UAV investigation into surface and tidewater mass loss processes of the Greenland Ice Sheet

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Summary

Accurately forecasting the contribution of the Greenland Ice Sheet to global sea-level requires precise observations to constrain present-day processes and incorporate them into models. However, the spatial and temporal resolution of satellite imagery and representativeness of in situ measurements often precludes or obscures our understanding of mass loss processes. This thesis investigates whether imagery from unmanned aerial vehicles (UAVs) have the potential to 1) bridge the scale gap between in situ and satellite observations and, 2) resolve processes of mass loss which are beyond the resolution of satellite imagery. It is found that the footprints of ground-based pyranometers are insufficient to capture the spatial heterogeneity of the ice surface as it progressively ablates and darkens. Point-to-pixel albedo comparisons are therefore often invalid, meaning that satellite-derived albedo measurements may be more accurate than previously thought. A 25 km transect intersecting the dark zone reveals that distributed impurities, not cryoconite nor surface water, govern spatial albedo patterns and may have implications for the future expansion of the dark zone. Repeat surveys over Store Glacier show that UAVs can be used to quantify calving rates and surface velocities of tidewater glaciers. The surveys indicate that large calving events cause short-term terminus velocity accelerations and can explain the seasonal pattern of acceleration and retreat. Any process which accelerates calving, such as removal of the ice mélange, therefore has important implications for the glaciers future behaviour.
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Chapter 1

Introduction and aims

1.1 Rationale

1.1.1 Processes of Greenland Ice Sheet mass loss

The contribution of the Greenland Ice Sheet to sea-level rise increased from less than 5% in 1993 to 25% in 2014 (Chen et al., 2017). The ice sheet mass balance is thought to have remained in near-equilibrium between 1960 and 2000, but, since 2000, it has been persistently negative with mass loss accelerating between 2006 and 2012 (van den Broeke et al., 2016). The rate of mass loss increased from 153 Gt a\(^{-1}\) in 2000 to 2005 to 265 Gt a\(^{-1}\) between 2005 and 2009 and 378 Gt a\(^{-1}\) between 2009 and 2012 (Enderlin et al., 2014). Mass loss rebounded to approximately 250 Gt a\(^{-1}\) between 2013 and 2015 but negative mass balance now seems to be the norm rather than the exception in the 21st century (van den Broeke et al., 2016). Mass loss is a result of surface meltwater runoff and ice discharge from tidewater glaciers (Rignot et al., 2008; van den Broeke et al., 2009; Enderlin et al., 2014; van den Broeke et al., 2016). The 2000 to 2005 mass loss was driven equally by the acceleration of many large tidewater glaciers and negative surface mass balance (Luckman et al., 2006; Howat et al., 2008; Joughin et al., 2010; Moon et al., 2012; Enderlin et al., 2014). But between 2006 and 2012, the discharge from tidewater glaciers remained constant and ice sheet mass loss became increasingly driven by negative surface mass balance (Enderlin et al., 2014). Surface mass balance was responsible for 68% of ice sheet mass loss between 2009 and 2012 (Enderlin et al., 2014).
The decrease in surface mass balance is driven by increased meltwater runoff in the ablation area, rather than a reduction in snowfall (van den Broeke et al., 2016). Surface albedo modulates the absorption of incoming shortwave radiation and is therefore the primary factor governing the surface energy balance of the ice sheet and determining the amount of meltwater runoff (Braithwaite, 1995; Knap and Oerlemans, 1996; van den Broeke et al., 2011). This is exemplified by the positive feedback between albedo and melt (Box et al., 2012). As snow melts, it darkens, increasing the absorption of shortwave radiation resulting in further melting (Box et al., 2012). Furthermore, bare ice, which has a much lower albedo than snow, is exposed once the snow has melted (Tedesco et al., 2011; Box et al., 2012). The longer bare ice is exposed, the more the Greenland Ice Sheet melts (Tedesco et al., 2011; Box et al., 2012). These feedbacks make the ice sheet very sensitive to the duration and extent of bare ice exposure but also to the albedo of the bare ice (Fettweis et al., 2011; Box et al., 2012; Tedesco et al., 2016).

Recently, the expansion of the bare ice area, particularly in southwestern Greenland (Tedesco et al., 2011; Box et al., 2012), and ongoing decline in bare ice albedo (Shimada et al., 2016; Tedstone et al., 2017) has motivated a new generation of surface energy balance models that assimilate spatial patterns of albedo derived from satellite data (van Angelen et al., 2012; van den Broeke et al., 2016). However, the drivers of bare ice darkening are yet to fully resolved, with some studies suggesting it is derived from melt-out of englacial dust, and others indicating that it has a biological origin (Wientjes and Oerlemans, 2010; Wientjes et al., 2012; Shimada et al., 2016; Tedstone et al., 2017). Recent research that promotes the concept of bioalbedo, suggests that the melt-out of particulates combine with available water to fertilise pigmented ice algae, which subsequently drive enhanced albedo reduction over the melt-season (Tedstone et al., 2017). However, the spatial distribution and concentration of surface algae, including its interaction with the availability of nutrients, changing meteorological conditions and ice sheet ablation and runoff, remains poorly understood.

Although the contribution of tidewater glacier discharge to mass loss decreased from 58% between 2000 and 2005 to 32% between 2009 and 2012 (Enderlin et al., 2014), understanding changes in ice flow from individual outlet glaciers is important because ice discharge appears to respond non-linearly to environmental forcing (Joughin et al., 2012; Straneo and Heimbach, 2013). This makes tidewater glaciers one of the most poorly constrained contributors to sea-
level rise in the 21st century (IPCC, 2013). Furthermore, in Antarctica, almost all mass is lost from tidewater margins (Depoorter et al., 2013). Tidewater glacier flow dynamics are thought to be sensitive to atmospheric and oceanographic forcing (Holland et al., 2008; Joughin et al., 2012; Straneo and Heimbach, 2013). But whilst, glaciers in southeast Greenland accelerated synchronously between 2002 and 2005 (Luckman et al., 2006; Joughin et al., 2010; Moon et al., 2012), the event was short-lived and neighbouring glaciers often exhibit contrasting retreat and acceleration which indicate that the interaction between tidewater glaciers and environmental conditions may be complex (Moon and Joughin, 2008; Straneo et al., 2012; Enderlin et al., 2014; Moon et al., 2012).

The coupling between retreat and ice flow acceleration demonstrates that tidewater glaciers are sensitive to resistive stress at the calving front (Thomas, 2004; Vieli and Nick, 2011; Joughin et al., 2008). Changes in calving front position are governed by calving rates which vary in style between each glacier (Amundson et al., 2008; James et al., 2014; Medrzycka et al., 2016). Calving processes, and their relationship with ice dynamics, are poorly understood because they are complex and difficult to observe. Calving includes a range of fracture processes which range in spatial and temporal scales. These processes range from the production of kilometer long tabular icebergs to the collapse of seracs (Medrzycka et al., 2016). Calving styles vary depending on the geometry of calving front and fjord in which the glacier terminates but also depending on the time of year and whether or not an ice mélange, a rigid mixture of icebergs and sea-ice floating, exists in front of the glacier terminus (Medrzycka et al., 2016). Field data collection from calving fronts is difficult and dangerous since these processes occur at the ice-ocean boundary. Calving rates may be derived from satellite remote sensing but this is only possible when repeat imagery is obtained at high enough temporal resolution (Sundal et al., 2013; Luckman et al., 2015). These data sets do not exist for most Greenland tidewater glaciers and so our understanding of the interplay between calving, ice flow velocity and terminus positions remains poorly understood.

1.1.2 Unmanned aerial vehicles in glaciology

Recent advances in microcontrollers, open-source autopilot software and Structure-from-Motion (SfM) software have enabled glaciologists to obtain high-resolution georeferenced orthomosaics
and digital elevation models (DEMs) with unmanned aerial vehicles (UAVs) equipped with consumer-grade digital cameras (d’Oleire Oltmanns et al., 2012; Westoby et al., 2012; Hugenholtz et al., 2013; Lucieer et al., 2014; Tonkin et al., 2014). UAVs offer flexibility to acquire customized aerial data in remote, hazardous and/or inaccessible regions. Depending on the required spatial resolution and application, flying height can be prescribed. Likewise, specific study sites can be repeatably surveyed at sub-daily intervals. Although digital cameras have tended to be the most common payload for recent glaciological applications, the miniaturization of laser altimeters, multispectral and hyperspectral sensors means that aerial data collected by UAVs may be further customized in the future.

Weaknesses of UAVs include relatively small ranges and limited payload capacity in comparison to manned aircraft, which can also provide aerial data with similar flexibility. UAVs tend to be battery powered, and even the most advanced batteries available on the market have around 100 times less energy density than petroleum-derived liquid fuels such as petrol. This means that UAVs are limited to a maximum of a couple of hours of surveying and may only be able to fly a few tens of kilometres. In contrast, NASA’s Douglas DC-8-72 can fly for up to 12 hours and has a range of nearly 10,000 kilometres. Likewise, the DC-8-72 can carry up to 14,000 kg of scientific payload whereas UAVs are limited to a couple of kilograms. However, UAVs can acquire data at a substantially lower cost than that of manned aircraft surveys. Furthermore, UAV pilots do not need extensive training and the UAVs themselves are not subject to the same safety regulations and maintenance required by manned aircraft. For these reasons, UAVs are often favoured over manned aircraft and UAV-based applications have become more and more common in the past decade.

The high spatial resolution of UAV data may lead to new insights into processes that occur on the surface of glaciers and ice sheets if the features are too small to be resolved by satellite imagery. For example, UAV imagery and DEMs with a pixel resolution of 5 and 10 cm, respectively, allowed Rippin et al. (2015) to quantify surface roughness and patterns of surface hydrology across the lower reaches of Midtre Lovénbreen, Svalbard. The scale of streams and roughness features (cm to m) precludes their observation from traditional satellite imagery, such as Landsat 8 with a pixel size of 15 m. Similarly, Hodson et al. (2007) quantified the distribution and area of cryoconite holes, also on Midtre Lovénbreen, which typically have areas ranging between a
few centimetres and a couple of metres. Over Arctic Ocean sea-ice near Barrow, Alaska, (Inoue et al., 2008) were able to determine the area of melt-ponds using 8 cm imagery. Although these features are small, they may have important implications for large-scale cryospheric processes such as sea-ice break-up and ice surface albedo feedbacks.

DEM acquired by UAVs using SfM are useful for monitoring glacier health, especially since the effects of warming air temperatures often manifest themselves in glacier surface elevation change. Whitehead et al. (2013) took advantage of UAVs to measure the spatial variability and interannual rates of glacier thinning or thickening at Fountain Glacier in the Canadian Arctic. Immerzeel et al. (2014) used a similar technique to study the evolution of ice cliffs and supraglacial lakes on Lirung Glacier in the Himalayas. The short time intervals between UAV surveys may also reveal insights into glaciological processes which occur over short timescales. Jouvet et al. (2017) used weekly UAV derived DEMs to track the propagation of fractures at the terminus of Bowdoin Glacier in Greenland. But, at the time of writing, further examples multi-day UAV surveys to investigate glaciological changes have yet to be published.

The increasing use of consumer-grade cameras for collecting aerial data has resulted in the development of techniques for quantifying the reflectivity of surfaces from digital imagery (Bhardwaj et al., 2016). Di Mauro et al. (2015) used raw digital camera imagery obtained from a multi-rotor UAV to measure the impact of mineral dust deposition on snow radiative properties in the European Alps. The visible band UAV imagery correlated strongly with coincident Analytical Spectral Devices (ASD) handheld spectroradiometer measurements \( R^2 \) of 0.83 and was used to upscale the point measurements of mineral dust content to the satellite scale (Di Mauro et al., 2015). Comparison between UAV and satellite imagery (and multi-temporal UAV surveys) requires radiometric correction of the UAV imagery since images are often acquired under variable atmospheric conditions, solar illumination and viewing angles. Radiometric correction removes the effects of variable illumination geometry and atmospheric transmittance to derive surface reflectance, which is defined as the fraction of incident radiation that is reflected by a surface for a single incidence angle (Schaepman-Strub et al., 2006). Burkhart et al. (2017) derived surface reflectance of snow from a 210 km fixed-wing UAV transect across the interior of the Greenland Ice Sheet for comparison and validation of satellite surface reflectance products (Burkhart et al., 2017). Albedo, on the other hand, is the directional integration of reflectance
over all solar and viewing angles (Schepman-Strub et al., 2006). If the surface is assumed to
reflect isotropically (i.e. equally at all angles) albedo may be measured by a single angular
measurement of reflectance. However, ice and snow are thought to reflect anisotropically which
means a bidirectional reflectance distribution function (BRDF) should be taken into consider-
ation. Hakala et al. (2010) derived a BRDF over snow using multi-angular surface reflectance
measurements from a digital camera mounted on a multi-rotor UAV.

The advantages of UAVs over ground-based measurements and satellites indicate that they may
be useful for investigating processes of Greenland Ice Sheet mass loss at both the ice-ocean and
ice-atmosphere boundaries. A number of processes at these boundaries such as the drivers of
bare ice darkening and calving patterns from tidewater glaciers are yet to fully resolved by the
current suite of observational technologies. Three applications where UAV-based approaches
may be useful are identified in this thesis. The first application is to exploit the fine spatial
resolution imagery. A number of features of the Greenland Ice Sheet such as meltwater streams
and cryoconite holes have areas ranging between a few centimetres and a couple of metres and
are beyond the spatial resolution of most satellite imagery (e.g. Landsat 8 with a pixel size of
15 m). UAV imagery may therefore be useful for quantifying the prevalence of these features
and investigating their contribution to the average reflectance of coarse satellite pixels. The
second application is to exploit the high temporal resolution enabled by repeat UAV surveys.
For example, the calving of large icebergs typically occurs over timescales of minutes to hours,
at temporal resolutions too small for satellite remote sensing (Landsat 7 and 8 have a repeat
coverage of eight days when combined). Calving events are also difficult to measure with ground-
based measurements because of the inaccessibility of tidewater glacier calving fronts. Repeat
daily or even sub-daily UAV surveys may therefore be useful for investigating the dynamics
occurring in these hazardous and dynamic environments. The third application is to bridge
the spatial gap between ground-based measurements and satellite imagery. For example, a
handheld spectroradiometer measurement of albedo over a 50 x 50 cm area of bare ice may
contain a lot of spectral information but may not be spatially representative over larger areas
of the ice sheet. Fine spatial resolution UAV imagery may therefore be useful for identifying
correlations between the handheld albedo measurement and RGB digital numbers and upscaling
the results to areas in the order of tens of kilometres squared.
1.2 Thesis structure and aims

The thesis is structured as a series of five integrated papers, two methods and three applications. The papers are separated and ordered into two broad themes, with each method followed by the application(s). This means that papers were written in a different order to that presented here. The first theme, Chapters 2, 3 and 4, addresses processes of mass loss from the surface of the ice sheet, at the ice-air boundary. The UAV imagery is used to generate albedo maps with high spatial resolution. These maps enable the quantification of the ablation zones’s spatial heterogeneity and characterize the surface types that dominate the dark zone, a region of the ice sheet with the lowest albedo. The second theme, Chapters 5 and 6, investigates processes of mass loss at the ice-ocean boundary. Here, repeat UAV surveys are used to produce a dense times series of calving rates, surface velocities and mélange thickness at Store Glacier between May and July which enable investigation into the seasonal dynamics of the glacier.

The first aim of the thesis is to establish whether UAVs are useful for investigating processes of ice sheet mass loss. Once this is established, in Chapters 2 and 5, the second aim of the thesis is to demonstrate the usefulness of UAVs with three applications in Chapters 3, 4 and 6. Chapter 3 bridges the spatial gap between in situ and satellite measurements; Chapter 4 demonstrates the importance of fine spatial resolution imagery; and Chapter 6 demonstrates the importance of fine temporal resolution imagery. Each paper is summarized at the beginning of each chapter in order to give an overview of the chapter as a whole and its context in the thesis. Chapters 2, 3 and 5 are published in Frontiers in Earth Science, Geophysical Research Letters and The Cryosphere, respectively, and are formatted according to the journals’ typesetting standards. Chapter 4 is in review with Nature Communications and Chapter 6 is yet to be submitted. Both are formatted for submission.
Chapter 2

Derivation of high spatial resolution albedo from UAV digital imagery: application over the Greenland Ice Sheet

2.1 Summary

Chapter 1 describes a method to derive a high spatial resolution albedo product using a consumer-grade digital camera and two broadband pyranometers mounted on a UAV. In doing so, the study addresses the first aim of the thesis which is to establish whether UAVs are useful for investigating processes of mass loss at the ice-atmosphere boundary. The study finds that visible band imagery obtained at nadir reliably measures albedo, even though it does not capture the full shortwave spectrum or account for the anisotropy of bare ice. The method is demonstrated using UAV imagery collected across the Kangerlussuaq Sector of the Greenland Ice Sheet in July 2015. The three resultant albedo products cover a total area of 280 km$^2$ with a pixel size of 20 cm and demonstrate that the ablation zone displays large spatial heterogeneity in albedo. We show that this is because the surface consists of numerous non-ice constituents and surface structures which include structural features such as crevasses, fractures and folia-
tions; surface water; snow patches; cryoconite that is concentrated in holes or in fluvial deposits; microbes and their humic by-products; and mineral dust and aerosols.

We conclude that the method outlined here may have two primary applications. The first is to assess the representativeness of albedo parameterizations in surface mass balance models, which are currently prescribed using coarse resolution (500 x 500 m pixel size) satellite products. The second is to assess the length scales at which bare ice surface albedo should be sampled to capture its spatial heterogeneity. Calibration and validation of satellite albedo products rely upon point-based in-situ measurements but direct comparison between the two is only valid when the footprint of the in-situ measurement is either the same as the corresponding satellite image pixel or when the surface is sufficiently homogeneous at the scale of both the in-situ measurement and satellite data pixel. The high spatial resolution albedo products may therefore be used to investigate the robustness of in-situ albedo observations for validating satellite-derived albedo. We address this application in Chapter 3.

2.2 Contribution

To complete this study, JR designed, tested and built the UAVs used for data acquisition. JR piloted the UAV during the summer of 2015 and obtained imagery over a total area of 280 km². JR developed and validated the method to convert UAV imagery to an albedo product processed all the imagery into three albedo products. JR wrote the first draft of the manuscript and edited it in response to suggestions from co-authors and three reviewers. JR made all the figures.
Derivation of High Spatial Resolution Albedo from UAV Digital Imagery: Application over the Greenland Ice Sheet

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Measurements of albedo are a prerequisite for modeling surface melt across the Earth’s cryosphere, yet available satellite products are limited in spatial and/or temporal resolution. Here, we present a practical methodology to obtain centimeter resolution albedo products with accuracies of ±5% using consumer-grade digital camera and unmanned aerial vehicle (UAV) technologies. Our method comprises a workflow for processing, correcting and calibrating raw digital images using a white reference target, and upward and downward shortwave radiation measurements from broadband silicon pyranometers. We demonstrate the method with a set of UAV sorties over the western, K-sector of the Greenland Ice Sheet. The resulting albedo product, UAV10A1, covers 280 km², at a resolution of 20 cm per pixel and has a root-mean-square difference of 3.7% compared to MOD10A1 and 4.9% compared to ground-based broadband pyranometer measurements. By continuously measuring downward solar irradiance, the technique overcomes previous limitations due to variable illumination conditions during and between surveys over glaciated terrain. The current miniaturization of multispectral sensors and incorporation of upward facing radiation sensors on UAV packages means that this technique could become increasingly common in field studies and used for a wide range of applications. These include the mapping of debris, dust, cryoconite and bioalbedo, and directly constraining surface energy balance models.

Keywords: albedo, digital camera, unmanned aerial vehicle, remote sensing, Greenland Ice Sheet, energy balance modeling
1. INTRODUCTION

The seasonal and interannual variability of melt on glaciers and ice sheets is primarily driven by the absorption of shortwave radiation at the surface (Braithwaite and Olesen, 1990; Munro, 1990; Arnold et al., 1996; Brock et al., 2000; van den Broeke et al., 2008). Albedo modulates the fraction of absorbed shortwave radiation and is known to be highly variable in space and time (e.g., Cutler and Munro, 1996; Jonsell et al., 2003; Klok et al., 2003). Quantifying albedo accurately and with high spatiotemporal resolution, is therefore critical for accurately determining the surface mass balance of glaciers and ice sheets (Gardner and Sharp, 2010). Despite this, patterns and dynamics of surface albedo remain one of the most prominent uncertainties in energy balance modeling (Hock, 2005; Noël et al., 2015).

Traditionally, energy balance models have relied on in-situ measurements, satellite imagery or numerical modeling for boundary condition albedo fields (e.g., Reijmer et al., 1999; van de Wal et al., 2005; Gardner and Sharp, 2010; van den Broeke et al., 2011; van As et al., 2012; Fettweis et al., 2013; Noël et al., 2015). However, the limitations of these methods, such as unrepresentative point measurements, coarse satellite pixel resolution, and incomplete representation of bare ice albedo in models, have motivated the development of new observation techniques. One such technique involves retrieving albedo from digital camera imagery (e.g., Corripio, 2004; Rivera et al., 2008; Dumont et al., 2011; Garvelmann et al., 2013; Rippin et al., 2015; Ayala et al., 2016). But while these approaches appear promising they range in complexity and suffer from large errors. For example, the relatively simple technique described by Rippin et al. (2015) for estimating an “albedo proxy” from the pixel values of RGB orthomosaics, likely suffers from high errors because, amongst other things, it does not account for variable illumination conditions. At the other extreme, Dumont et al. (2011) presented a theoretically rigorous approach for calculating albedo from oblique digital image pixels using a radiometrically calibrated camera, a bidirectional reflectance distribution function (BRDF) of snow and ice, a parametric transmittance model for downward spectral radiation, and a radiative transfer model for broadband to broadband conversions. Although comprehensive, their methodology fails to retrieve robust albedo measurements under clouds and relies on assumptions about atmospheric transmittance and reflectance anisotropy of ice and snow, both of which are difficult to model or measure.

Accordingly, the primary aim of this study is to describe a technique for determining the albedo of snow and ice surfaces at high spatiotemporal resolution (i.e., centimeter to decimeter pixel resolution) and accuracy (i.e., ±5% or better) using a consumer-grade digital camera mounted on an unmanned aerial vehicle (UAV). The technique aims to be reproducible and practical under rapidly changing illumination conditions such as those commonly experienced in glaciated regions. The application of this method is demonstrated using digital images acquired by a UAV over the Kangerlussuaq sector of the Greenland Ice Sheet between June and July 2015. The accuracy of the albedo product is validated with near-simultaneous ground-based albedo measurements and the MOD10A1 satellite albedo product.

2. METHODS

2.1. Assumptions

Estimating albedo from digital image pixels requires assumptions regarding the spectral reflectance of ice and snow and its relationship with the digital numbers of the image acquired by the camera. Our technique implicitly assumes that visible band (400–700 nm) digital imagery accurately captures the reflectance variability of surface types commonly found in glacier ablation areas. We evaluate this assumption by comparing albedo derived from the digital images to that obtained by a pair of Kipp & Zonen CM3 optical pyranometers (Section 3.3). We note that, at least for bare ice surfaces and snow with uniform grain size, most of the variability in reflected radiation lies in the visible band of the shortwave spectrum, between 350 and 695 nm (Wiscombe and Warren, 1980; Dozier et al., 1981; Cutler and Munro, 1996; Corripio, 2004) where a standard consumer-grade digital camera is most sensitive (e.g., Figure 1) (Jiang et al., 2013). In contrast, bare ice and snow albedo with uniform grain size does not display high variability in the near-infrared wavelengths (695–2,800 nm) (Wiscombe and Warren, 1980; Cutler and Munro, 1996).

We also assume there is a direct relationship between the digital numbers of the camera image, which represent brightness, and surface reflectance. We argue this assumption can be justified if the camera is suitably calibrated and corrected using the methodology presented in this study. Furthermore, previous studies have demonstrated that it is possible to retrieve albedo from digital images with accuracies of ±10% or less (e.g., Corripio, 2004; Rivera et al., 2008; Dumont et al., 2011; Garvelmann et al., 2013; Ayala et al., 2016). Based on these assumptions, a procedure (Figure 2) to obtain and process imagery suitable for conversion to albedo is described in Sections 2.2–2.6.

![Figure 1](adapted from Jiang et al., 2013)
2.2. Procedure Summary

Here, an example instrument configuration is outlined for obtaining albedo data using digital camera imagery and broadband pyranometers mounted on a UAV. Instruments include a standard consumer-grade digital camera, a pair of broadband silicon pyranometers and a white reference target (specific instruments used in our application are described in Section 3). Digital images of the white Teflon reference target, assumed to be a near-Lambertian reflector, were used to convert downward radiation measured by the upward facing pyranometer into digital numbers. Because the reflected radiation measured by the downward facing pyranometer is used to correct the digital images, both sensors were mounted in close proximity to each other in order to minimize the differences between their respective ground footprints. The instruments may be mounted on a multi-rotor UAV, kite, blimp (e.g., Carrivick et al., 2013), fiberglass rod (e.g., van der Hage, 1992), step-ladder or other handheld device depending on the required resolution and spatial extent of the investigation.

2.3. Broadband Pyranometers

Upward and downward hemispherical reflectance were measured using broadband pyranometers, the ratio of which provides an estimate of the surface albedo. Downward radiation may be measured at a fixed ground station but if illumination conditions are rapidly changing, perhaps due to cloud cover, it may be more accurate to measure downward radiation from the aerial platform. In this case, we recommend using silicon pyranometers due to their low weight (~100 g per sensor) and rapid response time of 1 ms. The pyranometers must be precisely leveled with a clear view of the sky or surface to ensure accurate measurements.

The disadvantage of silicon pyranometers is that they are only sensitive to wavelengths between 300 and 1,100 nm, which excludes some of the solar energy contributing to the surface energy balance. We evaluate this source of uncertainty by comparing albedo derived from a pair of Apogee silicon pyranometers (https://www.campbellsci.com.au/sp-110) with albedo obtained by a pair of Kipp & Zonen CM3 optical pyranometers. The Kipp & Zonen pyranometers are based on carbon black thermopiles which have a flat spectral response.
for wavelengths between 300 and 2,800 nm and an accuracy of 2%. Comparison between the two sensors was made by simultaneously sampling 25 surfaces with an albedo of between 0.11 and 0.64 in a 20 × 20 m study area. The instruments were mounted on a 1 m aluminum rod and held 0.5 m above each surface for 1 min. Albedo was calculated as the mean of five 20 s samples. At this height, the pyranometers have a ground footprint of around 5.5 m in diameter. A root-mean-square difference (RMSD) of 3.6% between the two instruments justified that silicon pyranometers can accurately measure albedo of ice sheet surfaces between 0.11 and 0.64.

The Apogee silicon pyranometers were factory-adjusted to account for their partial spectral sensitivity under standard clear-sky atmospheric conditions. Therefore the upward facing pyranometer has an uncertainty of 5%, which includes a cosine-response error of 2% at solar zenith angles of 45°. The cosine-response error increases as solar zenith angle increases so we recommend acquiring measurements as close as possible to solar noon.

2.4. Digital Images

The digital images were acquired in RAW format with fixed exposure, ISO, and aperture. These settings are necessary to preserve linearity between RGB digital numbers acquired by the camera and the number of photons hitting the sensor. We recommend a relatively fast shutter speed to minimize image blur (when mounted on a moving UAV) and a low ISO and F-number to ensure that pixels do not saturate. Most software cannot read the RAW images so they should be converted to lossless 16-bit TIFF images using RAW decoding software such as dcraw (http://cybercom.net/~dcoffin/dcraw/) (Figure 2). When converting the RAW images it is important to use a fixed white level, or balance, and avoid using a gamma correction to ensure the reflectance histogram is not altered and the RGB digital numbers remain directly proportional to the brightness of the surface (Lebourgeois et al., 2008). A gamma correction is a non-linear function used to convert between the brightness measured by the camera’s sensor and the output image pixel. In this study, we were only interested in the ratio between the total reflected and downward shortwave radiation and so the arithmetic mean of each pixel’s RGB values was calculated.

Image brightness attenuation away from the center, vignetting, is inherent to all consumer-grade digital imagery and can be corrected by subtracting a vignette mask from the original image (Goldman, 2010). The vignette mask can be calculated by fitting a third-order polynomial function to multiple Gaussian smoothed images (e.g., Lebourgeois et al., 2008). The mean vignette polynomial function can then be applied to correct all the images from the survey. Finally, the geometric distortion of the images should be corrected to ensure that each image pixel represents the same ground sampling distance. We used Agisoft Lens 0.4.0 software and an on-screen checker board to automatically determine the camera calibration parameters and correct for this distortion. After vignette and geometric correction, the digital numbers of every pixel in the image are directly comparable.

2.5. Digital Image Illumination Correction

The radiation recorded by an image pixel is not only related to the reflectance of the surface, but also to the downward radiation. The digital numbers in an image can be corrected for variations in illumination by including a leveled white Teflon reference target in every digital image (e.g., Hakala et al., 2010). An illumination corrected image can be produced from the ratio between the digital numbers of the ice surface pixels and the digital numbers of the white reference target, which are a proxy for downward solar irradiance. After applying an illumination correction, the digital numbers of pixels are comparable between images including those obtained under very different weather or illumination conditions.

If survey area is large, it may not be possible to include a white reference target in every image. Therefore, to facilitate the correction of UAV acquired imagery there is a need to quantify the relationship between downward radiation measured by an upward facing pyranometer and the RAW digital numbers of a white reference target. If linearity between the reflectance of the surface and digital numbers of the RAW images is preserved, then the relationship can be statistically represented by an orthogonal distance linear regression.

An example of this relationship between the digital numbers of a white reference target obtained by a Sony NEX-5N camera in RAW mode from the ground against the downward radiation measured by a static, upward facing silicon pyranometer with a clear sky view is presented (Figure 3). Here, the orthogonal distance regression fits the data well with a $R^2$ of 0.96 and an RMSD of 7.8%. The slope and intercept of the linear regression can subsequently be used to convert the downward radiation measured by the pyranometer to digital numbers. An illumination corrected image is then produced from the ratio between pyranometer to digital numbers.

![Figure 3](image.png)  
**Figure 3** | The relationship between the downward radiation measured by a broadband pyranometer and the digital numbers of a white reference target.
2.6. Digital Image Albedo Calibration

The final stage in the workflow is to calibrate the image using coincident measurements of albedo obtained from upward and downward facing pyranometer measurements. Snow and ice reflect incoming radiation anisotropically such that a nadir measurement of reflectance actually underestimates albedo by between 1 and 5% between 530 and 610 nm (i.e., the visible band) (Knap and Reijmer, 1998; Klok et al., 2003). The pyranometers account for the reflectance anisotropy of snow and ice because they measure hemispherical reflectance so can be used to correct the bidirectional reflectance measurements of the digital camera. This was achieved by multiplying the image pixel numbers by a factor calculated by dividing the mean pixel value of the illumination-corrected image by the albedo recorded by the pyranometers.

This step introduces uncertainty since the rectangular footprint of the digital image does not completely coincide with the circular footprint of the downward facing pyranometer (Figure 4). We overcome this issue by comparing the mean image pixel within the circular pyranometer footprint to the mean image pixel outside the pyranometer footprint for 300 representative sample images obtained from 600 m above the surface of the ablation area of the Greenland Ice Sheet. The silicon pyranometer has a circular FOV of approximately 60° whereas the Sony NEX-5N camera, used to test this uncertainty, has a rectangular FOV of 53.1 by 73.7°. The significant, strong positive correlation ($R^2 = 0.98; p < 0.01$) between the mean pixel value inside and outside the pyranometer footprint, with an RMSD of 1.2%, demonstrates that the downward facing pyranometer can be used to correct the digital images over length scales in the order of hundreds of meters (Figure 4). We note that this relationship may not hold when there is a spatial trend at the scale of the image or extreme heterogeneity in the surface being measured.

3. CASE STUDY: KANGERLUSSUAQ SECTOR OF THE GREENLAND ICE SHEET

Our methodology is demonstrated using 7,378 digital camera images acquired from a fixed-wing UAV flow over the K-sector of the Greenland Ice Sheet between June and July 2015. We refer to this albedo product as “UAV10A1.” UAV10A1 covers Isunguata Sermia, herein called UAV10A1-Isunguata, and two regions of the ice sheet referred to as UAV10A1-S6 and UAV10A1-Behar (Figure 8). UAV10A1-S6 and UAV10A1-Behar are so-called because of their proximity to the weather station at Site 6 on the K-transect (van de Wal et al., 2005) and a large supraglacial river informally named Rio Behar described in Smith et al. (2015), respectively.

3.1. Vehicle and Instrument Set-up

The fixed-wing UAV used in this study was described by Ryan et al. (2015) and based on a Skywalker X8 airframe constructed from expanded polypropylene foam with a wingspan of 2.12 m (Figure 5). Autonomous control was provided by a Pixhawk autopilot module which utilizes an L1 GPS, two inertial measurement units (IMUs), a compass, and a barometer. A 30 Ah 14.4V lithium-ion battery pack provides power for the 715W electromagnetic motor, two servos, the receiver and the autopilot module. The UAV cruising speed is regulated by
a digital differential airspeed sensor and targets 54 km h\(^{-1}\). The weight of the UAV without the sensor package was 4.79 kg. The sensor package weighs 0.715 kg and includes a Sony NEX-5N digital camera (0.345 kg), a Hobo micro data logger stripped of its case (0.075 kg) and powered externally, two Hobo pyranometers (0.170 kg), a temperature/humidity sensor (0.065 kg) and a GoPro HERO4 (0.060 kg). In this configuration, the UAV had an all-up-weight 5.5 kg and a 140 km range dependent on conditions.

The upward and downward facing Hobo silicon pyranometers were aligned within the airframe (Figure 5) so that they were level when the UAV was in stable flight. Errors due to instrument tilting were limited by omitting radiation data recorded when the pitch or roll of the UAV exceeded 3\(^\circ\). According to Bogren et al. (2016), this introduces a maximum bias of 8.1%. At a flight altitude of 600 m above the surface, a 60° FOV gives the pyranometer an estimated 700 m diameter footprint. The pyranometers were sampled by a micro data logger at 1 Hz and, before each flight, the data logger was synchronized to GPS time to ensure that upward and downward shortwave radiation data coincided with GPS and attitude data recorded by the PixHawk in post-processing.

Digital images were obtained by a downward facing Sony NEX-5N digital camera mounted in the UAV airframe (Figure 5). The camera was pre-set for a fixed exposure of 1/1,000 s, ISO 100 and F-stop of 8. The camera has a 16 mm fixed-focus lens (equivalent to 24 mm for a 35 mm film camera) which gives an FOV of 53.1 by 73.7\(^\circ\) and yields a ground footprint of approximately 900 \(\times\) 600 m at an altitude of 600 m above the ice surface. Vignetting was as high as 17.6% at the corners of the images and was corrected with a vignette correction mask (Section 2.4). Each image was tagged with the geographic location and attitude data recorded by Pixhawk's two IMUs and L1 GPS. The camera was triggered every 50 m of horizontal displacement (about every 3 s) which provided a forward overlap of 92%.

Gridded flight patterns over the ice sheet were planned and uploaded to the UAV using the Mission Planner ground station software (http://ardupilot.org/planner/). The UAV was programmed to maintain a constant altitude above sea level during missions and was targeted at 600 m above the surface. For the purpose of our study, this altitude was deemed an optimal trade-off between ground sampling distance and coverage. The distance between flight lines was set to 300 m which provided a 50% horizontal or side overlap between the aerial images.

### 3.2. Orthomosaic Production

The images were corrected and calibrated using the workflow described in Section 2 and mosaicked using Agisoft PhotoScan 1.3.0, herein PhotoScan, to produce UAV10A1. The images were imported into PhotoScan and automatically aligned based on the georeferencing information. This approach avoids the need for GPS-surveyed ground control points and reduces processing time because the software pre-aligns the images instead of performing global point matching to achieve accurate image alignment. The orthomosaics were produced in the software's "mosaic" mode, meaning that the pixels in the centers of the image were preferentially used to provide the output pixel value.

The orthomosaics were nearest-neighbor resampled to a pixel resolution of 20 cm and exported as GeoTIFFs for analysis.

### 3.3. Validation and Comparison

Validation of UAV10A1 was undertaken by comparison with surfaces sampled by the Kipp & Zonen CM3 pyranometers (Section 2.3). Nine of the twenty five surfaces were identified in three images acquired when the UAV was close to the ground and a comparison, made by sampling pixels within a 5.5 m diameter circle in the images, indicates that the two measurements compared well with an RMSE of 4.9% and a bias of 4.1% (Figure 6). The slight underestimation of CM3 in comparison to UAV10A1 may be partly explained by the obscuring of reflected radiation by the observer holding the CM3 on the meter long rod (e.g., van der Hage, 1992). The difference between the range of wavelengths measured by the CM3 pyranometers (300–2,800 nm) and digital camera (300–700 nm) may be a source of error. For example, changes in snow grain size have less effect on visible wavelength albedo than in the near-infrared wavelengths (Wiscombe and Warren, 1980). The visible band digital camera is not sensitive to albedo variation in the near-infrared which likely introduces error when compared to CM3 albedo. Generally, UAV10A1 measures albedo of the ice sheet surface adequately between 0.11 and 0.64, and within our target accuracy of ±5%.

UAV10A1 was also compared to the satellite-derived MOD10A1 Collection 6 daily albedo product (300–3,000 nm) obtained from the National Snow and Ice Data Center (NSIDC) (Hall and Riggs, 2016). The daily temporal resolution of MOD10A1 reduces uncertainties that may arise due to changes in ice surface albedo between data acquisition. MOD10A1. The value of each MOD10A1 pixel represents the best single albedo observation in the day, based on cloud cover and viewing
and illumination angles (Stroeve et al., 2006). Retrieving the
time of the MOD10A1 albedo observation was not possible
and we note that there may be up to a couple of hours
difference between UAV and satellite data acquisition which may
lead to discrepancies between the two data sources if there is a
significant diurnal albedo cycle caused by changing solar azimuth
and zenith angles or ice and snow metamorphosis during the day
(Stroeve et al., 2006; Bogren et al., 2016). Comparison between
both products was achieved by mean resampling UAV10A1 to
the same pixel resolution as the MOD10A1 product (∼463 m)
using the average pixel value of UAV10A1. Both products were
then clipped to the same extent and 347 cloud-free samples were
compared. The mean RMSD between the two albedo products
was 3.7% which is within the 6.7–7.0% estimated uncertainty of
the MOD10A1 product (Figure 7) (Stroeve et al., 2006; Wang
et al., 2011; Ryan et al., 2016).

3.4. High-Resolution Albedo Product
3.4.1. UAV10A1-Isunguata
UAV10A1-Isunguata is a 130 km² map that covers the
tongue of the land-terminating glacier in West Greenland
approximately 26 km north-east of the town of Kangerlussuaq
(Figures 8A,B). The most noticeable feature of UAV10A1-
Isunguata is the variation in surface topography between
highly crevassed and the lower relief, almost flat areas. The
crevassed regions generally occur near the margin of the
 glacier presumably where the subglacial topography is steeper.
and longitudinal and transverse strain gradients are high. The
surface topography appears to dominate the spatial albedo
variability with crevassed areas having a lower mean albedo (0.39)
than low-relief topography (0.50) (Figures 9A,B). Crevassed
regions have a lower albedo because rugged topography
increases the probability of a shading effect when radiation is
reflected between surfaces and trapped in troughs (Pfeffer
and Bretherton, 1987; Cutler and Munro, 1996). Additionally,
crevasses troughs may appear to be darker than the peaks
and plateaus of the bare ice because of the possibility of
liquid water storage due to enhanced radiative melting in the
depressions (Cathles et al., 2011). The slopes and troughs of
the crevasses could also facilitate the transport and residence
of impurities such as dust or ice algae. These impurities absorb
downward radiation and darken the bare ice in the depressions
(Figure 9B).

Supraglacial meltwater ponds and channels also show low
albedo and occupy areas of tens to hundreds of meters squared.
Supraglacial meltwater flowing over relatively clean ice has been
previously shown to have an albedo of between 0.19 and 0.26
(Ryan et al., 2016). But point sampling of surface water on
Isunguata Sermia reveals that many streams have an albedo
of between 0.09 and 0.12 due to the presence of underlying
cryoconite deposits (Hodson et al., 2007) (Figure 9A). The
cryoconite deposits cover the majority of the channel and pond
beds and lower the albedo of the observed surface. Cryoconite
holes, with a typical albedo of 0.11 when viewed from nadir, are
also apparent across Isunguata Sermia, as evinced by the small
quasi-circular black spots which are frequently observed on the
bare ice (Figure 9A) (Cook et al., 2015). Cryoconite granules
absorb solar radiation and melt down into the ice until they are
in radiative and thermodynamic equilibrium (Takeuchi et al.,
2001; MacDonell and Fittsimsons, 2008; Langford et al., 2010).
Although the holes have very low albedo when seen from directly
above, the cryoconite may become hidden at non-nadir viewing
angles and from non-zenith solar illumination (Bøggild et al.,
2010).

3.4.2. UAV10A1-S6
UAV10A1-S6 is a 70 km² map between 1,060 and 1,200 m
a.s.l. which includes the Institute for Marine and Atmospheric
(IMAU), Utrecht University weather station at S6 (Figure 8C).
The region is situated in an area of particularly low albedo
which has been defined as the “dark zone” (Wientjes and
Oerlemans, 2010; Wientjes et al., 2012; Shimada et al., 2016).
UAV10A1-S6 is predominantly characterized by a low-relief
surface topography and variable bare ice albedo ranging
between 0.27 and 0.58. The variable bare ice albedo appears
to cause the striking foliation patterns in the ice which are
also observed in satellite imagery (Figure 10A) (Wientjes and
Oerlemans, 2010; Wientjes et al., 2012; Shimada et al., 2016).
Possible causes of variable bare ice albedo include liquid water
content of the ice and/or the presence of light absorbing
impurities (Greuell, 2000; Wientjes et al., 2012; Lüthi et al., 2015).

UAV10A1-S6 is dominated by the presence of bare ice, but remnant snow can persist within ice fractures and supraglacial channel incisions exhibiting local albedo of between 0.65 and 0.71. The likely persistence of snow in these locations is due to a thicker snowpack and the redistribution of snow by the wind (Figure 10B) (van den Broeke et al., 2008). A 0.6 km² supraglacial lake in the south-west corner of UAV10A1-S6 (Figure 11A) is characterized by albedo ranging from 0.17 to 0.28. The difference in the lake's albedo is a consequence of varied water depth and the bed reflectance. At the lake's edge, slabs of bright ice are visible which are interpreted as grounded lake ice and a possible lake drainage event. Lake ice forms annually as a floating layer when the temperature of the water drops below freezing. It has therefore existed for less than a year and appears relatively impurity free with a mean albedo of 0.61.
3.4.3. UAV10A1-Behar

UAV10A1-Behar is an 80 km² map 10 km east of the S6 weather station between 1,200 and 1,350 m a.s.l. UAV10A1-Behar is at the edge of the dark zone but still within the ablation area (Figure 8D). UAV10A1-Behar is characterized by similar bare ice variability as UAV10A1-S6 but the former contains four supraglacial lakes and a more active drainage system. Large braided and single channel meltwater streams with widths of up to 25 m are common features in this area and have albedo values between 0.16 and 0.24, similar to supraglacial lakes (Figure 11B). UAV10A1-Behar concurs with recent studies which reveal that the channels are characterized by dendritic drainage patterns and ultimately terminate in a large moulin located at 67.05° N, −49.03° E (e.g., Yang and Smith, 2013; Smith et al., 2015; Yang et al., 2016). Although the lakes ephemerally store large fluxes of water over weekly to seasonal time-scales (Fitzpatrick et al., 2014), all the lakes in UAV10A1-Behar have laterally draining outlet streams showing that these features provide little obstruction to meltwater flow through the supraglacial hydrologic system (Smith et al., 2015).

4. DISCUSSION

Developments in economical remote sensing tools and the rapidly evolving field of UAVs are enabling glaciologists to obtain surface topographic and reflectance data at unprecedented spatial and temporal resolutions (Carrivick et al., 2013, 2016; Bhardwaj et al., 2016). In particular, structure-from-motion software has
facilitated the collection of topographic data using consumer-grade digital cameras (e.g., Westoby et al., 2012; Whitehead et al., 2013; Immerzeel et al., 2014; Tonkin et al., 2014; Ryan et al., 2015; Smith et al., 2016). More recently, there has been an interest in quantifying the reflectance of surfaces using similarly low-cost sensors (e.g., Hakala et al., 2010; Di Mauro et al., 2015; Rippin et al., 2015). This study has shown that it is possible to obtain albedo measurements over ice sheet surfaces with high spatial resolution and accuracy using a standard consumer-grade RGB digital camera mounted on a UAV.

The technique described has several advantages over current methods to generate background albedo fields. We overcome the commonly identified, but little addressed, challenge of variable illumination during surveying (e.g., Dumont et al., 2011; Rippin et al., 2015). Cloud cover is typical over glaciated regions and, during surveys lasting more than a couple of hours, solar zenith angles can change significantly. By continuously measuring downward radiation on the UAV, we account for these changes and can obtain high-quality reflectance data regardless of the meteorological conditions. By using the hemispherical reflectance measured by the downward facing silicon pyranometer, we account for the anisotropic reflectance of the surface and correct the bidirectional reflectance recorded by the digital images. The result is that we measure albedo with accuracies of better than ±5%, regardless of changing meteorological conditions. Although the described method relies on data collection during low solar zenith angles, to minimize the cosine response errors of pyranometer and specular reflection, an albedo field can be generated at short (i.e., daily) intervals allowing unprecedented mapping of evolving snow and ice surfaces over the melt-season. The technique and workflow outlined in this study should therefore have interest for users of UAVs including the latest proximal remote sensing solutions, such as MicaSense's RedEdge (https://www.micasense.com/rededge/) and Sequoia (https://www.micasense.com/sequoia/)

multispectral cameras, which include upward facing radiation sensors.

The visible band digital camera detects albedo variations which occur between 400 and 700 nm. However, UAV10A1 is not sensitive to albedo variations, such as those caused by changes in snow grain size, which only have an effect on albedo in the near-infrared wavelengths (Wiscombe and Warren, 1980). Improvements to the method presented would be to incorporate new instruments that are sensitive to a wider portion of the electromagnetic spectrum and are hence able to detect changes in surface reflectance at wavelengths shorter and longer than that recorded by standard digital cameras confined to the visible band (e.g., MacArthur et al., 2014; Von Bueren et al., 2015). Both the weight and cost of these sensors, such as Tetracam's ADC Snap (Tetracam, California, USA) and MicaSense's RedEdge and Sequoia (MicaSense, Seattle, USA), is decreasing and their use could become increasingly common in glaciological field studies or any other field that rely on precise high-resolution near-surface albedo data.

The methodology and UAV application described in this study can be used to evaluate and improve the parameterization of albedo in surface mass balance models. At a regional scale (e.g., the entire Greenland Ice Sheet), contemporary models either prescribe a background bare ice albedo using MODIS data (e.g., van Angelen et al., 2012; Noël et al., 2015) or simulate it as a function of melt (e.g., Zuo and Oerlemans, 1996; Lefebre et al., 2003; Alexander et al., 2014). However, these schemes may not adequately represent the sub-grid albedo variability of the ice sheet surface, leading to discrepancies between the modeled and observed surface mass balance. The high spatial resolution of UAV10A1 therefore presents an opportunity to assess the representativeness of albedo parameterizations and determine a grid scale at which ice albedo is sufficiently represented.

Another application of the technique is in the collection of UAV-based albedo data over areas where surface instruments cannot be deployed. Access to glaciers and ice sheets is often
hindered by steep slopes, crevasses, and impassable meltwater channels. But small UAVs can be deployed to survey these areas without risk to the observer. UAV10A1 may also be used to investigate the robustness of in-situ albedo observations for validating satellite-derived albedo. The calibration and validation of satellite albedo products relies upon in-situ measurements but direct comparison between the two is only valid when the footprint of the in-situ measurement is either the same as the corresponding satellite image pixel or when the surface is sufficiently homogeneous at the scale of both the in-situ measurement and satellite data pixel (Román et al., 2009, 2013). UAV10A1 provides an opportunity to assess whether the spatial heterogeneity of ice sheet and glacier surfaces is captured by the in-situ measurement footprint and whether direct comparison of such a “ground truth” with satellite-derived albedo data is fully justified.

5. CONCLUSION

The technique presented in this study demonstrates that an accurate (uncertainty less than 5%), high resolution and spatially continuous albedo product can be derived using a combination of a digital camera, broadband pyranometers, and UAV platform. This technique is reproducible and provides a practical approach for overcoming the limitations of previous studies caused by variable illumination. The method is demonstrated using measurements collected by a fixed-wing UAV across the Kangerlussuaq sector of the Greenland Ice Sheet in the 2015 melt-season. Validation of the albedo product, UAV10A1, indicates it has an RMSE of 4.9% in comparison to ground-based albedo measurements and an RMSE of 3.7% when compared to MOD10A1. UAV10A1 has numerous applications such as evaluating the accuracy of albedo fields used in energy balance modeling and understanding spatial and temporal variability of ice sheet albedo. The miniaturization of multispectral sensors and incorporation of upward facing radiation sensors on UAV packages means that this technique could become increasingly common in glaciological field studies and further developed to measure a wider range of wavelengths.

AUTHOR CONTRIBUTIONS

JR, AH, JB, and NS conceived the study. NS, JR, SB, and AH designed, tested and built the UAV used for data collection. JR and JB led data acquisition. JR developed and implemented the technique and wrote the article. AH provided supervision. JB collected handheld CM3 albedo measurements. AH, JB, MC, LP, AR, LS, MS, KC, TH, CJ, and SD provided support in the field and contributed to acquisition of data. AH, JB, SB, AE, SD, JC, and TI gave conceptual and technical advice and supported the interpretation of data. All authors edited and critically revised the article.

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Chapter 3

How robust are in situ observations for validating satellite-derived albedo over the dark zone of the Greenland Ice Sheet?

3.1 Summary

Chapter 3 uses digital imagery acquired by a UAV to evaluate point-to-pixel albedo comparisons across the western, ablating margin of the Greenland Ice Sheet. Calibration and validation of satellite-derived ice sheet albedo data require high-quality, in situ measurements commonly acquired by up and down facing pyranometers mounted on automated weather stations (AWS). However, direct comparison between ground and satellite-derived albedo can only be justified when the measured surface is homogeneous at the length-scale of both satellite pixel and in situ footprint. This study therefore tests whether this assumption is valid in the bare ice zone of the ice sheet and, by doing so, addresses the second major objective of the thesis which is to demonstrate how UAVs can be used to bridge the spatial gap between in situ and satellite measurements.

We find that the surface of the Greenland Ice Sheet is often not uniform at the scale of both
the AWS and satellite pixel due to the presence of impurities, surface water and crevasses. This means that a point measurement of albedo might not capture the full variability of the surface, resulting in discrepancies when compared to satellite image pixels. Furthermore, AWS are usually located on safe areas of flat, bare ice or snow, so they usually overestimate reflectivity in comparison to the satellite pixel. We argue that if unrepresentative ground measurements are removed from satellite comparison exercises then the uncertainty in satellite products could be reduced. Hence, the long-term decline in Greenland Ice Sheet reflectivity between 2000 and 2012 might be more significant than previously thought.

3.2 Contribution

To complete this study, JR designed, tested and built the UAVs used for data acquisition. JR piloted the UAV during the summer of 2015 and obtained imagery over the two AWS in June and July. JR downloaded and processed all the MODIS satellite data obtained from NSIDC (https://nsidc.org/data/mod10a1) and AWS data obtained from PROMICE (https://promice.org/DataDownload). JR compared the AWS and MODIS and quantified the discrepancies between the two datasets. JR developed the method to quantify the spatial heterogeneity of the ice sheet surface from UAV imagery and assess the representativeness of each in situ albedo measurement. JR wrote the first draft of the manuscript and edited it in response to suggestions from co-authors and one reviewer. JR made all the figures.
How robust are in situ observations for validating satellite-derived albedo over the dark zone of the Greenland Ice Sheet?

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Abstract Calibration and validation of satellite-derived ice sheet albedo data require high-quality, in situ measurements commonly acquired by up and down facing pyranometers mounted on automated weather stations (AWS). However, direct comparison between ground and satellite-derived albedo can only be justified when the measured surface is homogeneous at the length-scale of both satellite pixel and in situ footprint. Here we use digital imagery acquired by an unmanned aerial vehicle to evaluate point-to-pixel albedo comparisons across the western, ablating margin of the Greenland Ice Sheet. Our results reveal that in situ measurements overestimate albedo by up to 0.10 at the end of the melt season because the ground footprints of AWS-mounted pyranometers are insufficient to capture the spatial heterogeneity of the ice surface as it progressively ablates and darkens. Statistical analysis of 21 AWS across the entire Greenland Ice Sheet reveals that almost half suffer from this bias, including some AWS located within the wet snow zone.

Plain Language Summary Ground measurements of reflectivity, such as those made by automated weather stations, are often used to determine the accuracy of satellite measurements. But the footprints of the instruments mounted on automated weather stations are usually much smaller than the pixel of the satellite image, meaning that comparison between the two is only justified when the surface is relatively uniform. We use high resolution imagery collected by a UAV to demonstrate that the surface of the Greenland Ice Sheet is often not uniform at the scale of both the weather station and satellite pixel due to the presence of impurities, surface water and crevasses. This means that a point measurement of reflectivity might not capture the full variability of the surface, resulting in discrepancies when compared to satellite image pixels. Furthermore, weather stations are usually located on safe areas of flat, bare ice or snow, so they usually overestimate reflectivity in comparison to the satellite pixel. We argue that if unrepresentative ground measurements are removed from satellite comparison exercises then the uncertainty in satellite products could be reduced. Hence, the long-term decline in Greenland Ice Sheet reflectivity between 2000 and 2012 might be more significant than previously thought.

1. Introduction

Surface albedo modulates the absorption of incoming shortwave radiation and is a primary factor governing the surface energy balance and ablation of the cryosphere [e.g., Braithwaite and Olesen, 1990; Knap and Oerlemans, 1996; Broeck et al., 2000; van den Broeke et al., 2011]. Accurate measurements of albedo are therefore critical to understanding spatial patterns of melt, and an essential input to models for reliable prediction of surface runoff and the concomitant contribution of eustatic sea level rise from glaciers and ice sheets. Due to its inherent spatial and temporal variability, interpolating surface albedo from extremely sparse in situ measurements fails to represent albedo patterns realistically, particularly across the ablation zone. Hence, satellite remote sensing provides the only practicable method for accurate determination of spatial and temporal patterns of snow and ice albedo for constraining regional climate, melt, and runoff models across Greenland ice Sheet and elsewhere [Henderson-Sellers and Wilson, 1983].

Retrieval of surface albedo from satellite data is a complex process, dependent on the performance of atmospheric correction and the accuracy of the angular model used to describe the bidirectional reflectance.
distribution function [Ricchiazzi et al., 1998; Liang, 2001; Klein and Stroeve, 2002; Schaaf et al., 2002]. The calibration and validation of satellite albedo products therefore rely upon in situ measurements, the majority of which are made by broadband pyranometers mounted on automated weather stations (AWS) [Liang et al., 2005; Stroeve et al., 2005, 2006, 2013]. However, direct comparison is only valid when the footprint of the in situ measurement is either the same as the corresponding satellite image pixel or when the surface under scrutiny is homogeneous at the spatial scale of both the in situ measurement and satellite data pixel [Román et al., 2009, 2013; Shuai et al., 2011].

To date, quality assessment and calibration of satellite-derived albedo across ice and snow make an implicit assumption of surface homogeneity [e.g., Liang et al., 2005; Stroeve et al., 2005, 2006, 2013], which may be valid across relatively flat and uniform snow surfaces, such as the accumulation zone of the Greenland and Antarctic Ice Sheets. However, below the transient snowline, particularly during the summer melt season, ablating ice surfaces are not uniform. In Greenland, the ablation zone is comprised of a time-varying mixture of snow patches, ice with varying grain sizes, roughness features, biotic and abiotic impurities, and surface and shallow-subsurface water [Baggild et al., 2010; Gardner and Sharp, 2010; Moustafa et al., 2015; Ryan et al., 2017]. Ablating ice surfaces may therefore not be homogeneous at the scale of both the in situ measurement and the satellite pixel, leading to a potential discrepancy between the two measurements [Knap and Oerlemans, 1996]. Furthermore, because the in situ measurement is assumed to be accurate, and indeed, is often considered a “ground truth,” discrepancies are frequently attributed to bias in the satellite-derived albedo product [e.g., Stroeve et al., 2006]. This results in loss of confidence in satellite-derived albedo retrieval due to incorrect error attribution, thereby diminishing the statistical significance of long-term albedo trends and diluting capacity to accurately monitor the Earth’s cryosphere [e.g., Box et al., 2012; He et al., 2013; Stroeve et al., 2013; Alexander et al., 2014].

This study investigates whether direct point-to-pixel comparisons for calibration and validation across the Kangerlussuaq (K) transect of the Greenland Ice Sheet are justified. First, we evaluate the difference between in situ albedo, measured at three AWS, and the MODerate resolution Imaging Spectroradiometer (MODIS) albedo product, MOD10A1. Second, we quantify the spatial heterogeneity of the satellite pixel, in which each AWS is situated, using 20 cm pixel resolution aerial imagery acquired by a fixed-wing unmanned aerial vehicle (UAV). Finally, we assess whether the spatial heterogeneity of the surface is captured by the AWS footprint and whether direct comparison with satellite-derived albedo data is robust and justified.

2. Data and Methods
2.1. Satellite Albedo
Satellite albedo retrievals between 2002 and 2016 were obtained from the MODIS daily albedo (300 to 3000 nm) product, MOD10A1 Collection 6 (C6), collected by NASA’s Terra satellite [Hall and Riggs, 2016]. MOD10A1 is provided at 463 m (0.21 km²) pixel size, in a sinusoidal projection by the National Snow and Ice Data Center (NSIDC) [Hall and Riggs, 2016]. MOD10A1 was chosen because it is commonly used for estimating spatial and temporal albedo trends across the Greenland Ice Sheet [e.g., Box et al., 2012; Alexander et al., 2014] and surface mass balance modeling [e.g., van As et al., 2012]. Artifacts in MOD10A1 caused by undetected clouds, aircraft contrails, or shadows were filtered using the 11 day statistics technique proposed by Box et al. [2012] (Text S1 in the supporting information).

2.2. In Situ Albedo
In situ albedo measurements between 2009 and 2016 were obtained from three AWS (KAN-L, KAN-M, and KAN-U) situated on the K transect [Figure 1, Text S2; Ahlstrøm et al., 2008; van As, 2011]. Black thermopile Kipp & Zonen CNR1 or CNR4 net radiometers are mounted on the AWS, which measure downward and upward shortwave radiation fluxes with a specified uncertainty of less than 5% [van den Broeke et al., 2004; van As et al., 2012]. In the absence of accumulated snow, upward shortwave radiation is measured at a height of 2.8 m above the surface [van As, 2011]. The instruments have a field of view of 150° which yields a maximum ground footprint diameter of 21 m, equating to an area of 346 m². However, the effective footprint is smaller since the radiometers’ cosine response means that they are inherently biased towards incident radiation at angles perpendicular to (i.e., from directly beneath) the sensor.
2.3. Quantifying Surface Spatial Heterogeneity From Aerial Imagery

Visible wavelength (RGB) digital imagery was acquired by a fixed-wing UAV described by Ryan et al. [2015, 2017] (Text S3). We assume for the purposes of our analyses that the visible band imagery adequately captures the albedo variability of the surfaces surrounding the AWS. This assumption is justified since over half of the total solar energy arrives in the visible wavelengths [Painter et al., 2012] and that most of the variability in reflected radiation for bare ice surfaces and snow with uniform grain size occurs in the visible band (350 to 695 nm) of the shortwave spectrum [Warren and Wiscombe, 1980; Cutler and Munro, 1996] where our Sony NEX-5N camera is most sensitive [Jiang et al., 2013].

Semivariograms were constructed from the RGB aerial imagery to assess the spatial heterogeneity of the surfaces surrounding KAN-L and KAN-M (Text S4). The range of a semivariogram defines the distance from a point beyond which there is no further spatial correlation associated with that point [Isaaks and Srivastava, 1989; Román et al., 2009, 2013]. We quantify this point by fitting an exponential function to the semivariogram and define the range (otherwise known as the practical sampling distance) as the ordinate value at which the exponential function reaches 99.9% of the maximum semivariance or sill. If the range is smaller than the AWS pyranometer footprint, then it is apparent that the in situ measurement represents the spatial variability of the surface and can justifiably be used to validate MOD10A1. However, if the range is larger than the footprint of the in situ measurement, then the AWS does not capture the full spatial variability of the surface, which would lead to a bias between AWS and MOD10A1 albedo.

3. Results

3.1. Validation of Satellite Albedo Using AWS

The mean annual root-mean-square difference (RMSD) between AWS and MOD10A1 albedo varies between 0.025 and 0.077 and has a mean of 0.047 (Table 1). The mean annual bias between AWS and MOD10A1 is positive for all three AWS on the K transect (+0.016) suggesting that either MOD10A1 underestimates albedo or that the in situ measurements overestimate albedo. The mean bias becomes more positive as the melt season progresses from April/May (−0.017) to June (+0.034) and July (+0.056) and has a maximum of 0.10 at KAN-L (Table 1). The mean RMSD also increases from April/May (0.037) to June (0.040) and July (0.061) (Table 1). The MOD10A1-AWS bias has a statistically significant ($p < 0.001$) positive correlation with time since 1 April (Figures 2a–2c) and a statistically significant negative correlation with albedo (Figures 2d–2f).

**Table 1.** RMSD and Bias Between Albedo Measured by the Three K Transect AWS and MOD10A1 (2009 and 2016)$^a$

<table>
<thead>
<tr>
<th>AWS</th>
<th>Annual RMSD</th>
<th>Annual Bias</th>
<th>April/May RMSD</th>
<th>April/May Bias</th>
<th>June RMSD</th>
<th>June Bias</th>
<th>July RMSD</th>
<th>July Bias</th>
</tr>
</thead>
<tbody>
<tr>
<td>KAN-L</td>
<td>0.077</td>
<td>0.025</td>
<td>0.067</td>
<td>−0.029</td>
<td>0.073</td>
<td>0.072</td>
<td>0.100</td>
<td>0.100</td>
</tr>
<tr>
<td>KAN-M</td>
<td>0.038</td>
<td>0.013</td>
<td>0.028</td>
<td>−0.013</td>
<td>0.024</td>
<td>0.017</td>
<td>0.056</td>
<td>0.052</td>
</tr>
<tr>
<td>KAN-U</td>
<td>0.025</td>
<td>0.009</td>
<td>0.017</td>
<td>−0.009</td>
<td>0.022</td>
<td>0.013</td>
<td>0.026</td>
<td>0.017</td>
</tr>
<tr>
<td>Mean</td>
<td><strong>0.047</strong></td>
<td><strong>0.016</strong></td>
<td><strong>0.037</strong></td>
<td><strong>−0.017</strong></td>
<td><strong>0.040</strong></td>
<td><strong>0.034</strong></td>
<td><strong>0.061</strong></td>
<td><strong>0.056</strong></td>
</tr>
</tbody>
</table>

$^a$Bold numbers are the mean RMSD or bias of the three weather stations.
3.2. Spatial Heterogeneity of Ice Sheet Surface

Analysis of the UAV digital aerial imagery demonstrates that the surface of the ice sheet is spatially heterogeneous and changes significantly through time (Figure 3). In June, the surface surrounding KAN-L is predominantly characterized by snow but it is not deep enough to obscure underlying changes in surface topography (Figure 3a). By 14 July, the snow has completely melted leaving a predominantly bare ice surface with undulating topography (Figure 3b). A relatively homogeneous snow surface characterizes KAN-M in June which is replaced by ponded meltwater and superimposed ice by July (Figures 3c and 3d). The ice at the surface is characterized by different shades of grey, presumably depending on its saturation, crystallography, and impurity load.

Our derived semivariograms reveal that the practical sampling distance and area over which the inherent spatial variability of the ice sheet surface is captured depends on both the time of year and the location (Figure 4). In June, the exponential function attains the sill at separation distances of between 15 and 20 m indicating that albedo variability at these sites can be captured over relatively short sampling distances. Later in the melt season, the semivariograms do not plateau until between 45 to 50 m revealing that longer sampling distances are necessary to capture the spatial variability of the ablating ice sheet albedo. Furthermore, the sill or maximum semivariance of the ice sheet surfaces increases by an order of magnitude at both KAN-L and KAN-M from June to July (Figure 4). This indicates that the spatial heterogeneity of the ice sheet surface increases as the melt season progresses.

4. Discussion and Implications

Our analysis reveals that spatial variability in the melt processes and ablating surfaces across the Greenland Ice Sheet cause in situ measurements on the K transect to overestimate albedo (Table 1 and Figure 2). In June, our aerial imagery reveals that the, predominantly snow covered, surface is relatively homogeneous and that albedo sampled at length scales between 15 and 20 m is sufficient to capture that albedo variability (Figures 3 and 4). However, as the melt season progresses, more bare ice is exposed, the albedo of the ablation zone reduces and sampling distances of between 45 and 50 m are required to fully represent the inherent heterogeneity of the ice surface and associated albedo reduction (Figures 3 and 4). The increase in surface
heterogeneity is associated with an increase in the bias between AWS and MOD10A1 as represented by the negative correlation with albedo and positive correlation with time since 1 April (Table 1 and Figure 2). We therefore argue that the increase in the bias between MOD10A1 and AWS is a result of in situ measurements becoming less representative of the surrounding surfaces [e.g., Knap and Oerlemans, 1996].

Figure 3. High-resolution (20 cm) aerial images corresponding to the AWS locations within their respective MODIS pixels. The AWS footprint and practical sampling distance are shown by the black and red circles, respectively. The images are projected in UTM 22 N.

heterogeneity is associated with an increase in the bias between AWS and MOD10A1 as represented by the negative correlation with albedo and positive correlation with time since 1 April (Table 1 and Figure 2). We therefore argue that the increase in the bias between MOD10A1 and AWS is a result of in situ measurements becoming less representative of the surrounding surfaces [e.g., Knap and Oerlemans, 1996].

Figure 3. High-resolution (20 cm) aerial images corresponding to the AWS locations within their respective MODIS pixels. The AWS footprint and practical sampling distance are shown by the black and red circles, respectively. The images are projected in UTM 22 N.

Figure 4. Semivariograms showing how the semivariance of the ice surface changes with increasing separation distance between samples between June and July. Between June and July, the sampling distance needed to capture the heterogeneity of the surface becomes larger than the 21 m diameter footprint of the AWS-mounted pyranometers. Note the varying y axis scale.
We argue that in situ measurements mainly overestimate, rather than underestimate, albedo because the increase in spatial variability as the melt season progresses is predominantly associated with an increase in the extent of low albedo surfaces (Figure 3). These include bare ice with varying concentrations and types of impurity, cryoconite holes, surface water, crevasses, and rough/steep topography, all of which have a lower albedo of between 0.10 and 0.27 [Bøggild et al., 2010; Ryan et al., 2016]. In situ measurements are likely to undersample these darker surfaces because AWS are preferentially deployed on flat areas of bare ice, rather than meltwater channels or crevasses, to reduce tilt and reduce the risk of loss or inundation. Such bare ice surfaces, with a mean albedo higher than 0.50 [Ryan et al., 2016], are inherently brighter than the albedo of the corresponding MOD10A1 pixel footprint, which will capture the larger area including lower albedo surface types. This results in a systematic discrepancy between the in situ measurement and MOD10A1 product. Furthermore, as the melt season progresses, the extent of impurity-rich bare ice, surface water, and cryoconite holes tends to increase [e.g., Fitzpatrick et al., 2014; Chandler et al., 2015] which further drives this spatially derived bias between in situ and satellite albedo until snowfall resets the surface (Figure 2).

The bias between MOD10A1 and KAN-U indicates that albedo overestimation is not limited to in situ measurements in the ablation zone but is also evident in the accumulation zone (Figures 2e and 2f). While KAN-U was out of range for our UAV imagery, given sustained subzero temperatures at an elevation of 1800 m asl, we expect the surface to be snow covered, similar to KAN-M in mid-June (Figure 3c). The point-to-pixel bias at KAN-U increases between April/May and July suggesting that the heterogeneity of the albedo over snow later in the melt season is also insufficiently captured by the AWS pyranometer (Figures 2e and 2f). An extended analysis of 21 Greenland AWS suggests that the bias due to spatial heterogeneity in surface types reported here for the K sector of the ice sheet may potentially impact up to half of the PROMICE AWS network (Table S1 and Figures S1 and S2). The bias between MOD10A1 and AWS has a significant (p < 0.001) negative correlation with albedo for nine (43%) AWS (Figures S3 and S4) and a significant positive correlation over time from 1 April for seven (33%) AWS (Figures S5 and S6).

Correlations are particularly strong for AWS in the ablation zone of the K transect (Figures S1 and S2). These AWS are situated in the dark zone of the ice sheet where it has been reported that dust deposited in the accumulation zone throughout the Holocene is now emerging [Wientjes and Oerlemans, 2010; Shimada et al., 2016]. The spatial variability and seasonal redistribution of this dust may cause the surface surrounding the K transect AWS to be more heterogeneous than elsewhere on the ice sheet, leading to higher biases between MOD10A1 and AWS. At some sites outside of the K transect, the AWS-MOD10A1 bias is negatively correlated over time from 1 April (Table S1 and Figures S5 and S6). Surface processes may be responsible for these trends as well. For example, in the ablation zone, late-season snowfall which fills depressions and bridges gullies could preferentially increase the albedo of the surface outside of the AWS footprint [Smeets and van den Broeke, 2008]. Likewise, in the accumulation zone, the AWS may continue to observe the albedo of old snow during the melt season because freshly fallen snow is redistributed away from the AWS but remains inside the corresponding MODIS pixel [Lenaerts et al., 2012]. We did not observe these processes in our UAV imagery and any testing of these specific hypotheses would require further UAV surveys over additional sites.

Improving the representativeness of in situ albedo measurements can be achieved by increasing the height of the pyranometers above the surface (at 6.7 m the ground footprint would be 50 m), but this may be impractical for leveling the AWS and, even then, the cosine response of the pyranometers means that they would still be biased towards surfaces directly beneath them. Alternatively, attempts could be made to locate AWS at sites where the surface of the ice sheet is more homogeneous. These include areas of low strain where crevassing and fracturing is minimal, or where the distribution of impurities in the ice is more uniform. With this in mind, future research might benefit from installing local wireless networks of in situ pyranometers within a single MODIS pixel for site-specific satellite validation exercises. We also recommend implementing UAV surveys and the techniques outlined in this study to characterize the spatial heterogeneity of the surface during visits to AWS. It is also worth noting that other parameters measured by AWS used in regional climate and energy balance models may also suffer from the sampling biases documented here. For example, turbulent heat fluxes measured by AWS over flat surfaces are unlikely to truly represent complex flow over areas that have steep, rough, and crevassed topography.

Confidence in MODIS albedo can be improved by systematically ignoring areas of the ice sheet with high spatial variability in validation exercises and only selecting representative in situ measurements. Our analysis
indicates that the most representative measurements are likely to be found in the dry snow zone or during April and May when the ice sheet surface consists of a more homogeneous snow surface. In both these cases, the assumption that the ice sheet surface is homogeneous at both the scale of in situ and satellite image pixel is likely robust and the in situ albedo provides a more justifiable validation of satellite albedo retrievals. However, we note that this would only verify the satellite albedo product for relatively simple, near-Lambertian scattering surfaces provided by snow and would not hold true for the darker and optically complex ablating ice exposed later in the season.

It has previously been reported that Greenland Ice Sheet summer albedo declined by between 0.06 and 0.08 in the ablation zone and between 0.01 and 0.04 in the accumulation zone between 2000 and 2012 [Box et al., 2012; He et al., 2013; Stroeve et al., 2013; Alexander et al., 2014]. These long-term albedo trends are comparable to or below the 0.041 to 0.075 stated uncertainty of MODIS albedo products (e.g., MOD/MYD10A1 and MCD43A3) [e.g., Stroeve et al., 2006, 2013; Box et al., 2012] which raises questions regarding the true significance of these trends and current health of the cryosphere [e.g., Polashenski et al., 2015]. However, these reported uncertainties are based on the assumption that in situ AWS-based measurements provide an absolute ground truth. Here we show this assumption to be invalid for the ablating ice and snow over the K transect. We argue that if unrepresentative in situ measurements were removed from satellite validation exercises as outlined above, then the uncertainty in MODIS albedo products might be reduced to ~0.03. Such a result would improve statistical inferences regarding albedo decline across the Greenland Ice Sheet and, likewise at ablating ice masses elsewhere, increasing efficacy and confidence in the assimilation of albedo into regional climate and melt runoff models and prediction of global sea level rise.

5. Conclusions

We investigated temporal patterns of bias between in situ and satellite-derived albedo at three AWS situated on the K transect of the Greenland Ice Sheet. Aerial imagery acquired by a fixed-wing UAV allowed us to quantify the spatial heterogeneity at two AWS sites and determine whether comparison between albedo measured by in situ measurements and satellite products is justified. Our results suggest that the ice sheet surface is not necessarily homogeneous at both the scale of the AWS pyranometer footprint and the MODIS pixel footprint and that caution must be exercised when validating satellite albedo retrievals using in situ (e.g., AWS) measurements. At two sites in the ablation zone, aerial imagery demonstrates that AWS-mounted pyranometer footprints are insufficiently large to sample the true spatial heterogeneity of ice surface albedo in July, and hence, the in situ measurement cannot be justified as a valid ground truth. In situ, AWS-derived measurements tend to overestimate albedo, an issue that potentially affects almost half of the network of 21 AWS across the Greenland Ice Sheet that we analyzed and results in a degradation of precision and confidence in satellite-derived albedo products validated by this method.

References


Introduction

The supporting information includes an extended description of processing applied to the satellite and weather station albedo data and the geostatistical methods used to quantify the heterogeneity of the ice sheet surface from the UAV digital imagery. It also includes an extended comparison between MODIS and AWS albedo at 21 sites on the Greenland Ice Sheet.

Text S1: MODIS albedo

There is evidence of possible sensor degradation in the Terra MODIS instrument, which was uncorrected in MOD10A1 Collection 5 (C5) [Lyapustin...
et al., 2014; Polashenski et al., 2015]. Therefore we estimated the validity of MOD10A1 Collection 6 (C6) by comparing it with MYD10A1 C6, which is collected by the Aqua satellite and is assumed to have no sensor degradation. We performed the comparisons by calculating the rolling difference between MOD10A1 and MYD10A1, using a 30-day window between 2003 and 2016, and implementing a Dickey-Fuller statistical test for stationarity [Dickey and Fuller, 1979].

Values of between -16.3 to -11.8 from the Dickey-Fuller test are much less than the critical value (10%) of -3.43, revealing that, at least statistically, the bias between MOD10A1 and MYD10A1 is stationary. The latest MODIS collection (C6) therefore suitably accounts for Terra MODIS sensor degradation, which was uncorrected for in the previous collection (C5), and shows that MOD10A1 can be reliably used to estimate the albedo of the Greenland Ice Sheet between 2003 and 2016 [Lyapustin et al., 2014; Polashenski et al., 2015]. This is important because the Aqua MODIS instrument suffered a failure in the near infrared band, reducing its cloud detection capability, and reducing the reliability of MYD10A1 daily albedo data [Stroeve et al., 2006; Box et al., 2012].

MOD10A1 albedo was filtered using two techniques proposed by Box et al. [2012]. First, we removed MOD10A1 values higher than 0.84. The rational behind this is that the maximum clear sky albedo of freshly fallen snow does not exceed 0.84 [Konzelmann and Ohmura, 1995]. Values higher than this threshold in MOD10A1 are likely caused by low viewing angles or high solar zenith angles early or late in the melt-season [Klein and Stroeve, 2002; Stroeve et al., 2006; Box et al., 2012].

Second, artefacts in MOD10A1 caused by undetected clouds, aircraft condensation trails or shadows were removed by calculating the difference between daily albedo and an 11-day running mean. Those values that exceeded two standard deviations were removed unless they were within 0.04 of the 11 day running median [Box et al., 2012].

Text S2: In-situ albedo and comparison to satellite albedo

Daily albedo for 21 AWS across the Greenland Ice Sheet was calculated by integrating hourly upward and downward shortwave radiation measurements (by Kipp & Zonen CNR1 or CNR4 net radiometers) over a 24 hour period [e.g., Stroeve et al., 2005; Bogren et al., 2016] and filtered in the same way as the MOD10A1 data, using 11-day statistics. Downward shortwave radiation was corrected for tilt and hourly albedo was only calculated when solar zenith angles were larger than 70° [van As, 2011].

The mean annual RMSD between AWS and MOD10A1 albedo varies between 0.025 and 0.077 and has a mean of 0.047 (Table 1). This value is comparable to previous uncertainty estimates of the MOD10A1 product [e.g., Stroeve et al., 2006; Box et al., 2012; van As et al., 2012; Alexander et al., 2014]. For example, Box et al. [2012] found an RMSD of 0.041 and Stroeve et al. [2006] found a 0.067 RMSD between MOD10A1 and the GC-Net AWS network. Our
annual mean RMSD is ~0.02 less than that of Stroeve et al. [2006] because we used C6 MOD10A1 data, which are corrected for the sensor degradation that affected C5 (Text S1), and because we applied multi-day smoothing to both the MOD10A1 and AWS data that filters artefacts and outliers. Our RMSD value is higher than that estimated by Box et al. [2012] because the majority of the GC-Net AWS are situated in the accumulation zone where point-to-pixel comparisons are more likely to be valid whereas only one of the three K-transect AWS used here (KAN-U) is located in the accumulation zone.

Text S3: Aerial imagery

RGB aerial imagery was acquired by a fixed-wing UAV which was launched from a camp situated near KAN-B (Fig. 1). The UAV overflew two ice sheet AWS (KAN-L and KAN-M) at a mean height of ~600 m above the surface in June and July 2015. At this altitude above the surface, the imagery obtained by the Sony NEX-5N digital camera mounted onboard the UAV, has a footprint of around 600 x 900 m and a horizontal pixel resolution of ~15 cm. The images were georeferenced using the latitude, longitude, altitude and attitude data recorded by the UAV flight controller, and mosaicked using Agisoft PhotoScan Pro (http://www.agisoft.com/). The orthomosaics were nearest neighbors resampled to a common pixel resolution of 20 cm, clipped to the same size as an MOD10A1 pixel, and reprojected in UTM 22N.

Text S4: Quantifying surface spatial heterogeneity

The RGB aerial imagery was used to quantify the spatial heterogeneity of the snow/ice surfaces surrounding KAN-L and KAN-M AWS. To do this, we randomly sampled 1500 pixels surrounding KAN-L and KAN-M. Randomly sampling 1500 pixels was deemed an optimal trade-off between adequately representing the variability of the ice sheet surface and processing time. Semivariance was calculated for separation distances between up to 100 m at 2 m intervals. For more information about semivariograms and geostatistical techniques, see Isaaks and Srivastava (1989), Roman et al. [2009, 2013] and Shuai et al. [2011].
<table>
<thead>
<tr>
<th>AWS</th>
<th>MOD10-AWS bias vs albedo</th>
<th>MOD10-AWS bias vs time since 1 April</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Pearson’s r</td>
<td>p</td>
</tr>
<tr>
<td>KPC-L</td>
<td>-0.078</td>
<td>&gt;0.001</td>
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</tr>
<tr>
<td>SCO-U*</td>
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*AWS in accumulation area

**Table S1.** Statistical values for the correlation between MOD10A1-bias and albedo and time since 1 April for 21 Greenland AWS. Locations of AWS in Fig. S1.
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Mean 0.048 0.016 0.045 0.011 0.053 0.032 0.051 0.024

*AWS in accumulation area

Table S2. Annual and monthly RMSD and bias between albedo measured by 21 GAP/PROMICE AWS and MOD10A1 (2009 and 2016). Locations of AWS in Fig. S1.
**Figure S1.** Map showing the locations of the 21 AWS used in this study to extend the analysis beyond the K-sector. The dots represent the correlation between MOD10A1-AWS bias and time since 1 April.
Figure S2. Same as Fig. S1 but showing the correlation between MOD10A1-AWS bias and albedo.
Figure S3. Relationship between MOD10A1-bias and albedo for 8 AWS in North Greenland. The Pearson’s $r$ and $p$ values for each correlation are presented in Table S2.
Figure S4. Relationship between MOD10A1-bias and albedo for 10 AWS in South Greenland. The Pearson’s $r$ and $p$ value for each correlation are presented in Table S2.
Figure S5. Relationship between MOD10A1-bias and time since 1 April for 8 AWS in North Greenland. The Pearson’s r and p value for each correlation are presented in Table S2.
Figure S6. Relationship between MOD10A1-bias and time since 1 April for 10 AWS in South Greenland. The Pearson’s $r$ and $p$ value for each correlation are presented in Table S2.
References


Chapter 4

Greenland’s dark zone controlled by distributed, biologically-active, impurities

4.1 Summary

UAV imagery collected in Chapters 2 and 3 show that the bare ice zone of the Greenland Ice Sheet displays large spatial heterogeneity in albedo and includes an area with particularly low albedo, termed the dark zone. The dark zone appears every summer mainly across the west and southwest sectors of the ice sheet. At the Arctic Circle, in the vicinity of the Kangerlussuaq (K-) sector, the dark zone extends between 20 and 75 km from the land-terminating margin where Moderate Resolution Imaging Spectroradiometer (MODIS) data indicate a regional albedo minimum of 0.34. Hitherto, the surface characteristics of dark zone remain unquantified because the spatial resolution of satellite imagery is insufficient to fully resolve specific surface types and how they distinguish the dark zone from the brighter ice and snow surfaces adjacent to it. Chapter 4 expands on Chapter 2’s identification of surface types and determines which surface types are responsible for the low albedo of the dark zone and mesoscale albedo variation across the ablation zone.

We acquired UAV imagery across a 25 km east-west transect which dissected the dark zone in
2014. Seven distinct surface types were visually identified and automatically classified based on their reflectance and roughness properties. The fractional area of each surface type in each segment was determined using a supervised k-Nearest Neighbours (k-NN) classification. We found that between 80% and 95% of the dark zone is classified as ice containing uniformly distributed impurities and that the zone has low fractional areas of surface water (<1.0%), cryoconite holes (<0.5%) and crevasses (<0.2%). We suggest, based on grey/brown hue of the impurities and biological samples from another study, that the distributed impurities consist surface ice algae. The melt-out and release of surface particulates and nutrients onto the ice surface are thought to fertilize the pigmented ice surface algae and drive a reduction in albedo the duration of the melt-season as they multiply and bloom.

4.2 Contribution

To complete this study, JR designed, tested and built the UAVs used for data acquisition. JR piloted the UAV during the summer of 2014 and obtained imagery across the dark zone in August. JR downloaded and processed all the MODIS satellite data obtained from NSIDC (https://nsidc.org/data/mod10a1). JR developed the method to classify the seven distinct surface types from UAV images and quantified their fractional areas. JR wrote the first draft of the manuscript and edited it in response to suggestions from co-authors and nine reviewers. JR made all the figures.
Greenland’s dark zone controlled by distributed, biologically-active, impurities

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Abstract
Albedo, a primary control on ice surface melt, varies considerably across the ablation area of the Greenland Ice Sheet. However, the specific surface types that comprise the western dark zone, an area of bare ice with particularly low albedo, are unknown. Here, we use UAV imagery to attribute seven distinctive surface types to observed patterns of albedo along a 25 km transect across this sector. Our results reveal that, during the melt-season, the fractional area of ice containing distributed surface impurities – an admixture of dust, black carbon and pigmented algae – explains 73% of the observed spatial variability in albedo, and is responsible for the dark zone itself. Crevassing and supraglacial water also drive albedo reduction but due to their limited extent, explain just 12% and 15% of the observed variability, respectively. Cryoconite, concentrated in large holes or fluvial deposits, is the
The Greenland Ice Sheet has become the largest cryospheric contributor to global sea level rise predominantly through increased surface melt and runoff, which accounts for ~60% of its mass loss since 1991\textsuperscript{1-4}. The dominant energy source for snow and ice melt is direct solar shortwave radiation, the absorption and reflection of which is predominantly modulated by surface albedo\textsuperscript{5-7}. Accurately constraining spatiotemporal patterns of albedo across the ice sheet is hence fundamental to understanding and predicting surface melt and runoff along with their impact on ice sheet flow dynamics and sea-level rise. A conspicuous feature of Greenland’s ablation area is its dark zone, an area of bare ice with particularly low albedo that appears across the west and southwest sectors of the ice sheet each summer\textsuperscript{8-10}. At the Arctic Circle, in the vicinity of the Kangerlussuaq (K-) sector, the dark zone extends between 20 and 75 km from the land-terminating margin where Moderate Resolution Imaging Spectroradiometer (MODIS) data indicate a regional albedo minimum of ~0.34\textsuperscript{8}. From 2000 to 2012 the spatial extent of the dark zone increased by 12% but also exhibited considerable interannual variability\textsuperscript{11, 12}. The extent of the dark zone is weakly positively correlated with air temperature and negatively correlated with solar radiation during June, July and August (JJA)\textsuperscript{11, 12}. This suggests that the ongoing albedo decrease observed during the melt-season is not simply driven by melting of the winter snowpack to reveal the darker bare ice surface beneath, but, following exposure, there are changes in the nature of the bare ice itself\textsuperscript{11, 12}. However, the surface characteristics of the dark zone remain unquantified because the spatial resolution of satellite imagery is insufficient to fully resolve the specific surface types that comprise it, and how these surfaces evolve through time, distinguishing the dark zone from brighter ice surfaces adjacent to it.

Previous field-based, in-situ observations indicate that western Greenland’s ablation zone is characterized by highly variable non-ice constituents and surface structures\textsuperscript{8, 13, 14}. These include features such as crevasses, fractures and foliations\textsuperscript{15, 16}; supraglacial hydrological features, including streams, rivers, ponds and lakes\textsuperscript{17, 18}; snow patches and fracture cornices; cryoconite, concentrated in holes or in supraglacial fluvial deposits\textsuperscript{19, 20}; microbes and their humic by-products\textsuperscript{21-23}; mineral dust and
aerosols from outcropping or contemporary aeolian deposition including black carbon from wildfires, and other aerosols. Whilst the highest resolution optical satellite imagery currently available has facilitated the examination of crevasse fields and surficial hydrology, a quantitative assessment of the specific ice surface types that comprise the dark zone, and how they combine to yield observed albedo patterns across the ablation zone of the ice sheet, has yet to be made.

Here, we utilize high-resolution (15 cm pixel size) imagery acquired from an unmanned aerial vehicle (UAV) to characterize the specific ice surfaces across the dark zone and determine their impact on the mesoscale (1 to 10 km) albedo distribution during peak melt-season, as represented by the MODIS albedo product, MOD10A1. On 8 August 2014, a fixed-wing UAV equipped with a digital camera and upward and downward facing pyranometers was deployed from a field camp based in the vicinity of the K-transect, S6 automated weather station (AWS) on a 25 km east-west transect dissecting the dark zone (Fig. 1). Seven distinct surface types were visually identified on the ground by an expert and automatically classified based on their reflectance and roughness properties. The survey transect was divided into sixty 500 x 500 m segments, co-located to the footprints of corresponding MODIS pixels, and the fractional area of each surface type in each segment was determined using a supervised k-Nearest Neighbours (k-NN) classification (see Methods section: Surface classification for more information) (Fig. 1, 2). Finally, the mean albedo of each surface type was derived from the digital imagery and the relative contribution of different surface types to mesoscale albedo variability (defined by MOD10A1 pixels) was calculated using principal component regression (PCR).

Results

Surface type variation along the transect

Analysis of all UAV imagery allowed us to visually identify and automatically classify seven distinct surface types across the survey transect: i) clean ice, ii) ice containing uniformly distributed impurities, iii) deep water, iv) shallow water, v) cryoconite either in holes or fluvial deposits, vi) crevasses, and, vii) snow (Fig. 3). Distinction between clean ice and ice containing uniformly distributed impurities was guided by qualitative assessment of 112 oblique and nadir photographs taken from the ground (<1cm pixel footprint) at specific study-sites around the field camp (Fig. 1). These images confirm that distributed impurities across the ice surface are responsible for bare ice albedo variability at the local scale (1 to 10 m) (Fig. 4). In order to upscale and understand the impact of these surface impurities on
the mesoscale albedo distribution of the ablation zone, we divided bare ice into two categories: clean ice, with very low impurity concentrations, and ice containing some or an abundance of impurities. It is apparent that additional categories could be defined for bare ice given sufficiently high pixel resolution, but for the purpose of this study, and considering the spectral limitations of the onboard camera, we do not attempt to. We note that it would be a fruitful direction with multi- and hyper-spectral sensor payloads. Clean ice has 57.2% aerial coverage in the lower, western half of the survey transect between 0 and 17 km (Fig. 3). In the eastern half (17 to 27 km), clean ice coverage is lower at 23.0%. Ice containing uniformly distributed impurities (Fig. 5a) varies inversely to clean ice, with a higher fraction in the eastern half (74.5%) compared to the western half of the transect (40.0%) (Fig. 3).

Cryoconite is commonly found in holes but also in fluvial deposits near supraglacial streams and lakes (Fig. 6c). In this study, cryoconite is distinguished from ice containing distributed impurities by its very low albedo which is indicative of concentrated, rather than distributed, impurities. Cryoconite has a maximum aerial coverage of 1.6% at 8 km from the western end of the transect and a mean coverage of 0.6% across the entire transect (Fig. 3). It is possible that we underestimate the fractional area of small cryoconite holes due to the limited, 15 cm pixel resolution of our UAV imagery. However, we note that smaller cryoconite holes (< 15 cm) would also be hidden from virtually all aerial and satellite imagery obtained at low solar elevation angles. Furthermore, cryoconite hole depths tend to equilibrate as the melt-season progresses, due to their low albedo and preferential radiative absorption in comparison to brighter ice surfaces surrounding them. Coincident field measurements, made during UAV image acquisition, indicate that the cryoconite holes observed in our study were well developed. The implication is that once they have attained equilibrium depth, they cease to absorb additional energy (which would make them deeper) compared to surrounding ice and hence are effectively neutralized from the effects of incoming solar radiation. For these reasons, we argue that smaller cryoconite holes had a minimal net impact on MODIS-derived albedo compared to the ice surface surrounding them.

At 23 km (all distances refer from the western start point of the UAV transect), a braided meltwater channel network with a fractional area of 5.3%, intersects the transect (Fig. 1; 5b), but otherwise, surface water comprises only 1.9% of the survey area (Fig. 3). These results are consistent with Smith et al. (2015) who found that surface water accounted for 1.4% of a ~5,000 km² bare ice area in the K-sector of the ice sheet. At 28 km, a small, 0.83 km² supraglacial lake covers 32.8% of the segment (Fig. 4).
Crevasse density is highest on the western flank between 2 and 5 km and attains maximum coverage of 5.9% at 2 km (Fig. 3, 6a). Up-glacier of 10 km, crevasses are almost entirely absent (Fig. 6b). Remnant snow patches, which persist within ice fractures and supraglacial channel incisions, attain a maximum coverage of 1.7% with a mean aerial coverage of only 0.1%, at this time of year.

**Relationship between surface types and spatial albedo patterns**

The mesoscale albedo distribution, determined from MOD10A1 data, exhibits considerable variability along the survey transect with values between 0.27 and 0.47 (Fig. 1, 3). The dark zone, which is approximately located between 18 and 27 km along the survey transect, has a mean albedo of 0.29 and is characterized by distinct and conspicuous banding that specifically relate to foliation structures apparent in Landsat 8 imagery (Fig. 1). Between 80% and 95% of the dark zone is classified as ice containing uniformly distributed impurities (mean albedo $\alpha = 0.27$), with the remaining 5% to 20% consisting of predominately clean ice (mean $\alpha = 0.55$) (Fig. 5a). Application of PCR reveals that the fractional area of ice containing uniformly distributed impurities explains 73% of the observed mesoscale albedo variability. Although not the darkest surface type observed, distributed impurities dominate the mesoscale albedo signal due to their extensive coverage and large variations in their fractional area across the survey transect. Distributed impurities have been attributed to the outcropping of aeolian dust deposited during the early Holocene$^{9,10,26}$ and/or pigmented surface algal blooms and associated humic material$^{21–23}$.

Locally, defined at the scale of a single MODIS pixel, supraglacial water, contained in both lakes and channels, has a distinct impact on albedo due to its low albedo ($\alpha = 0.19$ to 0.26). The segment of the transect that corresponds to a large braided channel network at 23 km has an albedo of 0.28 (Fig. 5b). This is ~0.02 lower than the surrounding segments with <1% surface water yet are otherwise composed of similar ice surfaces (Fig. 3, 5a). However, in comparison to distributed impurities, supraglacial water has a minor impact on the mesoscale albedo pattern and explains only 15% of the albedo variability across our survey transect. Surprisingly, the supraglacial lake located at 28 km is not associated with a significant reduction of MOD10A1 albedo. This is because it is relatively narrow (260 m width) and covers just 33% of the segment whilst 25% of the remaining segment consists of very bright clean ice faculae (mean $\alpha = 0.58$): interpreted as a shoreline exposed when the lake level dropped (Fig. 5c).
Hence, the low albedo of the lake water (mean $\alpha = 0.19$) itself is offset by the brightness of the surrounding ice surface adjacent to it yielding minimal change in the net albedo (MOD10A1) signature.

Crevassing explains 12% of albedo variability and its impact is well illustrated at 3 to 4 km where a transition into a crevasse zone yields a significant reduction in mesoscale albedo compared to the adjacent, flatter surface (Fig. 6a, b). Crevasses enhance shortwave radiation absorption; radiative transfer modelling indicates that the presence of crevasses can double the downward energy absorbed relative to a homogeneous, flat ice surface and reduce albedo by between 0.10 and 0.25. The amount of radiation absorbed by crevasses is determined by their size, orientation, density and whether they are water-filled. One of the most densely crevassed areas across the transect (5.9% fractional area) (Fig. 6a), with crevasse widths up to $\sim$10 m and depths in excess of 8 m, yields an albedo reduction of $\sim$0.06 in comparison to the segment in figure 6b which has no crevassing but similar fractions of other surface types (Fig. 3, 6b). Elsewhere, at 6 to 7 km, smaller crevasses, with mean widths of 5 m, have a reduced impact on mesoscale albedo, lowering it by only 0.02 in comparison to the control segment in figure 6b. This observation is at odds with radiative transfer modelling because the modelled crevasses were compared against a flat, clean ice surface, which is not the case here (Fig. 6b). Cathles et al. (2011) modelled crevasses with width to depth ratios similar to the crevasses we observed and found that they have a melt enhancement factor of 1.14 to 1.20 at solar zenith angles of 45°, which is in broad agreement with our findings.

Increases in the fractional area of cryoconite, either in large holes or fluvial deposits, are not particularly associated with mesoscale albedo reduction across the survey transect, and surprisingly, the lowest concentration of cryoconite is actually observed within the dark zone itself (Fig. 5a). Cryoconite only occupies a very small fraction of the total coverage (1.6% maximum and 0.6% mean) which can be explained by the nature of the cryoconite material itself. The threadlike, filamentous structure of cyanobacteria enables them to entangle debris and facilitate the formation of granules. These granules absorb more solar radiation and melt down into the ice until they are in radiative and thermodynamic equilibrium. Although this mechanism means that the cryoconite hole has a very low albedo value (mean $\alpha = 0.10$) when observed from directly above, the hole occupies a relatively small area and is effectively hidden at non-zenith solar illumination resulting in an increase in mesoscale albedo.

Furthermore, cryoconite holes are often covered by an ice lid, caused by the refreezing of water that
has filled the hole during negative net radiation conditions\textsuperscript{30}. While thin frozen lids may undergo partial or complete ablation during the day, their higher albedo acts to further moderate the impact of cryoconite that they cover and render the holes indistinguishable from the adjacent ice surface (much to the dismay of many a field-campaigner with sodden feet).

**Discussion**

Our analysis reveals that the dark zone has low fractional areas of surface water (<1.0%), cryoconite holes (<0.5%) and crevasses (<0.2%). Instead, it appears that ice containing uniformly distributed impurities draped over a relatively flat surface are the primary agent responsible for the low (MOD10A1) albedo values observed during the melt-season (Fig. 5a). Near the S6 AWS, from where the UAV was launched, Stibal et al. (2017)\textsuperscript{21} report that samples of distributed impurities consist of an abundance of ice algae (Fig. 4), which are characterized by a grey/brown hue due to the brown-to-purple coloured pigments surrounding the algae chloroplasts\textsuperscript{23,32,33}. Correlations between dust content and the abundance of microbes suggest that the melt-out of particulates may provide nutrients for surface ice algae to grow\textsuperscript{21,34} and indirectly control the extent of dark zone\textsuperscript{12}. Further support for this hypothesis is provided by Tedstone et al. (2017)\textsuperscript{12} who argue that the large interannual variability in the extent of the dark zone, and its significant reduction in 2013 and 2015, demonstrates that bare ice albedo is not a consequence of summer ablation alone\textsuperscript{12}. Instead, positive correlations between the dark zone extent and proxies for the availability of liquid water and nutrients are interpreted as evidence that blooms of surface ice algae control bare ice albedo across the dark zone. Any increase in temperature and/or liquid water production in the presence of dust promotes further colonization of surface algae yielding an increase in pigmented biomass and net albedo reduction\textsuperscript{23,35}. The 12% expansion of the dark zone between 2000 and 2014 in western Greenland, corresponding with an increase in mean summer air temperature of 0.13° C per year over the same period\textsuperscript{11}, provides further support for these ongoing processes.

Whilst our results attest that the variation in the fractional area of supraglacial lakes, streams, crevasses and cryoconite do not significantly affect mesoscale albedo, they still play a secondary role in determining interannual and seasonal albedo variability. For example, supraglacial water may act to consolidate or distribute sediment and impurities across the ice sheet surface\textsuperscript{36}. Moreover, a relatively small expansion in the spatial extent of surface water would have a disproportionate impact on
mesoscale albedo and further amplify ablation. Melt rates at the base of supraglacial lakes and water bodies are double that of bare ice surfaces due to enhanced shortwave radiation absorption\textsuperscript{37, 38}. Atmospheric warming has been shown to increase the spatial extent and duration of ponded supraglacial water\textsuperscript{39, 40}. During years with higher summer temperatures, such as 2007, 2010 and 2012, supraglacial lakes formed earlier in the season and occupied a 40\% larger area than in cooler summers\textsuperscript{39}. It follows that increased storage of water in supraglacial lakes will play an important role in net albedo reduction across an expanding bare ice area in future.

Our analysis also demonstrates that crevasses reduce local (0.1 to 1 km) albedo, and hence any increase in crevasse extent will impact on mesoscale albedo patterns. Crevasses form due to localized concentration of tensile stresses which, due to high variables subglacial conditions and longitudinal stress coupling, are spatially and temporally variable across the Greenland Ice Sheet\textsuperscript{15, 41, 42}. In response to increased surface melt, GPS observations by Van de Wal et al. (2008)\textsuperscript{43} report reduced net flow over the marginal zone of the K-transect, whereas Doyle et al. (2014)\textsuperscript{44} report persistent ice flow acceleration above the equilibrium line up to 140 km from the ice sheet margin in this same sector. Recent modelling\textsuperscript{45} and observations\textsuperscript{46} of new crevasses forming over 160 km from the western margin reveals that the ice sheet interior is also more dynamically sensitive to transient stress perturbations originating from downstream than a previous steady-state model suggests\textsuperscript{47}. Regardless of ice dynamics, inland migration of the equilibrium line caused by atmospheric warming will drive increased bare ice extent, further exposing existing crevasses that were formerly snow and firn covered. Hence, surface crevasse extent will likely expand in future, resulting in mesoscale albedo reduction and enhanced surface absorption of incoming energy available for melt.

Finally, future spatial expansion of cryoconite could significantly impact surface albedo as it is so dark. Hodson et al. (2007)\textsuperscript{48} showed that 53\% of plot-scale (0.01 to 0.5 m) variation in albedo was correlated with the growth of cryoconite holes, and Chandler et al. (2015)\textsuperscript{20} report that the gradual seasonal reduction in albedo also correlates with an increase in cryoconite hole size and number. An increase in the extent of cryoconite holes may be caused by longer and warmer ablation seasons which would increase the heat energy to the walls and base of the hole, leading to further melting and hole expansion\textsuperscript{30}. On the other hand, an increase of meltwater may promote aggregation of distributed impurities and could have a surface cleaning effect, potentially raising the albedo of the surrounding
The growth and development of cryoconite holes on mesoscale albedo is hence complex and still somewhat ambiguous.

In this study, we characterized the spatial variability of surface types across the western ablating margin of the ice sheet towards the end of the melt-season when bare ice surfaces are most apparent. However, for much of the year (September to May), the ice sheet is snow-covered and it is likely that snow grain size and impurity concentration govern mesoscale albedo patterns across the K-transect and elsewhere during this period. Likewise, early melt-season albedo patterns are primarily governed by the relative proportions of snow and ice extent. Accurately determining snow melt and the timing of bare ice exposure has therefore been a priority for surface mass and energy balance models and the theoretical determinants of snow albedo and melt are relatively well established. In contrast, few studies have investigated the albedo of the bare ice surface types that characterize the ablation zone and they have commonly been treated as temporal and spatial constants in surface melt models.

The observed spatiotemporal variability in albedo across the ablation zone, has motivated a new generation of surface energy balance models that assimilate spatial patterns of albedo derived from MODIS data. Across our 25 km survey transect, the MODIS-derived surface albedo pattern is dominated by variations in the extent of uniformly distributed impurities, a result that contradicts previous research attributing it to an increased occurrence of supraglacial water. The source, processes and drivers of distributed impurities are yet to be unequivocally established, with some studies indicating a wind-blown origin, others revealing that they are derived from melt-out of englacial dust. Recent research has promoted the concept of bioalbedo, which argues that the melt-out and release of surface particulates and nutrients fertilizes pigmented ice surface algae, which drives albedo reduction over the duration of the melt-season. Further research though is required to determine how these factors combine to increase the spatial extent and concentration of pigmented surface algae, and their interaction with the availability of in situ and aeolian derived nutrients, changing atmospheric forcing and enhanced ice melt and runoff.
Methods

MODIS albedo

Albedo patterns were determined from a MOD10A1 C6 broadband (spectral range of 300 to 3000 nm) albedo product from the 8 August 2014 and available from the National Snow and Ice Data Center (NSIDC). MOD10A1 is gridded in a sinusoidal map projection and has a resolution of approximately 500 x 500 m or 0.25 km². The value of each pixel represents the best single albedo observation in the day based on cloud cover and viewing and illumination angles. We estimated that MOD10A1 has a root mean square difference (RMSD) of 7.0% in comparison to albedo measured by CNR1 or CNR4 thermopile pyranometers at the PROMICE/GAP automatic weather stations KAN-L, KAN-M and KAN-U between 2009 to 2014. This compares well to Stroeve et al. (2006) who estimated an RMSD of 6.7%.

UAV platform

Aerial imagery was acquired by a fixed-wing UAV identical to that described by Ryan et al. (2015, 2017). The UAV has a 2.1 m wingspan and is powered by a 10 Ah, 16.8 V LiPo battery pack which, with a total weight of 4 kg, yields a 1 hour endurance and 60 km range. The autonomous control system is based around an Arduino navigation and flight computer updated in real-time by a 10 Hz data stream comprising of a GPS, magnetometer, barometer and accelerometer. These data are logged along with a timestamp for each activation of the digital camera shutter which automatically triggers when a horizontal displacement threshold is exceeded. The UAV was hand launched on 8 August 2014 from a base camp at 67.08° N, 49.40° W, located at the site of the Institute for Marine and Atmospheric Research (IMAU), University of Utrecht S6 automatic weather station. It was pre-programmed to carry out a 25 km survey across the Kangerlussuaq sector of the western Greenland Ice Sheet (Fig. 1). The Greenland Ice Mapping Project (GIMP) digital elevation model (DEM) was used during the selection of 3D waypoints to ensure the UAV maintained a constant altitude of 350 m above the surface during the autonomous sorties. On return from the sortie, the UAV was manually landed into a 10 x 5 m net.

Digital imagery

Digital imagery was acquired by a Sony NEX-5N digital camera vertically mounted inside the front of the airframe. The camera has a 16 mm fixed focus lens (53.1 by 73.7° field of view) yielding an image
footprint of approximately 525 x 350 m during the autonomous sortie. The width of each image is approximately similar to the pixel footprint of MODIS. The camera was preset with a fixed shutter speed of 1/1000 s, ISO 100 and F-stop of 8, and triggered every 35 m to provide a 90% forward image overlap. The relatively fast shutter speed minimizes image blur whilst the low ISO and F-number ensures maximum image quality where even the brightest surfaces do not saturate the image. The camera was set to record the images in RAW format, an image format which contains minimally processed data from the camera’s sensor. During the survey, ~2000 RAW images were acquired at the camera’s maximum (4912 x 3264 pixels) resolution which, once the images were corrected for barrel, or geometric, distortion, equates to a ground sampling distance of ~11 cm.

**Orthomosaic and DEM generation**

The RGB images were used to produce an orthomosaic and DEM using Agisoft PhotoScan Pro (http://www.agisoft.com/) following the processing sequence described by Ryan et al. (2015)\(^60\). The images were georeferenced by providing latitude, longitude and altitude data recorded by the flight controller. The orthomosaic was produced in the software’s “mosaic” mode, meaning that pixels in the centre of the images were preferentially used to provide the output pixel value. The orthomosaic and DEM were nearest neighbour resampled to a ground resolution of 15 cm and 50 cm, respectively. We divided the orthomosaic into sixty 0.25 km\(^2\) segments and each segment was assigned a MOD10A1 value.

**Surface classification**

The fractional area of each surface type was calculated by dividing the number of pixels of each surface type by the total number of pixels in each orthomosaic segment. The number of pixels of each surface type was estimated using a supervised k-Nearest Neighbours (k-NN) classification from the scikit-learn Python module\(^63\). The pixels were classified using a majority vote based on the Euclidean distance to five equally weighted nearest neighbours (Fig. 7). The k-NN was manually trained with seven distinct and visually identified surfaces found in the orthomosaic: i) clean ice, ii) ice containing uniformly distributed impurities, iii) deep water, iv) shallow water, v) cryoconite either in holes or fluvial deposits, vi) crevasses and vii) snow. The training samples of the surface types were manually digitized from ten orthomosaic segments based on RGB brightness and a layer that specified whether or not the pixel was situated within a crevasse or fracture. This roughness layer was determined by
calculating the residual between the original 50 cm DEM subtracted from a 30 m Gaussian-smoothed DEM. Negative anomalies with a vertical displacement greater than 1 m were identified as crevasses. Small cracks and fractures were detected on the basis of sharp RGB contrast, and were discriminated using an edge detector algorithm62. Pixels within 2 m of a linear feature were also identified as crevasses.

The efficacy of the k-NN classifier was evaluated by comparison with independently digitized surface types in three orthomosaic segments, at the centre and extreme ends of the transect. We found that 92% of the pixels were classified accurately. The k-NN classified crevasses with an accuracy of 88% and performed better for deep and shallow water (96%) than for cryoconite, ice containing distributed impurities, clean ice and snow (90%). The classification of supraglacial water is relatively accurate because water has low reflectance in the red band and forms a unique cluster in the feature space (Fig. 7). The performance of the k-NN for clean ice and ice containing distributed impurities is less accurate (90%) because these surfaces reflect the RGB visible bands in similar relative proportions and the surface classes overlap in the feature space (Fig. 7). Misclassification could also be caused by shadows, especially in segments with steep topography (e.g. 6b). Shadows increase the proportions of the surface classified as ice containing distributed impurities and/or cryoconite. Accounting for shadows would reduce the proportion of ice containing distributed impurities in the crevassed zone (Fig. 6b) and subsequently increase their variation across the transect. This would make the impact of distributed impurities on the mesoscale albedo variability, and the conclusions of this study, more significant.

Finally, we used PCR to explore the dominant modes of surface type variation along the transect and assess which surface type, or set of surface types, best correlate with mesoscale albedo variability, as represented by MOD10A1. Linear regression was used to yield the correlation coefficients between the principle component scores and MOD10A1.

**Estimating albedo of digital image pixels**

An estimate for the albedo ($\alpha$) of each surface type was obtained from the UAV digital imagery. An explicit description of this method can be found in Ryan et al. (2017)14 but we briefly summarize it here. Firstly, the RAW digital numbers of each proprietary Sony RAW image were preserved by converting to a 16-bit TIFF image using dcraw (http://cybercom.net/~dcoffin/dcraw/). A vignette correction mask was universally applied to compensate for image and lens distortion due to edge
effects which were as high as 17.6% at the corners of some images. The correction mask was calculated from the mean vignette of all images acquired at nadir during the survey period. Barrel distortion was corrected using ImageMagick (http://www.imagemagick.org/) which utilised the coefficients stored in the image’s ancillary metadata also known as Exchangeable Image File Format (EXIF) data.

We then corrected the images for changing illumination conditions during the survey using downward irradiance measured by a ground-based upward facing Apogee SP-110 pyranometer. To do this, images of a 25 x 25 cm Teflon white reference target were acquired every 10 minutes using the UAV digital camera from the ground. The relationship between the mean RGB DNs of the white reference target and the downward irradiance recorded by the upward facing pyranometer were used to construct a calibration curve using a linear least squares regression ($R^2 = 0.96$). The ratio of reflected radiation recorded by the camera and the downward radiation estimated from the calibration curve enabled the illumination-corrected reflectance of each pixel to be defined.

Since snow and ice are non-Lambertian surfaces, a nadir measurement of reflectance underestimates albedo by between 1 and 5% in the visible band\textsuperscript{64}. The illumination-corrected images were therefore calibrated again by multiplying the image pixel numbers by a factor calculated by dividing the mean pixel value of the illumination-corrected image by the albedo recorded by upward and downward facing Apogee SP-110 pyranometers mounted on the UAV. Ryan et al. (2017)\textsuperscript{14} found that albedo determined using this method has an accuracy of ±5% over ice sheet surfaces typically found in the ablation zone.

**Data Availability**

The MODIS albedo data are available from the National Snow and Ice Data Center (NSIDC) at http://nsidc.org/data/MOD10A1 and https://nsidc.org/data/MYD10A1. The UAV imagery is available from the corresponding author on reasonable request.

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Author contributions statement

JR and AH designed, tested and built the UAVs. JB provided UAV on-board pyranometers and data logger. JR and JB programmed and deployed the UAV. JR analysed the data and wrote the original manuscript. AH provided direction/supervision and co-wrote the manuscript. AH and JB were project PIs. KC provided the ground photographs used for qualitative assessment of surface types. AH, MS, JB, JC, TIF, LCS and KC provided conceptual and technical advice and contributed to data interpretation. All authors commented and critically revised the manuscript. The authors declare no competing financial interests.

References


Figures

Figure 1. The UAV survey transect overlaid onto a Landsat 8 Operational Land Imager (OLI) true colour image of the Kangerlussuaq sector of the Greenland Ice Sheet from 6 August 2014. The transect was divided into sixty 0.25 km² segments for comparison with the MODIS albedo product, MOD10A1. High resolution aerial imagery and classifications of six segments are shown in figures 5 and 6 (a to c).
Figure 2. Schematic summarizing the aims of the study. Aerial digital imagery are used to characterize the surface types that are found in the ablation zone and assess their impact on mesoscale spatial albedo patterns as represented by MODIS.
Figure 3. Variation of the fractional area of each surface type, derived from UAV imagery, in comparison to MOD10A1 on 8 August 2014. The x-axis is displayed in figure 1. The results of the classification for six segments are shown in figures 5 and 6 (a to c).
Figure 4. Photograph showing close-up of bare ice found in the ablation zone, near the field camp located close to the S6 automated weather station at ~1,000 m a.s.l. (Fig. 1). (a) Ice containing distributed impurities on the surface and (b) clean ice with cryoconite holes.
Figure 5. RGB digital image, albedo map derived from digital image and classification of noteworthy and exhibitive segments whose locations are shown in figure 1. (a) Segment characterized by mostly ice containing uniformly distributed impurities. (b) Segments characterized by similar ice surface to (a) but with a larger fraction of channelized surface meltwater. (c) Segment dominated by a supraglacial lake with a previous shore consisting of clean ice.
Figure 6. RGB digital image, albedo map derived from digital image and classification of noteworthy and exhibitive segments whose locations are shown in figure 1. (a) Segment containing a high fraction of crevasses. (b) Segment characterized by a much lower relief surface and no crevasses. (c) Segment characterized by clean ice and numerous cryoconite holes.
Figure 7. Plots showing the attributes of the seven surface types identified in this study. DN is digital number of the Sony NEX-5N calibrated RAW image.
Chapter 5

UAV photogrammetry and structure from motion to assess calving dynamics at Store Glacier, a large outlet draining the Greenland ice sheet

5.1 Summary

Chapter 5 addresses the second theme of the thesis: tidewater glaciers outlet dynamics, and aims to establish whether UAVs are useful for investigating processes of ice sheet mass loss at the ice-ocean boundary. In the summer of 2013, we carried out three UAV surveys over Store Glacier, a major tidewater outlet glacier draining the western sector of the Greenland Ice Sheet. The paper describes the UAV platform, flight planning, and the generation of accurate, georeferenced orthomosaics and digital elevation models (DEMs) from the overlapping UAV imagery. We then describe methods to analyse individual and pairs of DEMs to investigate crevasse patterns, surface velocities, calving processes, and front positions at a daily and seasonal timescale. It is concluded that the use of repeat UAV surveys coupled with the processing techniques outlined
in this paper may be useful for resolving the complex frontal dynamics that characterize large calving outlet glaciers.

5.2 Contribution

To complete this study, JR planned the surveys and piloted the UAV during the summer of 2013 and obtained imagery over Store Glacier in July and August. JR developed the method to produce accurate orthomosaic and DEM products from the overlapping UAV imagery. JR developed the methods to derive surface velocities, surface elevation change, crevasses patterns and calving dynamics from the UAV products. JR wrote the first draft of the manuscript and edited it in response to suggestions from co-authors and two reviewers. JR made all the figures.
UAV photogrammetry and structure from motion to assess calving dynamics at Store Glacier, a large outlet draining the Greenland ice sheet

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Abstract. This study presents the application of a cost-effective, unmanned aerial vehicle (UAV) to investigate calving dynamics at a major marine-terminating outlet glacier draining the western sector of the Greenland ice sheet. The UAV was flown over Store Glacier on three sorties during summer 2013 and acquired over 2000 overlapping, geotagged images of the calving front at an ∼40 cm ground sampling distance. Stereo-photogrammetry applied to these images enabled the extraction of high-resolution digital elevation models (DEMs) with vertical accuracies of ±1.9 m which were used to quantify glaciological processes from early July to late August 2013. The central zone of the calving front advanced by ∼500 m, whilst the lateral margins remained stable. The orientation of crevasses and the surface velocity field derived from feature tracking indicates that lateral drag is the primary resistive force and that ice flow varies across the calving front from 2.5 m d⁻¹ at the margins to in excess of 16 m d⁻¹ at the centreline. Ice flux through the calving front is 3.8 × 10⁷ m³ d⁻¹, equivalent to 13.9 Gt a⁻¹ and comparable to flux-gate estimates of Store Glacier’s annual discharge. Water-filled crevasses were present throughout the observation period but covered a limited area of between 0.025 and 0.24 % of the terminus and did not appear to exert any significant control over fracture or calving. We conclude that the use of repeat UAV surveys coupled with the processing techniques outlined in this paper have great potential for elucidating the complex frontal dynamics that characterise large calving outlet glaciers.

1 Introduction

Observational and modelling studies have demonstrated that Greenland’s marine outlet glaciers have a complex and potentially non-linear response to both environmental forcing (e.g. Vieli et al., 2000; Benn et al., 2007; Holland et al., 2008; Howat et al., 2010; Hubbard, 2011; Joughin et al., 2012; Walter et al., 2012; Carr et al., 2013) and to changes in front position (Howat et al., 2007; Luckman et al., 2006; Joughin et al., 2008). To quantify these processes and feedbacks, regular and accurate high-resolution measurements are required to capture the key spatio-temporal linkages between rates of ice calving, flow, surface lowering and frontal advance/retreat. Despite significant advances in satellite remote sensing, limitations of spatial resolution (e.g. MODIS) and/or frequency of repeat imagery (e.g. Landsat or TerraSar-X) render detailed, day-to-day analysis of calving-front dynamics unfeasible. On the other hand, acquisition of digital imagery from unmanned aerial vehicles (UAVs) combined with the development of stereo-photogrammetry software has enabled the provision of high-resolution 3-D georeferenced data on demand for geoscience applications (e.g. d’Oleire-Olmanns et al., 2012; Hugenholtz et al., 2012, 2013; Whitehead et al., 2013; Lucieer et al., 2014). This represents a cost-effective technique for acquiring aerial data in remote, hazardous and/or inaccessible regions and recent applications for emerging snow and ice investigation abound the web (e.g. see the highly informative site of Matt
Table 1. Attributes of the flight surveys and image acquisition of the UAV.

<table>
<thead>
<tr>
<th>Flight no.</th>
<th>Date</th>
<th>Interval between pictures (s)</th>
<th>No. images</th>
<th>Glacier coverage (km²)</th>
<th>Resolution of DEM (cm/pixel)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>1 July</td>
<td>1.55</td>
<td>611</td>
<td>3.17</td>
<td>40</td>
</tr>
<tr>
<td>2</td>
<td>2 July</td>
<td>1.51</td>
<td>1051</td>
<td>4.95</td>
<td>38</td>
</tr>
<tr>
<td>3</td>
<td>23 August</td>
<td>2.36</td>
<td>567</td>
<td>5.02</td>
<td>39</td>
</tr>
</tbody>
</table>

Nolan; http://www.drmattnolan.org/photography/2013/). To date, published (peer-reviewed) application appears to be limited to the investigation of inter-annual changes of a land-terminating glacier on Bylot Island, Canadian Arctic (Whitehead et al., 2013).

Between July and August 2013, an off-the-shelf, fixed wing UAV equipped with a compact digital camera flew three sorties over the calving front of Store Glacier, West Greenland. The aerial photographs obtained during these flights were used to produce high-resolution (∼40 cm; Table 1) digital elevation models (DEMs) and orthophotos of the glacier terminus. These data allowed for the investigation of the spatially complex and time-varying glaciological processes operating at the glacier’s calving front. The aim of this paper is to

1. detail the UAV, in terms of its payload and camera settings, and its specific deployment to Store Glacier;
2. describe the techniques used for processing the aerial images and quantifying glaciological processes;
3. discuss the significance of the data we obtained which includes calving events, the character, orientation and morphology of crevasses, surface velocities, ice discharge and changes in thickness and position of the calving front.

2 Data and methods

2.1 Study site

Store Glacier is a large marine-terminating (tidewater) outlet glacier located in the Uummannaq district of West Greenland (Fig. 1). The calving front has a width of 5.3 km and an aerial calving front (freeboard) of up to 110 m a.s.l. (Ahn and Box, 2010). Aerial photography from 1948 onwards reveals that Store Glacier’s frontal position has remained stable over the last 65 years (Weidick, 1995). Seasonally, the calving front exhibits advance and retreat of up to 400 m (Howat et al., 2010). The study here focuses specifically on glacier dynamics during the melt season under open-water, tidal modulation of ice flow.

2.2 UAV platform

The UAV airframe is an off-the-self Skywalker X8 (www.hobbyking.com) which has a wingspan of 2.12 m and is made from expanded polypropylene (EPP) foam (Fig. 2). For this deployment, the X8 was powered by two 5Ah four-cell (14.8 V) lithium polymer batteries driving a 910 W brushless electric motor turning an 11 × 7 foldable propeller. In this configuration, the X8 has a flying mass of ∼3 kg (including 1 kg payload), which allows for a cruising speed of around 55–70 km per hour with a maximum range of ∼60 km in benign conditions at constant altitude. A small propeller/high-revolution motor combination was chosen to provide maximum instantaneous thrust to ensure a clean launch (for novice operators) and to handle the potentially strong katabatic winds encountered during its 40 km sortie.

The autopilot is an open-source project called Ardupilot (http://ardupilot.com/) based on an Atmel 2560 8 bit microcontroller and standard radio control parts including 2.4 GHz
results presented here are limited to the former. A SPOT GPS tracking device was also included in the payload to facilitate recovery should a mission fail (which it did). The focal length of the Lumix lens was adjusted to 5.1 mm (35 mm equivalent) to allow the widest possible coverage which gave the camera a 73.7° horizontal and 53.1° vertical field of view. A short exposure time of 1/1600 and a focal ratio of 8 were chosen to prevent overexposure and blurring of the ice surface. The Ardupilot open-source code was amended to trigger the camera automatically at user defined time or distance intervals at or between certain waypoints. The cameras were mounted pointing downwards within the airframe using neoprene and velcro straps to dampen vibration in a custom recessed aperture cut in the bottom with a UV filter to protect the lens and seal it.

2.3 Flight planning

The open-source software APM Mission Planner (http://plane.ardupilot.com/) was used for flight waypoint manipulation and planning in conjunction with the 30 m Greenland Mapping Project (GIMP) DEM (Howat et al., 2014). To optimise spatial coverage against required resolution, flight endurance and stability, the UAV was programmed to fly at a constant altitude of 500 m a.s.l. (Fig. 1). Based on the camera’s focal length and field of view (53.1° by 73.7°), the ground (sea-level) footprint at 500 m a.s.l. for each photo was ∼450 × 750 m. To ensure coverage of the entire glacier terminus and overlap for successful photogrammetric processing, the four transects broadly parallel to the calving front were flown with ∼250 m separation yielding a side overlap between photos of 70% (Fig. 1). The mean ground speed of the UAV was ∼70 km h−1 and the camera trigger interval was adjusted between surveys. On flights 1 and 2, the interval between camera triggers was 1.5 s corresponding to a forward overlap of 94% and over 1000 geotagged images acquired. Flight 3 had a 2.4 s interval yielding a 90% forward overlap and 581 images (Table 1).

UAV operations were based out of a field camp with the advantage of a 50 m area of flat alluvial terrace with relatively boulder and bedrock free ground for manual remote-control take-off and landing. This location did, however, require a ∼10 km transit to the calving front over a 450 m high peninsula which significantly reduced the useful endurance over the target. Of the six sorties flown over outlet glaciers in the region during July and August 2013, the three over Store Glacier were most successful. Each sortie was 40 km long and ∼35 min in duration after the UAV had attained its operating altitude at the start of the mission and was passed from manual remote-control mode into autopilot mode (Fig. 1). Visual and remote-control contact is lost within a few kilometres of the UAV being placed in autopilot mode; hence, validation of the mission plan is essential.

Two lightweight digital cameras were tested at the field site: a Panasonic Lumix DMC-LX5 10.1 megapixel (MP) camera with a 24 mm wide-angle zoom lens and a 16.1 MP Sony NEX-5N with a 16 mm fixed focal length lens though radio control and pulse-width modulation (PWM) controlled servos for aileron and elevon control (Fig. 2). Ardupilot implements a dual-level proportional-integral-derivative (PID) controller architecture. The lower level controls flight stabilisation and the higher level controls based navigation. Tuning of the PID parameters is necessary to suit the mass and dynamics of the airframe to ensure accurate stabilisation without pitch/roll oscillation (lower-level controller) or flight-path weaving (higher-level controller). The autopilot allows the UAV to fly autonomously according to a pre-programmed flight path defined by a series of 3-D waypoints chosen by the user. The autopilot utilises a GPS for navigation, a triple axis accelerometer and gyro for stabilisation and a barometric pressure sensor for altitude control. These parameters are logged to memory at 10 Hz throughout the flight (Fig. 2).

The advantage of this package is that it can be assembled within a day from off-the-shelf parts and is cost effective at less than USD 2000. The X8 is also relatively straightforward to fly, robust, easily repairable and floats, all added bonuses when being deployed in remote areas by potential novices. Furthermore, the Ardupilot firmware is open source and hence can be programmed for specific requirements, for example camera triggering (see below).

Two lightweight digital cameras were tested at the field site: a Panasonic Lumix DMC-LX5 10.1 megapixel (MP) camera with a 24 mm wide-angle zoom lens and a 16.1 MP Sony NEX-5N with a 16 mm fixed focal length lens though

Figure 2. Flowchart of the control set-up and picture of the UAV at base camp with the relative novices.
Figure 3. (a) Surface elevation difference between two DEMs collected on 1 July and 2 July. Red areas show elevation loss whilst blue areas show elevation gain. White circles highlight the calving events that occurred between the two UAV surveys. (b) The position of the calving front of Store Glacier during the summer of 2013. (c) Calving-front retreat observed between 12 June and 1 July. Inset is an orthorectified image of the water-filled crevasses observed on 1 July with a pixel resolution of 30 cm. (d) Calving-front advance observed between 1 July and 23 August. Inset is an orthorectified image showing water-filled crevasses observed on 23 August. The coverage and size of water-filled crevasses is smaller.

2.4 Three-dimensional model generation

Three-dimensional data were extracted from the aerial photos using Agisoft PhotoScan Pro software (Agisoft LLC, 2013). This software’s strength lies in its ability to fully automate workflow and enables non-specialists to process aerial images and produce 3-D models which can be exported as georeferenced orthophotos and DEMs (e.g. Figs. 3 and 6). The first stage of processing is image alignment using the structure-from-motion (SfM) technique. SfM allows for the reconstruction of 3-D geometry and camera position from a sequence of two-dimensional images captured from multiple viewpoints (Ullman, 1979). PhotoScan implements SfM algorithms to monitor the movement of features through a sequence of multiple images and is used to estimate the location of high-contrast features (e.g. edges), obtain the relative location of the acquisition positions and produce a sparse 3-D point cloud of those features. The Ardupilot flight logs of the onboard navigation sensors allow the camera positions and the 3-D point cloud to be georeferenced within instrument precision. SfM also enables the camera calibration parameters (e.g. focal length and distortion coefficients) to be automatically refined; hence, there is no need to pre-calibrate the cameras and lens optics (Verhoeven, 2011).

Once the photos have been aligned, a multi-view reconstruction algorithm is applied to produce a 3-D polygon mesh which operates on pixel values rather than features and enables the fine details of the 3-D geometry to be constructed (Verhoeven, 2011). The user determines the precision of the final 3-D model based on image resolution and pixel footprint. A medium quality setting was chosen yielding DEMs with between 38 and 40 cm/pixel ground sampling resolution (GSD), which were resampled to a Cartesian 50 cm grid to enable intercomparison (Table 1). Higher resolutions (< 30 cm GSD) are attainable but the increase in computational time and the accuracy of georeferencing limits the benefits of such apparent precision.

Two problems of accuracy were encountered in DEM production: (1) PhotoScan failed to reconstruct a flat sea level of constant elevation, and (2) relative positional errors between the DEMs constructed from different sorties were up to 17.12 m horizontally and 11.38 m vertically. Positional errors were due to the specified limits of the onboard L1 GPS of ± 5.0 m horizontally and, when combined with the barometric sensor, to a similar accuracy vertically. These were compounded by the time lag between the camera triggering and actual photograph acquisition. Hence, a secondary stage of processing was carried out which involved 3-D coregistration of the DEMs. To do this, the horizontal and vertical coordinates of common control points (CPs) based on distinct features such as cliff bases, large boulders and promontories were extracted from the georeferenced orthoimages. The CPs that were at sea level were nominally given elevation values of zero, re-imported into PhotoScan and subsequently reprocessed along with a geodetic GPS ground CP located at 70.401° N, −50.665° E and 335.85 m altitude on the bedrock headland overlooking the glacier’s northern flank. During this secondary stage of processing, PhotoScan’s optimisation procedure was run to correct for possible distortions. After processing with the CPs, a flat sea level across the glacier front was produced and the relative errors between the three DEMs were reduced to ± 1.41 m horizontally and ± 1.90 m vertically. The georeferenced 3-D DEMs and orthophotos were then exported at 50 cm pixel size for further analysis in ArcGIS and ENVI software packages.
2.5 Analysis

Changes in calving-front positions were obtained from these data combined with a Landsat 8 panchromatic image obtained on 12 June (Fig. 3b). Each calving-front position was digitised according to the procedure outlined by Moon and Joughin (2008), whereby a polygon of the calving-front retreat or advance is digitised and divided by the width of the glacier. This method has been used in previous studies (e.g. Howat et al., 2010; Schild and Hamilton, 2013) and enables intercomparison of results. Surface elevation change was calculated from the residual difference of the DEMs (Fig. 3a).

Ice flow across the terminus region was calculated by feature tracking performed on successive DEMs using the ENVI Cosi-CORR software module (Fig. 4b). These velocities were then used to estimate ice flux through the calving front for the same period under the assumption of plug flow (uniform velocity profile with depth) and using a calving-front cross section obtained from Xu et al. (2013) and modified by single and multi-beam echo sounder bathymetry obtained by S/V Gambo in 2010 and 2012 (Chauché, unpublished). The frontal cross section was divided into 10 m vertical strips and under the plug-flow assumption, each was assigned its corresponding horizontal velocity (Fig. 4a). The floatation depth and buoyancy ratio across the calving front was calculated using the ice surface (freeboard) elevation and total ice thickness with a value for the density of ice of 917 kg m$^{-3}$ and for sea water of 1028 kg m$^{-3}$ (Fig. 5a).

To investigate the distribution and patterns of crevassing, each DEM was Gaussian filtered at 200 pixels (100 m) in ArcGIS and subtracted from the original DEM to yield the pattern of negative surface anomalies. These anomalies were converted into polygons to map and hence quantify crevasse distribution and character (Fig. 6a). The resulting polygons were enclosed by a minimum bounding rectangle, which allowed the orientation, width, length and depth of crevasses to be extracted (Fig. 6a, Table 2). Water-filled crevasses were automatically located in the ENVI package using the supervised maximum likelihood classification (MLC) method. Representative training samples for water-filled areas were chosen from the colour composite orthophoto (Fig. 6b). The trained tool then classifies pixels that are interpreted as water into the desired class. The resulting raster image was converted into a shapefile and used to mask and define the area of the water-filled crevasses across the terminus. These procedures allow thousands of crevasses in multiple orthoimages and DEMs to be quantified easily without the difficulties and dangers associated with direct field measurements.

2.6 Uncertainties and limitations

The relative horizontal uncertainties between the DEMs were investigated by feature tracking the stationary bedrock at the sides of the glacier. The root mean square (rms) horizontal displacement was ±1.41 m which provides us with an approximate error estimate. The relative vertical uncertainties between the DEMs were estimated by calculating elevation differences between bedrock areas, which revealed an error estimate of ±1.9 m. The two-stage procedure outlined in Sect. 2.4 therefore enabled us to improve the relative positional uncertainties from nearly 20 m to less than 2 m. For future studies, it is thought that several CPs on the bedrock either side of the glacier front would further reduce these uncertainties. A telemetric differential GPS deployed on or near the calving front, which is sufficiently large/bright to identify within the aerial imagery would allow further ground control in the centre of DEMs, away from bedrock CPs.

Due to the lack of reflected light from deep crevasse recesses, the DEM generation process cannot quantify the narrowest sections of all fractures and resultant crevasse depths are therefore a minimum estimate. The technique is also clearly limited to line of sight precluding narrow fractures which extend for tens of centimetres horizontally and potentially up to a few metres vertically (Hambrey and Lawson, 2000; Mottram and Benn, 2009).

Finally, there are a number of practical difficulties when operating an autonomous aircraft in remote and inaccessible environments. Mission planning is critical; knowledge of the local weather conditions, as well as up-to-date satellite imagery and DEMs are a prerequisite.
Table 2. Attributes of mean crevasse width, length and orientation in each zone labelled in Fig. 5. Orientations are measured along the long axis of each crevasse with respect to the direction of flow which is 0°.

<table>
<thead>
<tr>
<th>Zone</th>
<th>Mean width (m)</th>
<th>Mean length (m)</th>
<th>Mean orientation (°)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Zone 1</td>
<td>3.6</td>
<td>9.4</td>
<td>9.2</td>
</tr>
<tr>
<td>Zone 2</td>
<td>4.8</td>
<td>14.0</td>
<td>36.7</td>
</tr>
<tr>
<td>Zone 3</td>
<td>10.5</td>
<td>32.6</td>
<td>85.1</td>
</tr>
<tr>
<td>Zone 4</td>
<td>6.5</td>
<td>17.8</td>
<td>60.4</td>
</tr>
<tr>
<td>Zone 5</td>
<td>3.5</td>
<td>8.5</td>
<td>10.8</td>
</tr>
</tbody>
</table>

3 Results

Three successful UAV sorties were flown over Store Glacier calving front providing imagery, orthophotos and DEMs on 1 and 2 July and the 23 August, herein referred to as flights and associated products 1 to 3, respectively (Table 1). The interval between flights 1 and 2 was 19 hours and comparison between these outputs enables identification of processes operating over a daily (short) timescale, be it a very specific snapshot. The third sortie was flown 52 days later and comparison between these outputs enables investigation of late-seasonal change. The footprint of the four cross-glacier transects flown extends just over 1 km upstream from the calving front and herein is referred to as the terminus.

3.1 Short timescale calving and surface elevation change

Residual elevation change between 1 and 2 July (Fig. 3a) reveals that the front retreated in two sections by up to 50 and 80 m, respectively. The more northerly calving event (A) resulted in a 450 m wide section of the terminus retreating by between 20 and 50 m, whilst event B produced between 20 and 80 m of retreat across a 400 m section (Fig. 3a). In addition to these two calving events (which are discussed in section 3.6), the central 4.5 km frontal section advanced between 12 to 16 m (Fig. 3a). At its lateral margins, the calving front shows no discernible systematic change though there are isolated, small calving events, for example, within 50 m of the southern flank (Fig. 3a). Upstream of the calving front, there is no net change in mean surface elevation away from the front and the dappled pattern of residual elevation change is a result of the advection of crevasses and seracs. Successive long profiles of the terminus between the 1 and 2 July reveal specific down-glacier crevasse advection with flow (Fig. 6) at a rate of 5 and 16 m d⁻¹ on profile 1 and 2, respectively. These results provide corroboration for the surface velocities derived by feature tracking in Sect. 3.4.

3.2 Seasonal timescale calving-front position and surface elevation change

Over the entire melt season, larger fluctuations in calving-front position are observed (Fig. 3b). Over the 19-day period from 12 June to 1 July, mean frontal retreat was 160 m (Fig. 3c) and between 2 July and 23 August, the calving front advanced by an average of ∼110 m to a position similar to that in 12 June (Fig. 3d). These mean values, however, do not convey the full extent and detail of the changes observed in the calving front. For example, the central section of the calving front retreated by up to 525 m between the 12 June and 1 July and advanced by up to 450 m between 2 July and 23 August (Fig. 3b). Furthermore, the lateral margins of Store Glacier (the southern 850 m and the northern 1.5 km) are relatively stable with <50 m change in position. Over the 52-day period between 2 July (Flight 2) and 23 August (Flight 3) widespread surface lowering of 6.1 m (or 0.12 m d⁻¹) was observed across Store Glacier terminus (Fig. 4a), which is significantly larger than the estimated vertical uncertainties of the DEMs (∼1.9 m). Despite the same dappled patterns caused by local advection of crevasses and seracs, we infer this to be associated with dynamic thinning 1 km upstream of the calving front, which is discussed in Sect. 4.2.

3.3 Bathymetry

The deepest sector of the calving front is located 1 km south of the centreline and exceeds 540 m below sea level (Fig. 5a). This 200 m wide sector also corresponds to the greatest thickness of ∼600 m. To the south of this deepest point, the bottom rises rapidly to a 200 m deep shelf located 500 m from the flank. To the north of the deepest point, the bottom shallows more gently to within 400 m where it becomes steeper towards the fjord wall.

3.4 Surface velocities

Maximum surface-flow velocities of 16 m d⁻¹ between 1 and 2 July are consistent with results obtained in previous studies using other techniques, such as feature tracking images from a land-based time-lapse camera (between 11 and 15 m d⁻¹) (Ahn and Box, 2010; Walter et al., 2012). The spatial pattern of surface flow from feature tracking of images between the 1 and 2 July varies considerably across the terminus of Store Glacier (Fig. 4b) attaining velocities of 16 m d⁻¹ (5.8 km a⁻¹) near the centre of the glacier down to 2.5 m d⁻¹ at the lateral flanks. Surface velocities are related to slope, depth, thickness and distance from the lateral margins (Fig. 5c, d). As would be expected, maximum velocities correlate with maximum depth and towards the northern flank are linearly correlated (R² = 0.90) with frontal depth (Fig. 5c). Towards the southern flank the relationship is less apparent especially between 200 and 350 m depths. There
The calving front of Store Glacier was 3.8 \times 10^7 \text{ m}^3 \text{ d}^{-1}. Bathymetric data we estimated that the mass flux through the flow pattern is combined with frontal toe was resting on the fjord bed (Chauché, unpublished). (Fig. 5a), side-scan sonar observations reveal that the glacier \sim 13.9 \text{ Gt} \text{ a}^{-1} due to lent to crevasses are up to 30 m deep, over 500 m long and nearly of 18 m, lengths of 68 m and widths of 31 m. The largest crevasses are up to 30 m deep, over 500 m long and nearly 200 m wide but no crevasses that penetrated below sea level were identified. Most crevasses in this region are arcuate with limbs pointing towards the calving front and are orientated obliquely to the direction of ice flow (Fig. 6). This arcuate morphology of crevasses continues across the central 3 km of the terminus in zone 3 (Fig. 6). Here, crevasses have mean a depth of 10.5 m, length of 50 m and widths of 18 m (Table 2).

In zone 2, 300 to 500 m from the northern flank, crevasses are aligned obliquely to the direction of ice flow (30–45\degree). Up to the fjord walls in zones 1 and 5, crevasses are generally orientated parallel to the ice flow (> 15\degree) (Fig. 6, Table 2) and are much smaller with a mean lengths of 22 m and width of 8 m (Table 2). No discernible difference in average crevasse depths, lengths or widths was observed between early July and late August and the pattern and character of crevassing was also similar.

Water-filled crevasses were clustered in zone 4, coinciding with the sector of larger crevasses (Fig. 6b). Water-filled crevasses covered 12,000 \text{ m}^2 or 0.24% of the survey area (to \sim 1 \text{ km from the calving front}) on 2 July (Table 1). Some 42 individual water-filled crevasses were identified with the largest having an area of 1200 \text{ m}^2. By 23 August, the number, size and total area of water-filled crevasses were lower: only 10 water-filled crevasses could be identified, the largest of which was 400 \text{ m}^2 and with a total area of 1230 \text{ m}^2 (0.025% of the survey area). We were not able to ascertain the depth of water in the crevasses as no common crevasses could be identified which drained or filled between flights but this would be a specific aim of future studies which, with regular sorties, could potentially determine the depth of a crevasse before filling or after drainage or otherwise exploit the light reflectance relationship with water depth (e.g. Fitzpatrick et al., 2014).

Successive profiles of the terminus from 1 and 2 July demonstrate how the UAV surveys are capable of capturing the displacement of crevasses, which advect downstream at a rate of 5 and 16 \text{ m} \text{ d}^{-1} in profiles 1 and 2, respectively (Fig. 6). The techniques used in this study are therefore capable of identifying changes in crevasses geometry, particularly width and depth through time.

### 3.5 Crevassing

The morphology and orientation of crevasses varies markedly across the terminus (Fig. 6). The largest crevasses occur in a sector south of the glacier centre line in zone 4 (Fig. 6, Table 2). Here, crevasses have mean minimum depths of 18 m, lengths of 68 m and widths of 31 m. The largest crevasses are up to 30 m deep, over 500 m long and nearly

![Figure 5. (a) Profiles showing the sea floor bathymetry and ice surface elevation at the calving front. These data were combined with surface velocities to estimate the ice flux of Store Glacier. Where the floatation percentage is over 100%, it is assumed that the ice is not thick enough to be fully grounded in hydrostatic equilibrium. (b) The relationship between effective basal shear stress and velocity. (c) The relationship between depth and velocity. At depths deeper than 400 m, velocities are fairly constant. The two differing relationships between 150 and 350 m represent velocities from different sides of the glacier. (d) Relationship between velocity and distance from the lateral margins. The positive correlation demonstrates the importance of the resistance provided by the fjord walls.](image-url)
crevasses that were deeper and located nearer to the front, yet did not calve. Water was not observed in any of the crevasses along which calving took place.

4 Discussion

4.1 Changes occurring over a daily timescale

The orientation of crevasses suggests that lateral drag is an important resistive stress on Store Glacier. The lateral margins of Store Glacier are characterised by crevasses that are orientated parallel to the direction of flow which suggests that they have formed in response to simple shear stresses associated with the drag of the fjord walls (Fig. 6) (Benn and Evans, 2010). The importance of lateral drag is further demonstrated by the morphology of crevasses found near the glacier flow line (Fig. 6). Their arcuate nature indicates that the principal tensile stresses operating on the ice have been rotated by lateral gradients in ice velocity. These gradients are caused by the simple shear stress between the fjord walls and the margins of the glacier which cause the ice to flow slower (Fig. 4b) (Benn and Evans, 2010).

The simple shearing caused by velocity gradients is further demonstrated by the differing relationship between velocity and depths between the north and south side of the glacier (Fig. 5c, d). On the north side, the velocity increases gradually from the fjord wall to the centre of the glacier, reflecting the gradual deepening of bathymetry and the resulting decrease of basal and lateral drag. On the south side, the velocities are higher than the north side for given depths and distances from the lateral margins (Fig. 5c, d). We hypothesise that, because the deepest part of the glacier is situated 1 km south of the centreline, the ice on south side is more influenced by faster flowing ice which exerts a simple shear stress on the shallower, adjacent ice (250–400 m thick). This causes the shallow ice to flow faster than ice with similar thicknesses and distance from the lateral margins on the north side (Fig. 5c).

The mass flux through the calving front was calculated at $3.8 \times 10^7$ m$^3$ d$^{-1}$ which needs to be balanced by three main frontal processes: calving, submarine melting and advective advance. Both calving and advance were observed in this study but it is likely that submarine melting also has a large role in ice output at a daily timescale. For example, Xu et al. (2013) used oceanographic data to calculate a melt water flux of between 0.5 and $1.1 \times 10^7$ m$^3$ d$^{-1}$ from Store Glacier in August 2010 equivalent to 13–29 % of the mass flux calculated by our study. For comparison, Rink glacier has an ice flux of $3.0 \times 10^7$ m$^3$ d$^{-1}$ of which 27 % is estimated to be lost through submarine melting each day (Enderlin and Howat, 2013).

Figure 6. (a) Distribution and patterns of crevasses on Store Glacier. Dry crevasses, which are large structural features, are shown in orange. Narrower crevasses that are observed in the orthorectified images but whose 3-D geometry is not constructed are shown in black. The areas of water-filled crevasses are shown in blue and occur almost exclusively in zone 4. The regions of the terminus discussed are designated by the dotted red lines and are referred to as the black numbers. Transects 1 and 2 shown in the inset demonstrate how crevasses advected downstream between 1 and 2 July. In T1, a series of calving events occurred which are discussed as calving event A. In T2, the calving front advanced 16 m. (b) Illustration of the terminus of Store Glacier with ellipsoids proportional to the average length, width and orientation of crevasses shown in (a) for the respective zones. The colour of the ellipsoids represents the proportion of crevasses that are water filled in each zone where WF refers to water filled in the legend. The italicised numbers denote the density of crevasses per 10 m$^2$ in each zone. Arrows illustrate inferred direction of principal strain.

4.2 Changes occurring over a seasonal timescale

The lack of variation in the position of the lateral margins of the glacier shows that a balance is maintained between the ice flux input and submarine melting and calving output in this zone throughout the melt season. The balance could be explained by the mechanism of calving events. At the lateral margins calving is characterised by small, regu-
lar events such as calving event A (Fig. 3a). The regularity of these small events means that any small advance or retreat is regulated almost instantly by changes in calving rate which returns the lateral margins of the glacier to the same position. Calving rate could also be moderated by changes in the bathymetry. When the lateral margins advance, calving rates increase due to the abrupt deepening of the bathymetry seaward of the lateral margins of the glacier which cause basal drag to be reduced. Ice-flow acceleration can lead to increased longitudinal stretching and deeper crevassing, thereby increasing calving rate and leading to retreat to its original, bathymetrically pinned position.

The centre of the calving front is much more active with calving and submarine melt rates that vary on a seasonal timescale. We propose that the main cause of variability is due to calving rates which are highly irregular throughout the melt season (Jung et al., 2010). Our observations also support the suggestion that calving rates are dominated by major calving events which have a time interval of around 28 days (e.g. Jung et al., 2010). If the calving front advances for 28 days at 16 m d\(^{-1}\), it will advance ∼448 m. A large, single calving event can therefore yield a retreat of ∼448 m and would explain the variation in the position of the calving front during the melt season (Fig. 3b). On 25 August 2013, a tabular iceberg with a length of ∼500 m was observed to calve from the central zone of Store Glacier.

Towards the end of the melt season (23 August), a widespread surface deflation of 0.12 m d\(^{-1}\) was observed (Fig. 4a). Application of a simple degree-day model reveals that part of this lowering can be attributed to ablation. Average daily air temperatures were recorded at an automated weather station (AWS) located near the UAV launch site (Fig. 1) and, using a melt factor of 6–10 mm per degree per day (Hock et al., 2005), surface lowering due to ablation is estimated between 0.038 and 0.064 m d\(^{-1}\). It follows that ablation alone cannot account for the entire lowering rate observed and, hence, we infer an additional component of dynamic thinning due to relative strain extension across this zone, related to reduced upstream delivery of flux and/or frontal kinematics associated with enhanced late-season submarine melting and/or calving rates. GPS measurements by Ahlstrom et al. (2013) tentatively support the former interpretation and reveal that surface velocities 8 km upstream of Store’s calving front tend to decrease between July and August. However, this raises questions regarding the timescales over which dynamic thinning and surface melt occur and whether or not the flow regime across the terminus is, to some extent, isolated or operating independent from processes upstream supplying mass. Either way, these questions are beyond the scope of the data sets presented here and require a study of greater areal extent and temporal coverage.

Another important observation is the order of magnitude reduction of the area of water-filled crevasses between early July and late August (Fig. 6). Surface air temperatures directly influence the extent of water-filled crevasses. AWS data reveal that mean daily air temperature was ∼6 °C during the 4 days prior to the UAV sortie on the 2 July. In contrast, mean temperature was ∼3.5 °C on the 4 days prior to the UAV sortie on 23 August. Water-filled crevasses have been hypothesised to penetrate deeper than crevasses without water (Weertman, 1973; van der Veen, 1998) and hence act as mechanism for calving (Benn et al., 2007). The calving events observed in this study did not specifically fail at water-filled crevasses and hence our limited results show no support for this mechanism. However, studies of greater scope with daily coverage will be required to determine definitively if water-filled crevasses have any appreciable impact on calving dynamics at Store Glacier or elsewhere.

5 Conclusions and future directions

A UAV equipped with a commercial digital camera enabled us to obtain high-resolution DEMs and orthophotos of the calving front of a major tidewater glacier at an affordable price. Airborne lidar currently presents the only alternative method for acquiring DEMs with comparable accuracy and precision. However, to fly consecutive sorties in a remote environment is likely to be prohibitively expensive and with sufficient ground control points the digital photogrammetry approach may also exceed the accuracy of this technique. The three sorties flown enabled key glaciological parameters to be quantified at sufficient detail to reveal that the terminus of Store Glacier is a complex system with large variations in crevasse patterns and calving processes, surface elevations and front positions at a daily and seasonal timescale. Surface velocities vary across the terminus and are influenced by both basal and lateral drag (Figs. 4b and 5c, d). The oblique orientation and arcuate nature of crevasses suggests that the principal extending strain rate is orientated obliquely to the direction of flow and we therefore propose that resistive stresses at the terminus of Store Glacier are dominated by lateral drag (Fig. 6). With this in mind, the retreat of Store Glacier into a wider trough could significantly increase the ice discharge. We estimated that the ice flux through the calving front of Store Glacier was 13.9 Gt a\(^{-1}\) and we observed a small terminus advance between 1 and 2 July (Figs. 3a and 5a). This advance reveals that, during this period, the sum of calving and submarine melt rates are less than the ice flux. Calving is an irregular process and that the position of the calving front returned to its 12 June position by 23 August suggests that over this timescale calving and submarine melting balance ice flux (Fig. 3b). Water-filled crevasses covered 0.24 % of the survey area on 2 July but this fell to 0.025 % on 23 August (Fig. 6). It remains to be seen whether water-filled crevasses are more likely to initiate calving events but our tentative results here indicate no support for this mechanism.

Future studies, with more frequent sorties could be used to compare and investigate further glaciological changes over...
a more continuous timespan. There is also the possibility of more sophisticated payloads with radiation, albedo and other multi-band sensors as well as radar and laser altimetry. There are many potential cryospheric applications for investigation, such as sea ice, marine and terrestrial-terminating glaciers and, with increased range, ice sheets, that can be achieved with the use of repeat UAV surveys. We have demonstrated that for calving outlet glaciers, a UAV carrying a high-resolution digital camera would be sufficient to investigate the following projects:

- analysis of the thickness and back stress exerted by the ice mélangé during the winter and the effect of its break out on glacier flow, calving rate and character;
- seasonal changes in the depth, density, orientation and nature of crevassing and their impact on calving rate and character;
- the influence of daily to seasonal melt and supraglacial lake drainage on downstream dynamics and calving;
- analysis of daily to seasonal fluctuations in calving flux, terminus position and impact on upstream dynamics and thinning.

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References


Chapter 6

Calving control on early-summer flow acceleration at Store Glacier, West Greenland

6.1 Summary

The final chapter of the thesis implements some of the techniques outlined in Chapter 5 to investigate the velocity and calving patterns of Store Glacier. The paper aims to test the two hypotheses which have been proposed to explain the late spring acceleration of flow observed at many of Greenland’s outlet glaciers. The first hypothesis attributes observed acceleration to the upstream propagation of processes at the tidewater glacier calving fronts. Correlations between frontal position and upstream surface velocity at four of Greenland’s largest tidewater glaciers (Helheim, Kangerdlugssuaq, Rink and Jakobshavn) suggests that variations in basal and/or lateral drag at the terminus control flow over seasonal and longer time-scales. However, the lack of correlation between terminus position and flow at other Greenland outlet glaciers has led some studies to prefer a second theory, which asserts that tidewater velocity patterns are predominantly controlled by subglacial processes. These studies emphasize that observed flow acceleration is caused by an increase the surface meltwater flux entering the subglacial system during summer, driving high basal water pressures and basal sliding. The widely observed
deceleration of outlet glacier flow towards the end of the melt-season is thought to be due to
a switching of the subglacial system to more efficient drainage with low water pressure and
higher basal drag and is often used as evidence for a strong subglacial control on tidewater
glaciers.

In 2014, 40 surveys, with an average periodicity of 1.8 days, were acquired over the terminus
of Store Glacier between 10 May and 19 July. The dense time series allowed us to quantify
frontal ablation rates (the combination of submarine melt rate and mechanical calving), sur-
face velocities, and mélange thickness at high temporal resolution and test the two competing
hypotheses. We find that Store Glacier is controlled by both changes to calving front position
and the structure of the subglacial drainage system. However, these processes have a contrary
impact on ice flow and operate over different periods of the melt-season. We observe an early-
season acceleration caused by an increase in the frequency of large calving events. Later in
the melt-season, net flow deceleration is observed and is attributed to the rapid drainage of
supraglacial lakes that drives a switch to a more efficient subglacial drainage system.

6.2 Contribution

To complete this study, JR designed, tested and built the UAVs used for data acquisition.
JR planned the surveys and piloted over Store Glacier the UAV during the spring/summer of
2014. JR produced the orthomosaic and DEM products from the overlapping UAV imagery. JR
derived the surface velocities, terminus position, and calving dynamics from the UAV products.
JR wrote the first draft of the manuscript and edited it in response to suggestions from Alun
Hubbard. JR made all the figures.
Calving control on early-summer flow acceleration at Store Glacier, West Greenland

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Remote sensing has greatly facilitated our understanding of flow and discharge variability of Greenland’s outlet glaciers, yet the mechanisms that control seasonal ice flow acceleration and deceleration remain poorly resolved. Some argue that tidewater glacier flow is determined by resistive stresses at the calving front, whilst others contend that subglacial water pressure is the primary influence on flow. Here, we investigate the patterns and drivers of Store Glacier’s daily flow regime from repeat unmanned aerial vehicle (UAV) surveys conducted from May to July 2014. Our results demonstrate that individual high magnitude calving events trigger abrupt, short-term ice flow accelerations of up to 39% across the terminus. Moreover, it is the increase in frequency and cumulative impact of such events after the breakup of ice mélange that accounts for the widely observed early melt-season acceleration of Store Glacier. In late July, terminus velocities markedly decrease after supraglacial lake drainage events, implying that changes in the subglacial hydrological system modulate the flow regime later in the melt-season. We propose that the complex interplay between calving and subglacial processes can explain the seasonal dynamic response of Store Glacier and potentially other tidewater glaciers draining the Greenland Ice Sheet.

Two hypotheses have been proposed to explain the late spring acceleration of flow observed at many of Greenland’s outlet glaciers¹-⁶. The first theory attributes observed acceleration to the upstream propagation of processes at the tidewater glacier calving fronts⁷,⁸. Correlations between frontal position and upstream surface velocity at four of Greenland’s largest tidewater glaciers (Helheim, Kangerdlugssuaq, Rink and Jakobshavn) suggests that variations in basal and/or lateral drag at the terminus control flow over seasonal and longer time-scales ²,⁵,⁷,⁹-¹¹. Several processes have been
proposed to explain the advance or retreat of the terminus on intra- and inter-annual timescales, including variations in; submarine melting, surface-melt-induced hydrofracturing of water-filled crevasses, and the disintegration of the ice mélange, a rigid floating mixture of icebergs and sea-ice that can buttress the calving front. However, the lack of correlation between terminus position and flow at other Greenland outlet glaciers has led some studies to prefer a second theory, which asserts that tidewater velocity patterns are predominantly controlled by subglacial processes. These studies emphasize that observed flow acceleration is caused by an increase in the surface meltwater flux entering the subglacial system during summer, driving high basal water pressures and basal sliding. The widely observed deceleration of outlet glaciers flow towards the end of the melt-season is thought to be due to a switching of the subglacial system to more efficient drainage with low water pressure and higher basal drag and is often used as evidence for a strong subglacial control on tidewater glaciers.

To investigate the effect of these processes on tidewater flow variability and resolve competing hypotheses, we conducted a suite of daily UAV surveys between 10 May and 19 July at the front of Store Glacier, a large tidewater outlet that drains the western sector of the Greenland ice sheet (Fig. 1). The resulting high temporal and spatial resolution datasets of frontal position, glacier and mélange thickness and ice flow enable near-terminus processes to be quantified in response to internal and external forcing. We find that Store Glacier is controlled by both changes to calving front position and the structure of the subglacial drainage system. However, these processes have a contrary impact on ice flow and operate over different periods of the melt-season. We observe an early-season acceleration caused by an increase in the frequency of large calving events. Later in the melt-season, net flow deceleration is observed and is attributed to the rapid drainage of supraglacial lakes that drives a switch to a more efficient subglacial drainage system.

**Early melt-season calving front controlled acceleration**

Near-daily surface velocities across the terminus of Store Glacier were derived by feature tracking pairs of digital elevation models (DEMs) generated from the overlapping UAV images (See Methods and Figures S1 to S4). The velocity fields indicate that during our study period (10 May to 19 July), the terminus had a mean velocity of 16.9 m per day (Fig. 2). In the spring, between 10 May and 10 June, the mean velocity remained stable at 16.2 m per day with no apparent response to ice mélange breakout which occurred between the 4 and 7 June (Fig. 3). After the 10 June, terminus velocities steadily accelerate at a rate of 0.08 m per day, attaining a maximum of 18.7 on 19 July: equivalent to a 17%
acceleration in velocity (Fig. 2). This early melt-season acceleration is similar in magnitude to that observed at many other Greenland tidewater glaciers5,19-21.

The time-series produced from our near-daily velocity fields is punctuated by five distinct episodes of flow acceleration, lasting up to 24 hours, which increase the terminus flow by 20 to 30% compared to the previous day. We find that these distinct acceleration episodes coincide with large calving events, similar to that observed by GPS and seismometers at Jakobshavn and Helheim glaciers30-32 (Fig. 4, 5). We name these events M1 (between 16:00 16 May and 16:00 17 May), O1 (16:00 8 June and 17:00 10 June), O2 (23:00 19 June and 06:00 20 June), O3 (15:30 7 July and 23:00 7 July) and O4 (15:00 11 July and 21:00 15 July) based on time since the start of the field campaign and whether they occurred during in the presence or not of mélange (Fig. 2). The calving events are evident as spikes in frontal ablation rates (Fig. 2 and Methods) and at Store Glacier, are characterized by the fracturing and production of large icebergs (Figures S1 to S4). During a 7 hour period after the calving event O2 terminus velocities accelerated by 23% to 22.8 m per day (Fig. 5b). Similarly, the terminus also accelerated 39% and attained 22.6 m per day during a 6 hour period after calving event O3 (Fig. 5c). Calving events M1, O1 and O4 were also associated with an increase in velocity of between 2.3 (20%), 4.4 (32%) and 1.6 (9%) m per day, respectively. We find a strong correlation ($R^2 = 0.81; \ p < 0.1$) between terminus velocity and calving rate over a 14-day period, suggesting that the large calving events results in an increase in weekly to monthly velocities and furthermore, that the frequency of large calving events explains the observed early-season flow acceleration at Store Glacier.

Our results suggest that the early melt-season flow acceleration is not driven by subglacial water pressure. Whilst the drainage of supraglacial lakes to the subglacial drainage system via hydrofracture26 has also been shown to cause short-lived ice flow accelerations of tidewater glaciers19, an analysis of daily MODIS imagery (see Methods) suggests that up-glacier supraglacial lake drainage events occurred between 18 July and 28 August, much later in the season than any of the five distinct short-term acceleration events we observed (Fig. 4). Likewise, the lack of coupling between terminus velocities and air temperature, measured by an automated weather station situated 8 km of the glacier at sea-level, indicates that the implied increase in surface melt does not directly yield a discernible flow response. Mean daily air temperatures plateau after 10 June whilst the 14-day mean terminus velocity continues to increase after 10 June. The lack of correlation between terminus velocities, air temperature and supraglacial lake drainage confirms that the observed early melt-season acceleration of Store Glacier flow is not directly controlled by changes to the subglacial drainage
system. Moreover, we find that 17% of the apparent early-season flow response diminishes when the abrupt acceleration events were removed from the time-series. These results are in general agreement with GPS data from Helheim Glacier which demonstrate that the flow response attributed to increased seasonal meltwater runoff is only 2 to 4% of the summer mean.

Stress perturbations at the calving front cause rapid accelerations

Before each of the large calving events, our UAV-derived DEMs indicate that the tongue of Store Glacier becomes ungrounded, or, is at least very close to positive buoyancy (Fig. S5). The formation of a floating or at least positive-buoyant ice tongue suggests that basal traction at the terminus is low and that driving stress is predominantly balanced by lateral resistive forces generated at the glacier margins: a scenario similar to that observed at Jakobshavn, Rink, Kangerdlugssuaq and Helheim Glaciers. However, none of the large calving events observed appear to cause the lateral margins to retreat, suggesting that the mechanism by which the resulting stress perturbations trigger flow acceleration may be more complex than previously thought.

Instead, we attribute the distinct flow acceleration events observed to changes in ice viscosity at the lateral shear margins in response to high magnitude calving. A recent study argues that increases in lateral stress, initiated by frontal retreat or large calving events, increases localized strain rates which reduce the effective ice viscosity at the margins. By driving an increase in terminus speed, the viscosity drop facilitates the geometrical adjustment of the glacier to the calving event. This mechanism has been used to explain the retreat and acceleration of Jakobshavn Isbræ on inter-annual timescales but we argue that it may also explain the intra-annual flow acceleration observed at Store Glacier. The model simulations imply that the soft lateral shear margins can rapidly transfer stress perturbations originating at the calving front tens of kilometers inland through longitudinal stress coupling. This process, therefore, may not only explain velocity fluctuations at the terminus, but also the reported coupling of calving rate and upstream flow perturbations distant from the calving front. Evidence for such a mechanism is demonstrated by a GPS velocity record, located 16 km upstream from Store Glacier’s calving front. At the time, the discrete peaks in velocity recorded by the GPS, which lasted several hours, were interpreted as a calving response to upstream originating flow acceleration events. Here, we demonstrate that the calving events precede terminus flow acceleration and the velocity peaks observed 16 km upstream of the calving front are therefore interpreted as evidence for the rapid upstream propagation of stress perturbations driven by large calving events.
We find that mélange breakup did not have a notable impact on terminus velocity, demonstrating that the presence of winter mélange does not impose significant backstress and impede ice flow at Store Glacier (Fig. 2, 3). Mélange breakup does though influence the number of large calving events, which results in a net flow acceleration across the terminus, albeit in distinct episodes. Hence, even though mélange does not directly affect ice flow (through buttressing), it has an indirect effect on seasonal tidewater glacier dynamics through its impact of calving magnitude and frequency (Fig. 2). Calving, in particular high magnitude events, are significantly lower before mélange break-up, demonstrating that the relatively limited buttressing force provided by the mélange (~60 kPa) inhibits ice fracturing: an observation that agrees with observations at other tidewater glaciers and model simulations (Fig. S6). The suppression of calving due to the presence of mélange, hampers large calving events resulting in a reduced flow acceleration events, and a net slower flow regime in comparison to when the mélange is not present. The continued and steady acceleration of 14-day mean terminus velocity after mélange breakout observed, rather than a distinct step change, supports our assertion that terminus flow acceleration is driven by increased calving activity rather than the specific loss of mélange backstress (Fig. 2).

**Late melt-season deceleration controlled by subglacial drainage system**

Whilst subglacial water pressure appears to have little effect on terminus velocity during the early melt-season, we find that it may play a potential role later in the melt-season due to the propensity of supraglacial lake drainage events. Feature tracking of Landsat 8 imagery, later in the melt-season (after our UAV survey), reveals that the terminus experiences a 21% deceleration from 17.5 to 14.5 m per day from 4 to 11 August (Fig. 4). This occurs after two supraglacial lakes, with an area of 1 and 2 km², drained between the 16 and 18 July, and 20 and 25 July, respectively. Both supraglacial lake drainage events cause an initial glacier wide acceleration, but were followed by flow deceleration to slower ice flow velocity (Fig. 4). We propose that the initial acceleration was caused by large fluxes of meltwater flooding into a relatively undeveloped subglacial drainage system thereby increasing the basal water pressure, reducing basal drag and facilitating faster flow. Increased meltwater flux allows the development of a more efficient subglacial drainage system characterized by either linked cavities or possibly channels melted into ice or eroded into basal sediments. Such a drainage system has a larger area of contact between the glacier sole and the bed, increasing basal drag. Store Glacier has experienced late melt-season decelerations in response to supraglacial lake drainages in 2002, 2005, 2008 and 2012. Late melt-season deceleration have been reported for other Greenland and Alaskan
tidewater glaciers and is likely important for stabilizing late summer flow when terminus positions are at a minimum and calving fronts are hence vulnerable to being pushed off their pinning points.

**Implications**
The seasonal velocity patterns we observe at Store Glacier are characteristic of other tidewater glaciers in Uummannaq Bay and elsewhere in Greenland. The mechanisms that we identify may therefore be relevant to other tidewater glaciers, particularly those whose frontal ablation rate is dominated by large calving events and driving stress is primarily balanced by the lateral shear margins. The four largest tidewater glaciers by discharge, Jakobshavn Isbræ, Rink, Kangerdlugssuaq and Helheim Glaciers, all display these characteristics. Furthermore, short-lived but detailed GPS measurements from Helheim found high correlation between calving magnitude and ice flow response. We propose that the early melt-season acceleration displayed by other major calving glaciers draining the Greenland ice sheet, can be explained by similar mechanisms and furthermore that these fast-flow outlet glaciers are insensitive to surface meltwater-enhanced basal lubrication.

The interplay between calving-induced ice flow acceleration and subglacially-induced late-summer deceleration, which is likely to vary uniquely between each glacier-fjord setting, may explain observations of both the synchronous and asynchronous Greenland tidewater glacier retreat. For example, large, fast-flowing tidewater glaciers whose calving rates are dominated by mechanical fracture and calving may be more sensitive to early ice mélange breakup than glaciers whose calving rates are dominated by submarine melting. When mélange breaks up earlier, calving rates increase earlier in the melt-season and the terminus will be vulnerable to retreat off its pinning point. This is exemplified in Uummannaq Bay, where three relatively fast-flowing glaciers (Ingia Isbræ, Umiamako Isbræ and Sermeq silardleq) all underwent synchronous multi-year retreat starting in 2003 coinciding with the least sea-ice formation and early mélange clearing. On the other hand, glaciers which experience late melt-season slowdown in response to supraglacial lake drainage events may be more resilient to environmental forcing than glaciers that do not experience late melt-season deceleration. There is evidence from both Store Glacier and others tidewater outlets draining the Greenland ice sheet that late-season deceleration does not always occur and it is during these years that glaciers will be more vulnerable to multi-year retreat.
References


Methods

UAV platform and imagery

Aerial imagery were acquired by a fixed-wing UAV identical to that described by Ryan et al. (2015). The UAV has a 210 cm wingspan and is powered by a 10 Ah, 16.8 V LiPo battery pack which, with a total weight of 4 kg, yields a 1-hour endurance and 60 km range. The autonomous control system is based around an Arduino navigation and flight computer updated in real-time by a 10 Hz data stream comprising of a GPS, magnetometer, barometer and accelerometer. These data are logged along with a timestamp for each activation of the digital camera shutter which automatically triggers when a horizontal displacement of 40 m is exceeded. The UAV was hand launched from a base camp located on the northern peninsula adjacent to Store Glacier (70.399º N, 50.668º W) and was belly landed on a short boulder-strewn patch of tundra (Fig. 1). The UAV was pre-programmed to carry out a 30 to 40 km survey at a constant altitude of 300 m above the calving front of Store Glacier every day between 10 May and 19 July 2014 (Fig. 1). Around 1,000 images of the glacier were obtained on each survey covering a mean area 12.8 km². The Greenland Ice Mapping Project (GIMP) digital elevation model (DEM) was used during the selection of 3D waypoints to ensure the UAV did not collide with the steep valley side to the south of the glacier.

Digital elevation model generation

Digital elevation models (DEMs) of the terminus were generated from the overlapping UAV images using Agisoft PhotoScan Pro software (http://www.agisoft.com/). Specific details of the processing chain can be found in Ryan et al. (2015). The twelve large boulders (six on each side of the glacier) were used as ground control points (Fig. 1). We surveyed the corner of each 5 boulder using an differential GPS and subsequently located these points in the UAV images for accurate georectification (uncertainty quantified in Sect. 2.3.1 and 2.3.3). The dense point clouds produced by PhotoScan, which contained about 68 million points for each survey, were exported to GeoTIFF format with a WGS84 UTM 22N (EPSG:32622) projection and a pixel size of 0.5 m.

Surface velocity fields from feature tracking

Feature tracking was applied to pairs of UAV DEMs using the ImGRAFT toolbox which uses a standard normalized cross-correlation algorithm. Template or correlation window sizes between 40 and 160 m were used depending on the time period between UAV surveys and 1 pixel in every 50 were tracked. Feature-tracked points were filtered using a strict correlation coefficient threshold of 0.4
and a signal-to-noise ratio threshold of 2. Point vector measurements of velocity were converted to magnitude using trigonometry and linearly re-interpolated over the original 0.5 m grid to produce a velocity field (e.g. Fig. 2). Terminus velocities were sampled with a rectangular polygon in the centre of the terminus, about 500 m from the calving front (Fig. 1). Uncertainties in the surface velocity fields were calculated by sampling areas of stationary bedrock on the north and south side of the glacier (Fig. 1). We estimated a mean horizontal uncertainty of ±0.55 m per day, which is likely comprised of co-registration errors and errors from interpolation of point vectors.

**Terminus position and frontal ablation**

Terminus position change was derived by locating the intersection of each calving front position with a set of equally spaced along-flow reference lines at 150 m intervals across the active glacier terminus width (e.g. 48) (Fig. 2). The magnitude and direction of terminus position change were calculated using the Euclidean distance between these intersections for each calving front pair. The average surface velocity for the corresponding calving front pairs was also extracted along each reference line, and the frontal ablation rate was calculated at each intersection by subtracting the change in terminus position from the terminus ice speed. These frontal ablation rates were averaged to produce a mean frontal ablation rate for the entire glacier width. The cumulative sum of terminus position change allowed us to monitor the relative state of front position advance or retreat during 2014 (Fig. 2). We were unable to digitize the front position in the northern section of the glacier because it was unclear where the boundary between the mélange and the glacier was (Fig. 2, 3). Terminus position and frontal ablation rates were averaged over the southern 3.5 km of the terminus and may therefore be higher than studies which measured terminus position change over the entire width of the glacier. Uncertainties in the rate of terminus position change are likely due to digitization error which is thought to be ±4 pixels. At a pixel size of 0.5 m, this translates to an uncertainty of ±2 m per day in terminus position and frontal ablation rates.

**Mélange thickness**

The mean surface elevation (freeboard) of the ice mélange, a rigid mixture of icebergs and sea-ice floating in front of the glacier terminus, was sampled at two sites of the fjord (Fig. 1). We estimated a mean vertical uncertainty of ±0.97 m in the surface elevation by sampling areas of stationary bedrock on the north and south side of the glacier (Fig. 1). The thickness of the mélange was calculated based on the assumption of an ice density of 917 kg m$^{-3}$ and a sea water density of 1028 kg m$^{-3}$. 
At the start of the survey period, the mélange thickness averages nearly 140 m on the northern part of the fjord and 38 m in the centre (Fig. 3). Mélange thickness remains constant at both portions of the glacier until it clears between the 4 and 7 June (Fig. 4). After clearing, the average mélange thickness remains below 10 m for the rest of the survey period with variation attributed to the presence or absence of icebergs. Our field observations and environmental data suggest that a combination of calving and high wind speeds destroyed the mélange. These observations are consistent with discrete-element model simulations which indicate that calving activity is able to fracture the sea-ice bonds between icebergs, reducing the mélange’s rigidity. The high wind speeds then push the fractured mélange away from the calving front. In early spring, when air and ocean temperatures are below zero, the mélange may reform over a period of a few days. But the high wind speeds and positive air temperatures between 4 and 7 June precluded the reformation of the mélange.

**Supraglacial lake drainage events**

Supraglacial lake drainage events were detected using MOD02 Level 1B raw product from NASA’s Terra satellite following the method described in Cooley and Christoffersen (2017). The MOD02 product has a resolution of 250 m and, although it is available at subdaily time scales, quality filtering was used to select the best daily image. The criteria for lake drainage was that it should drain in less than 4 days between two cloud-free images and lose 90% or more than 1.5 km² of its area.

**Climate variables**

Air temperature and wind speed were measured a few metres above sea-level by an automated weather station situated on an island about 8 km from the calving front of Store Glacier. Air temperature and wind speed were measured and logged every 15 minutes and were resampled to daily means for analysis. The onset of the melt-season was defined as the first day of a period of least three consecutive days where air temperatures were > 0°C.
References


Figure 1. (a) Landsat 8 OLI true colour image of Store Glacier on 19 July 2014. (b) Orthomosaic generated from overlapping UAV images obtained on 19 July 2014. Coloured lines show terminus positions between 10 May and 19 July 2014.
Figure 2. Timeseries of glacier variables: velocity, frontal ablation, terminus position and environmental variables: melange thickness, air temperature and wind speed.
Figure 3. UAV-derived velocity fields and DEMs first showing the lack of response to melange breakup between the 4 and 7 June and secondly showing the speed-up of the calving front immediately after a calving event. After the calving event between the 19 and 20 June, the centre of the calving front accelerates to ~22 m per day. See supplementary information all the velocity fields and DEM residuals.
Figure 4. Seasonal velocity of Store Glacier from Landsat 8 feature tracking showing timing of three key events: calving, melange breakup and supraglacial lake drainages.
Supplementary Information for:

**Calving control on early-summer flow acceleration at Store Glacier, West Greenland**

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**TABLES**

Table S1: Dates and periodicity between UAV surveys over Store Glacier in 2014 and whether or not the ice melange was present.

Table S2: Glacier flow speeds, frontal ablation rates and melange thickness during the survey period.

**FIGURES**

Figure S1: Velocity fields, velocity residuals and DEM residuals for UAV surveys between 10 May and 30 May 2014.

Figure S2: Same as Figure S1 but for 30 May to 16 June.

Figure S3: Same as Figure S1 but for 16 June to 08 July.

Figure S4: Same as Figure S1 but for 08 July to 16 July.

Figure S5: Floatation thickness of Store Glacier on 16 May, the day before a large calving event. A floatation thickness higher than zero indicates that the terminus may be ungrounded or very close to positive buoyancy.

Figure S6: DEMs and elevation profiles showing the melange freeboard in (A) spring, (B) preceding breakup and (C) after breakup.
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**Table S1.** Dates and periodicity between UAV surveys over Store Glacier in 2014 and whether or not the ice melange was present.
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**Table S2.** Glacier flow speeds, frontal ablation rates and melange thickness during the survey period.
Figure S1. Velocity fields, velocity residuals and DEM residuals for UAV surveys between 10 May and 30 May 2014.
Figure S2. Same as Figure S1 but for 30 May to 16 June.
Figure S3. Same as Figure S1 but for 16 June to 08 July.
**Figure S4.** Same as Figure S1 but for 08 July to 16 July.

**Figure S5.** Floatation thickness of Store Glacier on 16 May, the day before a large calving event. A floatation thickness higher than zero indicates that the terminus may be ungrounded or very close to positive buoyancy.
Figure S6. DEMs and elevation profiles showing the melange freeboard in (A) spring, (B) preceding breakup and (C) after breakup.
Chapter 7

Conclusions

7.1 Methodological summary

The first aim of the thesis was to establish whether UAVs could be useful for investigating processes of Greenland Ice Sheet mass loss. This was achieved using UAV imagery acquired over a 280 km$^2$ area in the ablation zone (Chapter 2) and a tidewater outlet glacier draining the ice sheet (Chapter 5) in 2013 and 2015, respectively. In these chapters, methods were developed to generate scientifically accurate, georeferenced products from UAV imagery that could be used for understanding glaciological processes.

Chapter 2 demonstrated that accurate (uncertainty less than 5%), high resolution and spatially continuous albedo products could be derived using UAVs equipped with a digital camera and broadband pyranometers. The technique was found to be reproducible and provided a practical approach for overcoming the limitations of previous studies caused by variable illumination. The high-resolution albedo products offered a number of opportunities for studying bare ice albedo that could not be achieved using ground-based or satellite data. Two of these applications were pursued in Chapters 3 and 4.

Chapter 5 demonstrated that key glaciological parameters of a large tidewater glacier could be quantified with repeat UAV surveys. The analysis revealed that the terminus of Store Glacier is a complex system that undergoes large variations in crevasse patterns, surface velocities, calving rates, surface elevations and front positions at a daily and seasonal timescale. It was concluded
that more frequent surveys might be useful for understanding this complex behaviour. The next year, this was accomplished with 40 surveys between early-May and mid-July and the results of this study were presented in Chapter 6.

7.2 Glaciological summary

The second aim of the thesis was to apply the methods developed in Chapters 2 and 5 to processes of Greenland Ice Sheet mass loss at the ice-ocean and ice-atmosphere boundaries.

At the ice-atmosphere boundary, Chapter 3 demonstrated how UAV imagery could bridge the spatial gap between in situ and satellite measurements. Aerial imagery acquired by a fixed-wing UAV allowed us to quantify the spatial heterogeneity at two AWS sites and determine whether comparison between albedo measured by in situ measurements and satellite products is justified. It was found that the ice sheet surface was not necessarily homogeneous at both the scale of the AWS pyranometer footprint and the MODIS pixel footprint and that caution must be exercised when validating satellite albedo retrievals using in situ (e.g., AWS) measurements. At two sites in the ablation zone, aerial imagery demonstrated that AWS-mounted pyranometer footprints were insufficiently large to sample the true spatial heterogeneity of ice surface albedo in July, and hence, the in situ measurement could not be justified as a valid ground truth. This result justified the removal of unrepresentative in situ measurements from satellite validation exercises and, as a result, improved statistical inferences regarding albedo decline across the Greenland Ice Sheet, increasing confidence in the assimilation of albedo into regional climate and melt runoff models for the prediction of global sea level rise.

Also at the ice-atmosphere boundary, Chapter 4 demonstrated the usefulness of fine spatial resolution imagery for resolving small features of the Greenland Ice Sheet such as meltwater streams and cryoconite holes. The analysis revealed that the dark zone of the ice sheet, which is an area of bare ice with particularly low albedo, is characterized by ice containing uniformly distributed impurities draped over a relatively flat surface. The impurities were interpreted as evidence for surface ice algal that feed off the melt-out of particulates and grow in response to increased air temperatures and liquid water availability. The result suggests that further research is needed to determine how the spatial extent and concentration of pigmented algae
responds to an increase in temperature and/or liquid water production.

At the ice-ocean boundary, Chapter 6 demonstrated how fine temporal resolution imagery could be used to understand seasonal velocity variations at Store Glacier. Repeat UAV surveys were used to produce a dense times series of calving rates, surface velocities and mélange thickness at Store Glacier between May and July. It was found that five distinct episodes of flow acceleration, lasting up to 24 hours, which increased the terminus flow by 20 to 30% coincided with large calving events. The results indicated that the early melt-season flow acceleration was not driven by subglacial water pressure which is contrary to many other studies. Later in the melt-season (late July) terminus velocities markedly decreased after a supraglacial lake drainage event, implying that changes in the subglacial hydrological system modulate the flow regime later in the melt-season. It is concluded that the seasonal acceleration and deceleration displayed by other major calving glaciers draining the Greenland Ice Sheet, could be explained by similar mechanisms.

7.3 Synthesis

In summary, this thesis has satisfied its two aims, 1) to establish whether UAVs are useful for investigating processes of Greenland Ice Sheet mass loss and 2) to demonstrate the usefulness of UAVs for understanding these processes at both the ice-ocean and ice-atmosphere boundaries. The UAV imagery collected in this thesis has provided insights into ice sheet processes which were previously unobservable by the current suite of satellite and ground-based data acquisition technologies. The choice to collect data over a large tidewater glacier was obvious since the calving processes are so dynamic and occur on timescales of minutes to hours. But Greenland has many other dynamics environments. Supra-glacial lakes have been observed to drain in a matter of hours, delivering large volumes of water into the ice-bed interface causing vertical uplift and enhancing horizontal movement (Doyle et al., 2013). Likewise, Greenland’s pro-glacial lakes, which have also been observed to drain in a matter of hours, have received attention recently as they regulate the storage and discharge of sediment and water to the oceans. UAV data may therefore be useful for understanding the mechanisms by which these lakes drain and the impact these drainages have on the surrounding environments. The choice of investigating bare
ice was also obvious since it shows large spatial heterogeneity. But snow also exhibits spatial heterogeneity in reflectance and topography due to aeolian erosion and deposition. UAVs may therefore be useful for understanding the dynamics of these features and how they respond to environmental forcing. UAVs are likely to become increasingly common in glaciological field studies and their use may lead to further insights into processes of Greenland Ice Sheet mass loss.


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