposed ice makes up to 40% of the total net accumulation, with the remaining 60% coming from firn that has been heavily densified by refreezing (Bassford et al., 2006b). However, this does not imply by itself density changes unless the meteorological conditions influencing surface melting change substantially. The Arctic region is known to have undergone important climatic changes during recent decades, including sustained warming (Comiso and Hall, 2014; Hartmann et al., 2013) and decline in sea-ice cover (Stroeve et al., 2011; Vaughan et al., 2013). Curiously, the summer warmth at the surface of the summit seems to have changed in the opposite direction, at least until 1998. The amount of melt layers in ice cores is a proxy for the summer warmth at the ice-cap surface (Koerner, 1977). In the analysis by Opel et al. (2009) of the deep ice core taken in 1999-2001 at the Academy of Sciences Ice Cap summit, they found a strong increase in melt-layer content at the beginning of the 20th century, which remained on a high level until about 1970 and then decreased markedly until the last year sampled by the core (1998). Even if no more recent data are available, these large temporal variations in melt-layer content indicate that the hypothesis of absence of temporal changes of firn thickness or density is not suitable for the Academy of Sciences Ice Cap.

The above limitations on the available calving flux estimates, however, do not justify the large differences observed between the various periods. Seasonal or other intra-annual

variations cannot be claimed as an explanation for the low value estimated for 1995, which was based on two snapshots in time (September and December 1995). As seen in Section 5.1, the magnitude of the seasonal velocity signal and of other intra-annual variations is insufficient to justify such deviation. Moreover, September-December is a period in the year with rather high velocities (especially September, when the highest seasonal velocities are recorded; see Figure 5.3). The calving flux in the most recent decade seems to have been rather stable, because the difference between the estimate by Moholdt et al. (2012a) for 2003-2009 and our own estimate for 2016-2017 is of $\sim 0.5 \text{ Gt a}^{-1}$, and corresponds almost entirely to the set of basins grouped under 'Others' (Table 5.5). In particular, this difference is due to the changes observed in Basin BC and we attribute it to increased dynamic thinning associated to the recent initiation of fast stream flow in Basin BC, as discussed in Section 6.2.

Let us analyse some environmental variables that could have an impact on calving flux. Summer air temperature is clearly one of them, through its influence on surface melting, and associated drainage of meltwater to the glacier bed, enhancing bed lubrication and basal sliding (Zwally, 2002). However, these accelerations are mostly short-lived and do not contribute to increased calving (Sundal et al., 2011). Air temperature could still play a role if it had an influence on sea-ice or ice mélange concentration, as these are known to affect calving, especially on glaciers confined in fjords (Moon et al., 2015; Otero et al., 2017). In particular, in the neighbouring Novaya Zemlya, Carr et al. (2014) have shown a close correspondence between glacier front position, sea-ice concentration and the number of ice-free months. The two latter, in turn, had a close correspondence with sea-surface temperature (SST). Thus, periods of higher SSTs are likely to promote glacier retreat. However, Carr et al. (2014) did not find a clear correspondence between changes in air temperature and sea ice, and only a limited correspondence between air temperatures and front position variations, concluding that air temperature changes are not a primary driver of marine-terminating glacier retreat in Novaya Zemlya (Carr et al., 2014). In a later study, Carr et al. (2017) have questioned the causal relationship between sea-ice concentration and frontal retreat, indicating that it is unclear whether the two variables simply co-vary or whether sea ice can drive ice loss, by extending the duration of seasonally high calving rates. This was based on the lack of correspondence, in their study, between step changes in the air temperature and sea-ice data and significant changes in outlet glacier retreat rates.

The possible effects of sea-ice cover on the dynamics and calving flux of the Academy of Sciences Ice Cap are expected to be much weaker, as the marine fronts are not confined in fjords where sea ice or ice mélange could exert a significant backpressure. Moreover, the seas surrounding Severnaya Zemlya are characterised by relatively thin first-year ice, because this is a region where new ice is produced and soon moved away by the oceanic currents flowing

northwards and then turning to the west in its way to the Fram Strait (Serreze and Barry, 2014). Nevertheless, Sharov and Tyukavina (2009) have pointed out that medium-term (from decadal to half-centennial) changes in glacier volumes are interrelated with the extent and duration of sea-ice cover nearby, so that slow-moving maritime ice caps grow when the sea ice cover in adjacent waters is small, and thin when the sea ice cover consolidates. This, however, would apply only in our case study to the slow-moving basins West and A. In fact, in general we did not find any clear relationship between summer (June-July-August) average temperature or sea-ice concentration that could explain the observed long-term changes in calving flux. Actually, the highest calving fluxes correspond to the period 1988-2006, which had, overall, lower air temperatures and larger end-of-September sea-ice extents than the periods 2003-2009 and 2016-2017, which had lower calving fluxes (figures C5 and C6). Regarding the two latter periods, the calving flux in 2003-2009 was half of that of 2016-2017, in spite that the summer air temperatures for the latter were some lower (though the end-of-September sea-ice extent was also lower). Only for the lowest calving flux estimate, which corresponds to particular snapshots in time (September and November 1995), we found that the sea-ice extent at the end of September that year was larger than those of the preceding and following years. The summer before the SAR image acquisition, however, was relatively warm, though it was followed by a marked drop in air temperature. This latter feature was common to the period of our SAR acquisitions (November 2016-November 2017), which was characterised by markedly decreasing temperatures following a year (and summer) relatively warm. In this case, however, the sea surrounding northern Severnaya Zemlya was virtually ice free at the end of September.

As we have not found any clear climate-related driver for the large interannual changes in calving flux observed along the last three decades, we are inclined to associate the observed dynamic instabilities with intrinsic characteristics of the ice cap, in agreement with Moholdt et al. (2012a). One of the characteristics that could influence long-term variations in front position and calving fluxes would be the complex geometry of the subglacial and seabed topography in the terminal zones of the eastern basins (C and D), as shown in Dowdeswell et al. (2002). The variations would be associated with changes in floatation conditions (Howat et al., 2007). The floating or near-floating conditions of these marginal zones has been suggested by various lines of evidence, such as the very low ice surface gradients, the strong radar returns from the ice-cap bed in several areas at the margin of the ice streams, and the large numbers of tabular icebergs observed near their margins (Dowdeswell et al., 2002).

Table 5.6 Mass balance for the main drainage basins of the Academy of Sciences Ice Cap and different periods. Basin North here groups our basins North and West, and 'Others' groups our basins South, BC and Southeast. We have used these names for compatibility with Moholdt et al. (2012a). The value given for 'Ice Cap total' in this study corresponds to the values marked with an asterisk, i.e. we have taken the values for 2012/13-2016 and, when no available, those for 2004-2016.

Drainage Basin	Moholdt et al. (2012a)		This study	
	1988-2006 (m w.e. a ⁻¹)	2003-2009 (m w.e. a ⁻¹)	ICESat-WV 2004-2016 (m w.e. a ⁻¹)	WV-WV 2012/13-2016 (m w.e. a ⁻¹)
Basin north	0.03±0.18	0.07±0.06	0±0.08*	-
A	0.14±0.23	0.14±0.09	-0.09±0.09	-0.11±0.10*
B	-1.13±0.28	-0.23±0.12	-0.25±0.10	-0.52±0.16*
C	-2.30±0.23	-0.86±0.09	-0.90±0.13	-0.86±0.23*
D	-1.57±0.26	-1.11±0.11	-0.92±0.12	-0.76±0.19*
Others	-0.29±0.23	-0.02±0.09	-0.53±0.15*	-
Ice Cap total	-0.55±0.16	-0.19±0.05	-	-0.31±0.12
	-3.06±0.89 Gt a ⁻¹	-1.06±0.28 Gt a ⁻¹	-1.72±0.67 Gt a ⁻¹	

5.6.5 Comparison of mass balance rates with those of previous studies

The various estimates of the total mass balance of the Academy of Sciences Ice Cap for different periods within the last three decades are shown in Table 5.6. All of them have been obtained by the geodetic method, through conversion of volume changes (calculated from DEM differencing) to mass changes, assuming in all cases Sorge's law. The limitations involved in the latter assumption have already been discussed in the previous section. Another shortcoming stems from the limited coverage of the ice cap by the ICESat tracks (Fig. 5.4), and the associated interpolation errors (Zwally et al., 2011). Our estimates and their associated errors, though, are in line with other ICESat-based studies of Arctic regions and, in particular, with those by Moholdt et al. (2012a) for the Academy of Sciences Ice Cap. Possible changes in ice cap area, if significant, would be another source of uncertainty in the geodetic mass balance estimates. Dowdeswell et al. (2002) reported an ice cap area of 5575 km², based on Landsat images of 1988. Moholdt et al. (2012a), using multitemporal satellite imagery from Corona and Landsat acquired between 1962 and 2010, concluded that there no clear temporal trends in the terminus fluctuations of the Academy of Sciences Ice Cap, and calculated a total marine-terminating glacier area loss of 5 km² between 1988 and 2009, but with several examples of both advance and retreat. Our own observations, using Landsat-7 and Sentinel-2 optical images of July 2002 and March 2016, respectively, show local advances and retreats of the eastern and southern marine margins of up to $\sim 1 - 2$ km with respect to the margins of Moholdt et al. (2012a), but the net change in area is negligible and thus we have used the same ice cap area of 5570 km². Moholdt et al. (2012a) noted that, assuming an average ice thickness of 100 m at the marine termini, the rate of ice-volume loss between 1988 and 2009 was of only 0.02 km³ a⁻¹, which is not significant in terms of ice-cap mass balance.

The total mass balance starts at ~ -3 Gt a⁻¹ for 1988-2006, then increases to ~ -1 Gt a⁻¹ in 2003-2009 and afterwards decreases to ~ -2 Gt a⁻¹ in 2012/13-2016. Both our estimates of *climatic* mass balance for 2004-2016, and those of Moholdt et al. (2012b) for 1988-2006 and 2003-2009, indicate a near-zero balance. The limited earlier observations

of climatic mass balance available, for the Vavilov Ice Cap on October Revolution Island some 120 km to the south, also indicate a near-zero average balance of -0.03 m a^{-1} for 10 years during the periods 1975-1981 and 1986-1988 (Barkov et al., 1992), and a similar value of -0.02 m a^{-1} for the complete period 1974-1988 from mass-balance modelling experiments (Bassford et al., 2006b). We may thus consider that the climatic mass balance of the Academy of Sciences Ice Cap has been close to zero on average during at least the last three decades. This implies that the net total mass balance must be equal, and opposite in sign, to the calving flux. This is confirmed by the comparison between the ice cap totals (in Gt a^{-1}) of Table 5.5 and Table 5.6. In fact, the only minor difference ($\sim 0.3 - 0.4 \text{ Gt a}^{-1}$) corresponds to the period 2003-2009, and it is due to the use by Moholdt et al. (2012b) of Basin North as an analogue for the climatic mass balance of the whole ice cap (the slightly positive climatic mass balance of Basin North times the area of the whole ice cap accounts for the mentioned difference).

The climatic mass balance estimates by Barkov et al. (1992) for the Vavilov Ice Cap, though close to zero on average, had a large interannual variability ranging from -0.63 to 0.46 m w.e. The further modelling experiments by Bassford et al. (2006b) not only confirmed this large variability but also that it is caused primarily by variations in the amount of summer melting, in turn driven by summer temperatures and further amplified by the albedo feedback, as shown in other Arctic archipelagos such as Svalbard (Ostby et al., 2017). However, in the context of our case study these yearly variations have limited interest, since we only count on average mass balance estimates over periods of several years, up to more than a decade. Let us thus focus on the factors that could control long-term changes and trends in mass balance; in particular, on the climatic mass balance.

Summer air temperature and precipitation are the most evident controlling factors. Using NCEP-NCAR and ERA-Interim data for Novaya Zemlya and Severnaya Zemlya over 1995-2011, Zhao et al. (2014) have analysed the influence of summer (June-September) mean 850 hPa geopotential height temperature on snowmelt. They analysed the trends of both total melt days (TMD) and melt offset date (MOD). For Severnaya Zemlya, the temperature trends 1995-2011 were of $0.80^\circ\text{C/decade}$ (NCEP-NCAR, $p\text{-value} < 0.05$) and $0.88^\circ\text{C/decade}$ (ERA-Interim, $p\text{-value} = 0.065$), and Zhao et al. (2014) found a positive correlation between mean TMD and the average June-September NCEP-NCAR air temperature at 850 hPa, with slope of the linear regression of $10 \text{ days } ^\circ\text{C}^{-1}$ ($r=0.843$, $p\text{-value} < 0.0001$). Using simple regression, they also found that Severnaya Zemlya's TMD is significantly anti-correlated to Laptev Sea ($r=-0.735$, $p\text{-value} < 0.001$) and Kara Sea ($r=-0.678$, $p\text{-value} < 0.003$) September sea-ice extent. However, since sea-ice extent and glacier surface melting can co-respond to the regional temperature increase, Zhao et al. (2014) additionally used partial

correlation analysis to remove the large-scale influence of air temperature on both variables. After removing these effects, partial correlation analysis suggested that glacier melt on Severnaya Zemlya is still statistically anti-correlated to Laptev Sea and Kara Sea sea-ice extent. The explanation is that reduced offshore sea-ice concentration, i.e. increased open-water fraction, can enhance onshore advection of sensible and latent heat fluxes (Rennermalm et al., 2009). However, even if long-term changes in summer (and annual) temperatures have been observed during our period of analysis (Fig. C6), and regional sea-ice concentration has also shown a clear decreasing trend (Fig. C5), these changes seem to not have exerted an appreciable impact on the long-term climatic mass balance estimates, which remain close to zero. As noted by Zhao et al. (2014), the explanation could be that sea-ice reduction exposes larger areas of open water in summer to evaporation and change the large-scale atmospheric circulation, which has been shown to intensify summer precipitation over the Arctic (Francis, 2013; Serreze et al., 2012).

In particular, the influence of sea-ice concentration on precipitation has been observed for the Academy of Sciences Ice Cap. The analyses by Opel et al. (2009) of the deep ice core drilled at the ice cap summit in 1999-2001 has revealed this influence. Deuterium excess (the difference between the two stable water-isotope ratios $\delta^{18}\text{O}$ and δD) in precipitation depends mostly on the evaporation conditions in the moisture source region, and to a lesser degree on condensation temperatures. The main controlling factors are the relative air humidity and the sea-surface temperature (SST) and, to a lesser extent, wind speed during evaporation. Based on relationship between deuterium excess and SST, Opel et al. (2009) noted that in hemispherical warmer periods the Academy of Sciences Ice Cap receives more precipitation from moisture evaporated at lower SSTs, for example due to a northward shift of the moisture source. Since most precipitation on Severnaya Zemlya is brought by air masses moving from the south and southwest, the Kara Sea seems to be a regional moisture source and its sea-ice cover the main controlling factor for summer and autumn evaporation. Lower sea-ice extent in the Kara Sea would allow higher evaporation rates and enhance the contribution of regional moisture to precipitation over the Academy of Sciences Ice Cap. Moholdt et al. (2012a) have analysed some evidences of this precipitation increase for Novaya Zemlya and Severnaya Zemlya, finding a slightly higher precipitation rate in 2004-2009 with respect to the 1980-2009 mean, especially for Novaya Zemlya. However, the close-to-zero climatic mass balance of Severnaya Zemlya suggests that the recent precipitation anomaly is very likely to be real also for this archipelago, as it provides the most reasonable mechanism to counterbalance the observed increasing melt trend. Summarizing, the near-equilibrium climatic mass balance of the Academy of Sciences Ice Cap (and Severnaya Zemlya in general) would be the result of two opposing effects. On one hand, sea-ice cover loss would

enhance precipitation by exposing larger areas of open water to evaporation. On the other, these larger areas of open water would allow onshore advection of heat fluxes from warming mixed ocean layers, accelerating surface melt on the ice cap. With the climatic mass balance kept near zero, the role of the calving flux is critical in determining the total mass balance of the Academy of Sciences Ice Cap.

5.6.6 Relative shares of surface ablation and frontal ablation to total ablation

Even if we know that the climatic mass balance of the Academy of Sciences Ice Cap has been close to zero during the last decades, it is of interest to know the share to the total mass losses by surface ablation and frontal ablation. To do so we need an estimate of the accumulation, since zero climatic mass balance means that surface accumulation and ablation are equal in magnitude but of opposite sign. We can roughly estimate the total accumulation over the ice cap as done by Dowdeswell et al. (2002), based on the measured net accumulation at the ice cap summit and its variation with altitude. At the summit, analysis of an ice core detected the layers of maximum radioactivity (in terms of Cesium¹³⁷) corresponding to the 1963 atmospheric nuclear tests and to the 1986 Chernobyl event. The resulting average net accumulation rates were 0.45 m w.e. a⁻¹ from 1963 to 1999, and 0.53 m w.e. a⁻¹ from 1986 to 1999 (Fritzsche et al., 2002). Later analyses by Fritzsche et al. (2005) gave an average accumulation rate of 0.46 m w.e. a⁻¹ over 1956-1999 based on stable-isotope investigations. All of these values are also in agreement with the the mean annual net mass balance of 0.43-0.44 w.e. a⁻¹ observed by Zagorodnov for 1986/87 using structural-stratigraphic methods. However, they are in disagreement with the annual-layer thickness of 0.26-0.28 m suggested by Klementyev et al. (1988) and used by Kotlyakov et al. (1990) for dating the Academy of Sciences ice core drilled in 1986/87. On the other hand, measurements elsewhere in Severnaya Zemlya suggest that annual precipitation may vary with altitude from 0.25 to 0.45 m w.e. a⁻¹ (Bryazgin and Yunak, 1988). Assuming, as done by Dowdeswell et al. (2002), a value of 0.30 m w.e. a⁻¹ as average accumulation over the entire ice cap (because, as shown by Bryazgin and Yunak (1988), the accumulation is expected to decrease with altitude), we obtain a total accumulation of 1.67 Gt a⁻¹. The total ablation will thus be of -1.46 Gt a⁻¹. As the net mass balance is equal to the sum of surface accumulation, surface ablation and frontal ablation (the two latter terms being negative quantities), and the total mass balance over 2012/13-2016 is of -1.72 ± 0.67 Gt a⁻¹ (-1.74 Gt a⁻¹ if considered over 2004-2016) this gives a surface ablation of -1.46 Gt a⁻¹ (-1.49 for 2004-2016). Total ablation (surface+frontal) will thus be of -3.18 Gt a⁻¹

(2012/13-2016) or -3.21 Gt a^{-1} (2004-2016), which means that frontal ablation represents, in average, $\sim 54\%$ of the mass losses over 2012/13-2016 ($\sim 55\%$ over 2004-2016), with the remaining 46% (45%) corresponding to surface ablation.

Even if we consider for the ice-cap averaged accumulation rate its upper bound, given by the net accumulation rate at the ice cap crest of $0.46 \text{ m w.e. a}^{-1}$ provided by Fritzsche et al. (2005), the total accumulation over the ice cap would be of 2.56 Gt a^{-1} , which implies a surface ablation of -2.35 Gt a^{-1} over 2012/13-2016 (-2.38 Gt a^{-1} over 2004-2016). The total ablation (surface+frontal) would then be of -4.07 Gt a^{-1} over 2012/13-2016 (-4.10 Gt a^{-1} over 2004-2016), which means that frontal ablation represents $\sim 42\%$ of the total ablation, with the remaining 58% corresponding to surface ablation. Calving losses are, therefore, a substantial component of the total mass losses in the Academy of Sciences Ice Cap (Dowdeswell et al., 2002).

5.7 Conclusions

The following main conclusions can be drawn from our analysis:

1. During the period November 2016-November 2017, the marine-terminating margins of the Academy of Sciences Ice Cap remained nearly stable, so ice discharge and calving flux are equivalent in our case study, at $1.93 \pm 0.12 \text{ Gt a}^{-1}$.
2. This estimated ice discharge is quite close to that determined by Moholdt et al. (2012a) for 2003-2009, of $\sim 1.4 \text{ Gt a}^{-1}$, and most of the difference can be attributed to the initiation of ice stream flow in Basin BC, which accounts for $0.41 \pm 0.05 \text{ Gt a}^{-1}$. The initiation of ice stream flow in basin BC took place after 2002, and most likely within the period 2003-2009.
3. The long-term (interannual/interdecadal) variations of ice discharge during the last three decades have been large, between 0.6 and 3.2 Gt a^{-1} . The lack of clear environmental drivers for the observed changes suggests that this dynamic instability is associated with intrinsic characteristics of the ice cap, as proposed by Moholdt et al. (2012a). We suggest that this instability could be caused by the long-term changes in flotation conditions associated with the complex geometry of the subglacial and seabed topography in the terminal zones of the fast-flowing eastern basins (B and C).
4. The purely seasonal variability in ice velocity (and correspondingly in ice discharge) during the period November 2016-November 2017 is of up to $\sim 10\%$ of the yearly

averaged velocity values (20% peak-to-peak), but the intra-annual variability, considering changes other than seasonal, has average values of up to 16%, depending on basin, but with maximum values of up to 32% of the yearly-averaged velocities (32% and 64% peak-to-peak). This gives an idea of the errors that could be incurred if the ice discharges calculated for particular snapshots in time were extrapolated to yearly values.

5. The intra-annual variations in ice velocity during the summer seem to be associated with episodes of surface melting, routing of meltwater to the glacier bed and enhanced basal sliding. Maximum velocities are reached during a sustained high-velocity event lasting for about two months and followed by an abrupt slow-down in velocities well before the end of the melt season. This abrupt drop in velocities could be due to the transition from a hydraulically inefficient distributed drainage system to an efficient channelized system. The explanation for the short-term intra-annual velocity variations during the winter, however, remains elusive.
6. The average total mass balance of the ice cap for the period 2012/13-2016, determined by the geodetic method, is of $-1.72 \pm 0.27 \text{ Gt a}^{-1}$ (equivalent to $-0.31 \pm 0.05 \text{ m w.e. a}^{-1}$ over the whole ice cap area).
7. The average climatic mass balance of the ice cap during the period 2012-2016 (very similar to that of the period 2004-2016), is not significantly different from zero, at $0.21 \pm 0.30 \text{ Gt a}^{-1}$, or equivalently $0.04 \pm 0.05 \text{ m w.e. a}^{-1}$. This is in line with the scarce in-situ observations in the region during the 1970s and 1980s and with the indirect estimates from geodetic mass balance and calving flux by other authors for the periods 1988-2006 and 2003-2009. The average climatic mass balance has remained nearly around zero during the last four decades, and the total mass balance has therefore been governed by the calving flux.
8. The near-equilibrium climatic mass balance under a scenario of regional warming and observed increase in the total melt days during the period 1995-2011 can only be explained by a concurrent increase in accumulation. As suggested by Zhao et al. (2014), loss of sea-ice cover would enhance precipitation by exposing larger areas of the ocean to precipitation, while these larger areas of open water would allow onshore advection of heat fluxes from warming mixed ocean layers.
9. The current average total ablation over 2012-2016 (surface ablation plus frontal ablation) is estimated to be of about -3.18 Gt a^{-1} , which means that frontal ablation (mostly calving, in our case study) represents $\sim 54\%$ of the mass losses, with the

remaining $\sim 46\%$ corresponding to surface ablation. Even considering a scenario of extreme surface accumulation, the shares of total ablation by frontal ablation and surface ablation would be $\sim 42\%$ and $\sim 58\%$, respectively, indicating that calving losses are a significant contributor to the ice mass losses from the ice cap.

Summarising, with a climatic mass balance averaging to zero over the last four decades, in spite of regional warming, the total mass balance of the Academy of Sciences Ice Cap is mostly driven by ice discharge, which does not seem to respond to environmental changes but to intrinsic characteristics of the ice cap such as the subglacial and seabed topography near the marine-terminating margins.

Chapter 6

Conclusions and Outlook

The main conclusions resulting from the present PhD thesis work have already been presented in the last sections of chapters 3, 4 and 5. In this final chapter, for the convenience of the readers we present a compilation of such conclusions, restructured separating the conclusions referred to methodological aspects from those referred to the applications of the methodologies developed in this thesis (as well as other existing state-of-art methodologies) to glaciers of the Canadian High Arctic and of the Russian Arctic. Similarly, some outlook was already sketched in chapters 3 and 4. We also compile those comments here, but also discuss briefly some further possibilities regarding lines of future research.

6.1 Methodological conclusions

6.1.1 Offset tracking using ascending and descending passes

SAR intensity offset tracking is a well established technique in the scientific community (Pritchard, 2005; Strozzi et al., 2008, 2002; Werner et al., 2005). SAR Differential Interferometry is also a widespread and accepted methodology. Joughin et al. (1998) estimated the 3D velocity vector using the D-InSAR technique over ascending and descending acquisitions and the surface-parallel flow assumption. Fallourd et al. (2010) inferred the 3D glacier displacement field applying the intensity offset tracking methodology to ascending and descending passes for alpine glaciers. The latter study used both the azimuth and range offsets within a least-squares approach for estimating the glacier surface velocity field.

Sentinel-1 TOPS acquisition mode generates a SAR image with a resolution of 20 metres in the azimuth direction and 5 metres in the range direction. This disparity in resolutions translates into a worse result in the derived surface velocity products that use the along-track direction data. The Sentinel-1 observation scenario, with its acquired ascending and

descending scenes in several Earth's regions (Potin et al., 2014), allows for a better use of the retrieved scenes. This wealth of SAR satellite imagery allowed us using only the estimated intensity offset tracking displacements in the range direction from ascending and descending passes, which ultimately led to an improvement in the estimated surface velocity field (Sánchez-Gómez and Navarro, 2017), as shown in this thesis. Furthermore, the ionospheric effect strongly influences the derived surface velocity product in the azimuth direction, especially in the regions closer to the magnetic poles (Gray et al., 2000; Wegmüller et al., 2006). Therefore, the use, as proposed in this thesis, of only range offsets prevented the presence of azimuth streaks in our final velocity product.

There are three main error sources in intensity-tracking methodology: 1) the matching procedure (which is a function of the co-registration between images, template size, and the quality of the image features) (Nagler et al., 2015); 2) the ionospheric effect and its influence on the azimuth offsets in the form of azimuth streaks (Gray et al., 2000; Wegmüller et al., 2006); and 3) the geocoding error (with the high quality ephemerides of Sentinel-1 data this error is reduced to the quality of the DEM used for the topographic correction (Nagler et al., 2015)). With the use of the technique proposed in this thesis, these error sources were ultimately reduced to only the matching-related uncertainties.

The proposed intensity offset tracking algorithm demonstrated its ability to improve the offset tracking velocity product both in its resolution and in avoiding the azimuth streaks. The specific figures given in Sánchez-Gómez and Navarro (2017) (and Chapter 3 of this thesis) show a noticeable improvement in the RMSE over the traditional intensity offset tracking technique. The error analysis showed that the latter approach casts a RMSE value of $\sim 0.051 \text{ m d}^{-1}$ (with mean error of $\sim 14.2 \text{ m yr}^{-1}$) while the former drops to 0.012 m d^{-1} (with mean error of $\sim 3.3 \text{ m yr}^{-1}$) (Sánchez-Gómez and Navarro, 2017).

Finally, we acknowledge that the intensity offset tracking methodology is well suited for estimating glacier surface velocities on regions with large displacements and velocity gradients due to its better resilience to signal de-correlation. On the other hand, the D-InSAR interferometry shows a good performance in slowly moving areas and small velocity gradients, providing high resolution, smooth and continuous velocity fields for land-terminating glaciers. This illustrates the complementarity of both techniques. These observations suggest the interest to develop a hybrid velocity product combining DInSAR and offset tracking results, as will be discussed in the Outlook section.

6.1.2 Error estimates in ice discharge to the ocean

In Chapter 4 we analysed the contributions of the various error components involved in the estimation of ice discharge through predefined flux gates, distinguishing two cases: 1)

ice-thickness data is available for glacier cross-sections close to the glacier terminus, and 2) ice-thickness data is only available along the glacier centreline. In the latter case, we analysed the performance of three different U-shaped cross-sectional approaches, and gave hints for the choice of a suitable location of the flux-gate. We list below the main conclusions drawn from our study:

Regarding the relative contribution from the various error components, and focusing on the case study of the Canadian High Arctic glaciers:

1. The velocity field is the dominant source of error for small and medium-size glaciers (discharge $< 100 \text{ Mt a}^{-1}$ with low velocities ($< 100 \text{ m a}^{-1}$)).
2. For large glaciers (discharge $> 100 \text{ Mt a}^{-1}$) with high velocities ($> 100 \text{ m a}^{-1}$) the error in cross-sectional area becomes the main contributor to the total error. This stresses the need of measuring radar cross-sectional profiles for the largest glaciers.
3. The bias (systematic error) implied by glacier thinning/thickening between the radar and SAR acquisitions is variable according to subregions, oscillating between 8% and -8% of the discharge value over the period of 2-5 years between acquisitions, with an average of the absolute values of $\sim 3\%$. Temporally coincident radar and SAR acquisitions are recommended to reduce the effect of the bias, especially for the largest glaciers (Trinity, Wykeham, Belcher, Ekblaw), which contribute to most of the ice discharge in the region.

Of the above conclusions, two have particular relevance because are applicable to any glacier region: the interest of measuring radar cross-sectional profiles, especially for the largest glaciers, and the importance of using radar and SAR acquisitions close in time to each other; otherwise, it is crucial to correct for ice-thickness changes between both acquisitions.

Concerning the performance of the U-shaped cross-sectional approaches:

1. If the radar flight line is not too far from the glacier centreline, the off-centred parabolic approach shows the lowest bias and acceptable standard deviation, so it is the recommended approach.
2. The centred parabolic approach shows nearly constant standard deviation, but is more strongly biased than the off-centred approach for common distances flight line-centreline.
3. The off-centred quartic approach shows large variable bias and its standard deviation, though nearly constant, is large.

These last conclusions, though based on our case study of the Canadian High Arctic, are of a general nature and could be assumed applicable to other glacier regions.

6.2 Ice discharge to the ocean in Arctic Regions

6.2.1 Canadian High Arctic

Regarding the comparison of the ice discharge results for Canadian High Arctic glaciers presented in Chapter 4 of this thesis, referred to the winters of 2016 and 2017, with the results for 2015 presented by van Wychen et al. (2016, 2017) and Millan et al. (2017), in general we see comparable results with only small differences. When there is a difference between the results by these authors (usually a larger estimate by Millan et al. (2017)), our data in general agree better with those by van Wychen et al. (2016, 2017), so we will comment on the comparison with the latter to gain some understanding on the interannual changes. There is an increase of ice discharge from the main glaciers (Trinity and Wykeham) of the Prince of Wales Icefield from 2015 to 2016, by 5% and 20%, respectively, but this is followed by a decrease in 2017, by 10% and 15% respectively, consistent with the reduction in surface velocities of the main Canadian Arctic tidewater glaciers pointed out by Strozzi et al. (2017). Among the largest glaciers, only Belcher Glacier, in the Devon Ice Cap, maintains similar discharges during the period 2015-2017. Two small glaciers show significant decreases in ice discharge. Tanquary, part of the Prince of Wales Icefield, changes by 70% from 2015 to 2016, and a further 20% from 2016 to 2017. Good Friday Bay, part of the Steacie Ice Cap, decreases by 80% from 2015 to 2016, remaining stable in 2017.

Some additional conclusions referred to aspects that deserve further investigation in the case of Canadian High Arctic glaciers are sketched in the Outlook section.

6.2.2 Russian Arctic. Academy of Sciences Ice Cap

As presented in Chapter 5, the following main conclusions can be drawn from our analysis referred to the Academy of Sciences Ice Cap in Severnaya Zemlya, Russian Arctic:

1. During the period November 2016-November 2017, the marine-terminating margins of the Academy of Sciences Ice Cap remained nearly stable, so ice discharge and calving flux are equivalent in our case study, at $1.93 \pm 0.12 \text{ Gt a}^{-1}$.
2. This estimated ice discharge is quite close to that determined by Moholdt et al. (2012a) for 2003-2009, of $\sim 1.4 \text{ Gt a}^{-1}$, and most of the difference can be attributed to the

initiation of ice stream flow in Basin BC, which accounts for $0.41 \pm 0.05 \text{ Gt a}^{-1}$. The initiation of ice stream flow in basin BC took place after 2002, and most likely within the period 2003-2009.

3. The long-term (interannual/interdecadal) variations of ice discharge during the last three decades have been large, between 0.6 and 3.2 Gt a^{-1} . The lack of clear environmental drivers for the observed changes suggests that this dynamic instability is associated with intrinsic characteristics of the ice cap, as proposed by Moholdt et al. (2012a). We suggest that this instability could be caused by the long-term changes in flotation conditions associated with the complex geometry of the subglacial and seabed topography in the terminal zones of the fast-flowing eastern basins (B and C).
4. The purely seasonal variability in ice velocity (and correspondingly in ice discharge) during the period November 2016-November 2017 is of up to $\sim 10\%$ of the yearly averaged velocity values (20% peak-to-peak), but the intra-annual variability, considering changes other than seasonal, has average values of up to 16%, depending on basin, but with maximum values of up to 32% of the yearly-averaged velocities (32% and 64% peak-to-peak). This gives an idea of the errors that could be incurred if the ice discharges calculated for particular snapshots in time were extrapolated to yearly values.
5. The intra-annual variations in ice velocity during the summer seem to be associated with episodes of surface melting, routing of meltwater to the glacier bed and enhanced basal sliding. Maximum velocities are reached during a sustained high-velocity event lasting for about two months and followed by an abrupt slow-down in velocities well before the end of the melt season. This abrupt drop in velocities could be due to the transition from a hydraulically inefficient distributed drainage system to an efficient channelized system. The explanation for the short-term intra-annual velocity variations during the winter, however, remains elusive.
6. The average total mass balance of the ice cap for the period 2012/13-2016, determined by the geodetic method, is of $-1.72 \pm 0.27 \text{ Gt a}^{-1}$ (equivalent to $-0.31 \pm 0.05 \text{ m w.e. a}^{-1}$ over the whole ice cap area).
7. The average climatic mass balance of the ice cap during the period 2012-2016 (very similar to that of the period 2004-2016), is not significantly different from zero, at $0.21 \pm 0.30 \text{ Gt a}^{-1}$, or equivalently $0.04 \pm 0.05 \text{ m w.e. a}^{-1}$. This is in line with the scarce in-situ observations in the region during the 1970s and 1980s and with the indirect estimates from geodetic mass balance and calving flux by other authors for the

periods 1988-2006 and 2003-2009. The average climatic mass balance has remained nearly around zero during the last four decades, and the total mass balance has therefore been governed by the calving flux.

8. The near-equilibrium climatic mass balance under a scenario of regional warming and observed increase in the total melt days during the period 1995-2011 can only be explained by a concurrent increase in accumulation. As suggested by Zhao et al. (2014), loss of sea-ice cover would enhance precipitation by exposing larger areas of the ocean to precipitation, while these larger areas of open water would allow onshore advection of heat fluxes from warming mixed ocean layers.
9. The current average total ablation over 2012-2016 (surface ablation plus frontal ablation) is estimated to be of about -3.18 Gt a^{-1} , which means that frontal ablation (mostly calving, in our case study) represents $\sim 54\%$ of the mass losses, with the remaining $\sim 46\%$ corresponding to surface ablation. Even considering a scenario of extreme surface accumulation, the shares of total ablation by frontal ablation and surface ablation would be $\sim 42\%$ and $\sim 58\%$, respectively, indicating that calving losses are a significant contributor to the ice mass losses from the ice cap.

Summarising the above conclusions, with a climatic mass balance averaging to zero over the last four decades, in spite of regional warming, the total mass balance of the Academy of Sciences Ice Cap is mostly driven by ice discharge, which does not seem to respond to environmental changes but to intrinsic characteristics of the ice cap such as the subglacial and seabed topography near the marine-terminating margins.

6.3 Outlook

We have seen in Chapter 3 that the intensity offset tracking methodology is well suited for estimating glacier surface velocities on regions with large displacements and velocity gradients due to its better resilience to signal de-correlation. On the other hand, the D-InSAR interferometry has shown a good performance in slowly moving areas and small velocity gradients, providing high resolution, smooth and continuous velocity fields for land-terminating glaciers. We also noticed the complementarity of both techniques (Hu et al., 2014; Luckman et al., 2007). We suggest here the possibility to develop a hybrid velocity product combining DInSAR and offset tracking results as suggested by Joughin (2002) and Liu et al. (2007). The second Sentinel-1 constellation satellite (Sentinel-1B) will improve the interferometric capabilities of the Sentinel-1 constellation by minimizing

de-correlation and the amount of movement between acquisitions (Joughin et al., 1998). Moreover, we acknowledge the possibility to apply the proposed ascending and descending passes algorithm to speckle tracking. This approach opens the door for obtaining surface velocity fields within featureless accumulation areas, while at the same time avoiding the ionospheric effects typical of Antarctic and Arctic areas (Nagler et al., 2015).

Regarding the error estimates in ice discharge, in Chapter 4 of this thesis we have focused on the analysis of the errors in discharge through given flux gates and at a given time, and the quantification of the influence of the various error components. Consequently, in our view, most of the work remaining to be done corresponds to the approximation of the frontal ablation of tidewater glaciers by the ice discharge calculated at flux gates close to the calving front. Aside from seasonality and interannual variability considerations (which we have considered for the Academy of Sciences Ice Cap in Chapter 5, but not for the Canadian High Arctic glaciers in Chapter 4), we believe that two critical aspects deserve further investigation in the case of the Canadian High Arctic glaciers: 1) the surface mass balance between the flux gate location and the calving front, and 2) the front position changes. If the surface mass balance effects are ignored, the overestimate of frontal ablation is expected to be large for Canadian High Arctic glaciers, because of the strongly negative recent surface mass balance of the Canadian Arctic, with values between -1.5 and -2.0 m w.e. a^{-1} at the lowermost part of the tidewater glaciers during 2003-2009 (Gardner et al., 2011). Our own preliminary estimates for the studied glaciers suggest typical overestimates by $\sim 30\%$ of the calculated ice discharge, reaching up to 50% for individual glaciers. Regarding the effect of the terminus advance/retreat, it is difficult to quantify for Canadian High Arctic glaciers, due to the pulsating behaviour of many of them. As noted by van Wychen et al. (2016), pulse-type and surge-type glaciers have some common characteristics, such as periods of speedup and slowdown, and terminus advance coincident with acceleration, but their key difference is that all of the velocity variability of the pulse-type glaciers appears to be restricted to their lowermost terminal region, which is grounded below sea level. Very little interannual variability is observed upglacier from this area. This poses difficulties in the approximation of frontal ablation by ice discharge through flux-gates, as these estimates are heavily dependent on flux-gate location for the pulsating glaciers. This stresses the importance of monitoring the terminus advance/retreat for the glaciers in this region, as has been done during the last decades (Burgess and Sharp, 2004; Burgess et al., 2005; van Wychen et al., 2016; Williamson et al., 2008). In the case of the Academy of Sciences Ice Cap, the flux gates were located closer to the calving fronts, and the front position changes during the recent decades have been nearly negligible, so the corrections for mass balance between flux gate and calving

front, and for front advance and retreat, are much less relevant as compared with the case of the Canadian High Arctic glaciers.

The Sentinel-2 satellite constellation provides further possibilities for retrieving ice surface velocities using optical imagery (Kääb and Vollmer, 2000). Surface velocity fields derived from optical satellite imagery present their own advantages and disadvantages (Kääb et al., 2005). Commonly, ice-surface features are present in most tidewater glacier's termini, allowing to infer deformations using cross-correlation. This permits to extract successfully ice-surface velocity fields, provided that no cloud cover is present in the image pairs under consideration. The resolution of the surface velocities depends on the resolution of the original optical imagery from which they are inferred. Furthermore, summer changes in the spectral signature of these surface features can also be avoided using several image-enhancing techniques (Heid and Kääb, 2012). Therefore, the temporal resilience to de-correlation of optical image-matching methods rely mostly on the effects of shear and deformation on ice surface features, provided that no other weather phenomena hinders the acquisition of such elements (e.g. temporal snow cover or cloud cover). This wealth of data, available for processing and obtaining velocity products, together with the already-existing SAR Sentinel-1 imagery, build up an excellent pool of data. We foresee a future possibility of combining both types of velocity products, as suggested by Kargel et al. (2014). The ice-surface velocities obtained in such a way would benefit from a higher temporal resolution and would also inherit a greater robustness regarding both spatial coverage, quality and confidence.

The spatial resolution of several new SAR active sensors such as TerraSAR-X allows to analyse the surface deformation of smaller-size glaciers and ice caps present in the peripheries of Greenland and Antarctica and in neighbouring islands and archipelagos. Therefore, we also see a possibility to deepen and update the already-existing analyses of glacier dynamics made in the South Shetland Islands using the ALOS-PALSAR sensor (Osmanoğlu et al., 2013, 2014), using in the near future the increased spatial resolution provided by TerraSAR-X. Analysis of dynamics and surface mass balance of glaciers using combinations of satellite Laser altimeters such as ICESat and very high-resolution optical sensors capable of generating highly-accurate DEMs are nowadays becoming common (Melkonian et al., 2016; Noh and Howat, 2015; Noh et al., 2016; Sánchez-Gómez et al., 2018; Willis et al., 2015). The data retrieved by these sensors, in combination with ice-surface velocities inferred from SAR, provide state-of-the-art information for assessing and evaluating the dynamics and physical parameters of any glacier.

Finally, we open the door to using the available satellite information for applying inversion modelling in ice-thickness estimation (Farinotti et al., 2017). The use of the different types of satellite data which are available nowadays, and its possible combinations, together with the

use of state-of-the-art software tools such as the Ice Sheet System Model (Larour et al., 2012) or Elmer Ice (Gagliardini et al., 2013), can lead to substantial advance in the understanding of glaciers dynamics, and the roles of the various physical processes involved. There is plenty of research on how to optimally apply these modelling software tools on satellite-retrieved data for inversion modelling. A couple of remarkable examples are the ice-thickness inversion scheme developed by Morlighem et al. (2011) and the basal friction inversion presented in Morlighem et al. (2013).

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Appendix A

Supplementary Materials. Chapter 3

Fig. A.1 Ice velocities using offset tracking for southern Ellesmere Ice Caps. Velocity map overlaid on top of Sentinel-1 SAR images. Velocities calculated from images acquired between the 22nd and 12th of March 2016. Background images acquired between the 22nd and 29th of February 2016.

Fig. A.2 Ice velocities using D-InSAR for southern Ellesmere Ice Caps. Velocity map overlaid on top of Sentinel-1 SAR images. Velocities calculated from images acquired between the 10th and 29th of March 2016. Background images acquired between the 22nd and 29th of February 2016.

Appendix B

Supplementary Materials. Chapter 4

Fig. B.1 Ellesmere, Axel Heiberg and Devon islands, Nunavut, Canadian Arctic (Wessel and Smith, 1996). The glacier outlines are from the Randolph Glacier Inventory (RGI) version 5.0 (Pfeffer et al., 2014).

Fig. B.2 Agassiz Ice Cap and Northern Ellesmere and Prince of Wales Icefields (Ellesmere Island), and Müller and Steacie Ice Caps (Axel Heiberg Island), Nunavut, Canadian Arctic.

Table B.1 Characteristics of the main glaciers of Ellesmere, Axel Heiberg and Devon islands, Nunavut, Canadian Arctic.

Glacier	Latitude	Longitude	Area (m ²)	Profile length (m)	Mean thickness (m)	Mean surface velocity (m a ⁻¹)
Prince of Wales Icefield						
North 1	78.94	-78.05	714000	2594	275	25
North 2	78.85	-78.24	900500	3216	280	36
Stygge	78.77	-78.24	330000	3445	96	28
Leffert	78.69	-74.92	340000	3371	101	59
Ekblaw	78.51	-76.71	1056000	4047	261	134
Tanquary	78.46	-76.08	850000	3316	256	21
Cadogan	78.23	-76.94	890000	4995	178	53
Trinity	77.97	-78.57	1638000	7150	229	771
Wykeham	77.89	-78.61	1104500	5909	187	486
South Margin	77.71	-77.88	7544400	58178	130	11
South	77.33	-79.59	350000	2147	163	107
Palisade	77.39	-80.99	180000	2153	84	42
Devon Ice Cap						
Sverdrup	75.72	-83.18	587000	2506	234	25
Eastern	75.79	-82.00	338000	2307	147	94
Belcher	75.67	-81.39	730000	4701	155	284
Fitzroy	75.45	-80.46	535000	3256	164	189
East	75.07	-80.41	550000	5070	108	36
South East 1-2	74.98	-80.44	978000	10420	94	41
South Crocker	74.85	-83.20	626000	3369	186	75
North Crocker	74.91	-83.62	700000	4326	162	12
Northern Ellesmere Icefield						
Disraeli North	82.84	-70.79	504000	2779	181	14
Disraeli	82.67	-72.50	314000	2020	155	38
M'Clintock	82.43	-76.15	108000	1920	56	33
Milne	82.44	-80.22	1230000	3823	322	44
Vanier	82.14	-80.75	290000	2440	119	33
DeVries	82.01	-79.60	565000	2890	196	2
Yelverton	81.84	-79.43	658000	4393	150	141
Otto	81.30	-84.70	445000	4200	106	2
Marine	82.24	-81.74	385000	2503	154	7
Marine North	82.41	-82.56	703000	4610	152	11
Agassiz Ice Cap						
Tuborg	80.89	-76.14	780000	3367	232	49
Antoinette	80.81	-76.30	550000	2400	229	42
d'Iberville	80.56	-77.92	426000	3650	117	14
Cañon	79.68	-79.64	656000	4290	153	144
Sawyer Bay	79.36	-78.05	462000	3200	144	7
Parrish	79.57	-77.18	347000	2457	141	7
Eugenie	79.82	-74.93	402000	4747	85	61
Unnamed 4	80.07	-72.39	505000	3200	158	24
Manson Icefield						
Mittie West Arm	76.90	-79.53	1747000	6382	274	4
Mittie East Arm	76.87	-79.12	1990000	10247	194	4
Müller and Steacie Ice Caps						
Iceberg	79.43	-92.37	330000	4628	71	32
Good Friday Bay	78.55	-91.76	566000	5378	105	22
Sydkap Ice Cap						
Sydkap	76.62	-85.11	414000	2520	164	82

Table B.2 Sentinel-1 images used in this study.

Platform	Acquisition date	Polarisation	Orbit number	Pass direction	Slice number
Sentinel-1A	2016-02-09	HH	9875	Ascending	3-6
Sentinel-1A	2016-02-10	HH	9890	Ascending	4-9
Sentinel-1A	2016-02-11	HH	9905	Ascending	5-7
Sentinel-1A	2016-02-14	HH	9935	Ascending	3-4
Sentinel-1A	2016-02-16	HH	9972	Descending	1-4
Sentinel-1A	2016-02-17	HH	9986	Descending	1-5
Sentinel-1A	2016-02-20	HH	10030	Descending	1-3
Sentinel-1A	2016-02-22	HH	10065	Ascending	4-9
Sentinel-1A	2016-02-23	HH	10080	Ascending	5-7
Sentinel-1A	2016-02-25	HH	10109	Ascending	3-4
Sentinel-1A	2016-02-28	HH	10147	Descending	1-4
Sentinel-1A	2016-02-29	HH	10161	Descending	1-5
Sentinel-1A	2016-03-03	HH	10205	Descending	1-3
Sentinel-1A	2016-03-04	HH	10225	Ascending	3-6
Sentinel-1A	2016-03-05	HH	10240	Ascending	4-9
Sentinel-1A	2016-03-06	HH	10255	Ascending	5-7
Sentinel-1A	2016-03-08	HH	10284	Ascending	3-4
Sentinel-1A	2016-03-11	HH	10322	Descending	1-4
Sentinel-1A	2016-03-12	HH	10336	Descending	1-5
Sentinel-1A	2016-03-16	HH	10400	Ascending	3-6
Sentinel-1A	2016-03-17	HH	10415	Ascending	4-9
Sentinel-1A	2016-03-18	HH	10430	Ascending	5-7
Sentinel-1A	2016-03-23	HH	10497	Descending	1-4
Sentinel-1A	2016-03-24	HH	10511	Descending	1-5
Sentinel-1A	2016-03-27	HH	10555	Descending	1-3
Sentinel-1B	2017-01-28	HH	4054	Ascending	3-6
Sentinel-1B	2017-01-29	HH	4084	Ascending	5-7
Sentinel-1A	2017-01-29	HH	15047	Descending	1-4
Sentinel-1B	2017-01-30	HH	4069	Ascending	4-9
Sentinel-1A	2017-01-30	HH	15061	Descending	1-5
Sentinel-1B	2017-02-01	HH	4112	Ascending	7-9
Sentinel-1B	2017-02-01	HH	4113	Ascending	3-4
Sentinel-1A	2017-02-02	HH	15105	Descending	1-4
Sentinel-1A	2017-02-03	HH	15125	Ascending	3-6
Sentinel-1B	2017-02-04	HH	4151	Descending	1-4
Sentinel-1A	2017-02-04	HH	15140	Ascending	4-9
Sentinel-1B	2017-02-05	HH	4165	Descending	1-5
Sentinel-1A	2017-02-05	HH	15155	Ascending	5-7
Sentinel-1A	2017-02-07	HH	15183	Ascending	7-9
Sentinel-1A	2017-02-07	HH	15184	Ascending	3-4
Sentinel-1B	2017-02-08	HH	4209	Descending	1-4
Sentinel-1B	2017-02-09	HH	4229	Ascending	3-6
Sentinel-1A	2017-02-10	HH	15222	Descending	1-4
Sentinel-1B	2017-02-10	HH	4244	Ascending	4-9
Sentinel-1A	2017-02-11	HH	15236	Descending	5-9
Sentinel-1B	2017-02-11	HH	4259	Ascending	5-7
Sentinel-1B	2017-02-13	HH	4287	Ascending	7-9
Sentinel-1B	2017-02-13	HH	4288	Ascending	3-4
Sentinel-1A	2017-02-14	HH	15280	Descending	1-3
Sentinel-1A	2017-02-15	HH	15300	Ascending	3-6
Sentinel-1B	2017-02-16	HH	4326	Descending	1-4
Sentinel-1A	2017-02-16	HH	15315	Ascending	4-9
Sentinel-1B	2017-02-17	HH	4340	Descending	1-5
Sentinel-1A	2017-02-19	HH	15358	Ascending	7-9
Sentinel-1B	2017-02-20	HH	4384	Descending	1-4
Sentinel-1A	2017-02-22	HH	15397	Descending	1-3
Sentinel-1A	2017-02-23	HH	15411	Descending	1-5
Sentinel-1A	2017-02-26	HH	15455	Descending	1-4

Fig. B.3 Sydcap Ice Cap and Manson Icefield (Ellesmere Island) and Devon Ice Cap (Devon Island), Nunavut, Canadian Arctic.

Appendix C

Supplementary Materials. Chapter 5

Table C.1 Sentinel-1 images used in this study.

Platform	Acquisition date	Polarisation	Orbit number	Pass direction	Cycle number	Slice number
Sentinel-1B	2016-11-06	VV	2831	Descending	23	1
Sentinel-1B	2016-11-09	VV	2875	Descending	23	1
Sentinel-1B	2016-11-30	VV	3181	Descending	25	1
Sentinel-1B	2016-12-03	VV	3225	Descending	25	1
Sentinel-1B	2016-12-12	VV	3356	Descending	26	1
Sentinel-1B	2016-12-15	VV	3400	Descending	26	1
Sentinel-1B	2016-12-24	VV	3531	Descending	27	1
Sentinel-1B	2016-12-27	VV	3575	Descending	27	1
Sentinel-1B	2017-01-05	VV	3706	Descending	28	1
Sentinel-1B	2017-01-08	VV	3750	Descending	28	1
Sentinel-1B	2017-01-17	VV	3881	Descending	30	1
Sentinel-1B	2017-01-20	VV	3925	Descending	30	1
Sentinel-1B	2017-01-29	VV	4056	Descending	30	1
Sentinel-1B	2017-02-01	VV	4100	Descending	30	1
Sentinel-1B	2017-02-13	VV	4275	Descending	31	1
Sentinel-1B	2017-02-22	VV	4406	Descending	32	1
Sentinel-1B	2017-02-25	VV	4450	Descending	32	1
Sentinel-1B	2017-03-06	VV	4581	Descending	33	1
Sentinel-1B	2017-03-09	VV	4625	Descending	33	1
Sentinel-1B	2017-03-18	VV	4756	Descending	34	1
Sentinel-1B	2017-03-21	VV	4800	Descending	34	1
Sentinel-1B	2017-03-30	VV	4931	Descending	35	1
Sentinel-1B	2017-04-02	VV	4975	Descending	35	1
Sentinel-1B	2017-04-11	VV	5106	Descending	36	1
Sentinel-1B	2017-04-23	VV	5281	Descending	37	1
Sentinel-1B	2017-04-26	VV	5325	Descending	37	1
Sentinel-1B	2017-05-05	VV	5456	Descending	38	1
Sentinel-1B	2017-05-17	VV	5631	Descending	39	1
Sentinel-1B	2017-05-20	VV	5675	Descending	39	1
Sentinel-1B	2017-05-29	VV	5806	Descending	40	1
Sentinel-1B	2017-06-01	VV	5850	Descending	40	1
Sentinel-1B	2017-06-10	VV	5981	Descending	41	1
Sentinel-1B	2017-06-13	VV	6025	Descending	41	1
Sentinel-1B	2017-06-22	VV	6156	Descending	42	1
Sentinel-1B	2017-06-25	VV	6200	Descending	42	1
Sentinel-1B	2017-07-07	VV	6375	Descending	43	1
Sentinel-1B	2017-07-16	VV	6506	Descending	44	1
Sentinel-1B	2017-07-19	VV	6550	Descending	44	1
Sentinel-1B	2017-07-31	VV	6725	Descending	45	1
Sentinel-1B	2017-08-09	VV	6856	Descending	46	1
Sentinel-1B	2017-08-12	VV	6900	Descending	46	1
Sentinel-1B	2017-08-21	VV	7031	Descending	47	1
Sentinel-1B	2017-09-02	VV	7206	Descending	48	1
Sentinel-1B	2017-09-05	VV	7250	Descending	48	1
Sentinel-1B	2017-09-14	VV	7381	Descending	49	1
Sentinel-1B	2017-09-17	VV	7425	Descending	49	1
Sentinel-1B	2017-09-26	VV	7556	Descending	50	1
Sentinel-1B	2017-09-29	VV	7600	Descending	50	1
Sentinel-1B	2017-10-08	VV	7731	Descending	51	1
Sentinel-1B	2017-10-11	VV	7775	Descending	51	1
Sentinel-1B	2017-10-20	VV	7906	Descending	52	1
Sentinel-1B	2017-10-23	VV	7950	Descending	52	1
Sentinel-1B	2017-11-01	VV	8081	Descending	53	1
Sentinel-1B	2017-11-04	VV	8125	Descending	53	1
Sentinel-1B	2017-11-16	VV	8300	Descending	54	1

Table C.2 ICESat tracks used in this study.

Platform	Acquisition date	Track
ICESat	2003-03-05	20
ICESat	2003-03-21	35
ICESat	2003-03-28	85
ICESat	2003-03-02	100
ICESat	2005-03-02	100
ICESat	2004-03-01	115
ICESat	2005-03-05	154
ICESat	2004-03-04	169
ICESat	2004-03-07	219
ICESat	2004-03-08	234
ICESat	2004-03-11	273
ICESat	2004-03-12	288
ICESat	2004-03-16	338
ICESat	2004-03-17	353
ICESat	2005-03-21	392
ICESat	2004-03-20	407
ICESat	2004-02-21	1320
ICESat	2007-03-15	1335

Table C.3 WorldView DEM used in this study.

Object ID	Platform	Acquisition date	Function
52915	WorldView-1	2012-05-11	DEM differencing
54035	WorldView-1	2013-07-01	DEM differencing
58189	WorldView-2	2013-07-01	DEM differencing
56797	WorldView-2	2016-05-10	ICESat-DEM differencing and DEM differencing
59961	WorldView-3	2016-05-11	ICESat-DEM differencing and DEM differencing
56205	WorldView-2	2016-06-22	ICESat-DEM differencing
52027	WorldView-1	2016-07-13	ICESat-DEM differencing
59867	WorldView-3	2016-07-14	ICESat-DEM differencing and DEM differencing
59841	WorldView-3	2016-07-29	ICESat-DEM differencing
52092	WorldView-1	2016-07-30	ICESat-DEM differencing
59897	WorldView-3	2016-07-30	ICESat-DEM differencing

Fig. C.1 dh/dt WorldView-WorldView for Basin B

Fig. C.2 dh/dt WorldView-WorldView for Basin C

Fig. C.3 dh/dt WorldView-WorldView for Basin D

Fig. C.4 Monthly-averaged Arctic sea-ice concentration

Fig. C.5 End of September Arctic sea-ice concentration. NSIDC Sea Ice Polar Stereographic North projection.

Fig. C.5 End of September Arctic sea-ice concentration. NSIDC Sea Ice Polar Stereographic North projection.

Fig. C.6 Summer and year-averaged air temperatures 1987-2017 over Northern Komsomolets Island (from NCEP/NCAR Reanalysis 1 data)