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Mass loss from glaciers and ice caps represents the largest terrestrial component of current sea level rise. However, our understanding of how the processes governing mass loss will respond to climate warming remains incomplete. This study explores the relationship between surface elevation changes (dh/dt), glacier velocity changes (du/dt), and bedrock topography at the Trinity-Wykeham Glacier system (TWG), Canadian High Arctic, using a range of satellite and airborne datasets. We use measurements of dh/dt from ICESat (2003-2009) and CryoSat-2 (2010-2016) repeat observations to show that rates of surface lowering increased from 4 m yr^{-1} to 6 m yr^{-1} across the lowermost 10 km of the TWG. We show that surface flow rates at both Trinity Glacier and Wykeham Glacier doubled over 16 years, during which time the ice front retreated 4.45 km. The combination of thinning, acceleration and retreat of the TWG suggests that a dynamic thinning mechanism is responsible for the observed changes, and we suggest that both glaciers have transitioned from fully grounded to partially floating. Furthermore, by comparing the separate glacier troughs we suggest that the dynamic changes are modulated by both lateral friction from the valley sides and the complex geometry of the bed. Further, the presence of bedrock ridges induces crevassing on the surface and provides a direct link for surface meltwater to reach the bed. We observe supraglacial lakes that drain at the end of summer and are concurrent with a reduction in glacier velocity, suggesting hydrological connections between the surface and the bed significantly impact ice flow. The bedrock topography thus has a primary influence on the nature of the changes in ice dynamics observed over the last decade.

Keywords: Glacier change, optical remote sensing, Landsat, altimetry, subglacial topography, dynamic thinning.

1. Introduction

The rate of mass loss from glaciers and ice caps to the ocean has accelerated in response to global heating (Zemp et al., 2019), but our understanding of the mechanisms controlling these changes remains incomplete, making future projections of mass loss from the cryosphere highly uncertain. Central to this problem is the complex nature of marine-terminating glacier dynamics and their sensitivity to changes at the ice front (Howat et al., 2008; McMillan et al., 2014; Willis et al., 2018). This is of particular concern in the Arctic, which continues to warm at twice the rate of lower latitudes (Overland et al., 2016) due to an amplified increase in northern hemisphere high latitude temperatures attributed to the strengthening of the ice-albedo positive feedback (Serreze and Francis, 2006). Elevated Arctic temperatures have resulted in an increasingly negative surface mass balance for Glaciers and Ice Caps (GIC) across the region (Serreze and Barry, 2011). However, understanding how the flow of marine-terminating glaciers is affected by climate warming remains elusive and large uncertainties exist in estimates of their impact on future sea level.

As the climate continues to warm, changes in ice flow have led, in general, to a long-term increase in ice discharge from Arctic tidewater glaciers (Gardner et al., 2013), but the magnitude of this effect varies between individual glacier catchments. Understanding the drivers of ice dynamic changes across catchments and over time is key to understanding this variability. In particular, thinning of marine-terminating glaciers can lead to ungrounding of the glacier from its bed (McMillan et al., 2014), causing a loss of basal traction and enhancing the flow of ice near the terminus. Increased ice velocity and buoyancy can lead to further mass loss through increased calving (James et al., 2014), as well as enlarging the area of contact between ice and the ocean, causing a positive feedback loop of enhanced submarine melting, thinning, and retreat (Murray et al., 2010).

Outlet glaciers transport ice from the accumulation area in the ice cap interior to the surrounding ocean and the rate at which they flow is highly sensitive to local fjord geometry, i.e. the morphometry of the subglacial bedrock or valley sides (Joughin et al., 2004). However, this effect remains poorly understood due to the lack of high-quality bedrock elevation data across highly crevassed glaciers confined to small valleys, which represent technical challenges for the ice penetrating radar instruments used to measure ice thickness (Conway et al., 2009). The relative influence of other forcing factors also remains unresolved; high-elevation thickening in interior ice-cap drainage basins, ocean-induced melt at the glacier terminus and increased surface runoff may all increase ice discharge in the future. Further, the bedrock geometry may significantly impact the rate of ice discharge from marine-terminating glaciers where they are grounded below sea level (van Wychen et al., 2016) or are reverse-sloping.

In this paper, we combine observations from aerogeophysical surveys, satellite altimeters and satellite multi-spectral imagers acquired between 2000 and 2016 in order to understand the influence of local fjord geometry on recent dynamic changes at the Trinity-Wykeham Glacier system (TWG), Nunavut, Canada (van Wychen et al., 2014, 2016; Millan et al., 2017). We had three objectives: (i) to map the bedrock geometry of the TWG using ice-penetrating radar and analyse the spatial pattern of subglacial landforms; (ii) to derive annual velocity estimates for each year between 2000 and 2016, and seasonal (spring, summer and winter) velocity estimates for each year between 2013 and 2016 and to compare this to the pattern of surface elevation change (dh/dt) over the ICESat (2003-2009) and CryoSat-2 (2010-2016) study periods; (iii) to assess the interactions between glacier velocity changes, dh/dt and bedrock topography in order to understand the influence of the glacier bed on changes in glacier dynamics between 2010 and 2016.

2. The study area

The Queen Elizabeth Islands (QEI), located in the Canadian Arctic (see Figure 1), contain almost one third of the land-based ice outside the Greenland and Antarctic ice sheets (Radic and Hock, 2010). Regional rates of mass loss have increased from 6.3 Gt yr^{-1} (1991-2005) to 33.1 Gt yr^{-1} (2005-2014) (Millan et al., 2017). Previously, this change was thought to be primarily driven by an increase in summer air temperatures after 2005 leading to a more negative surface mass balance (Gardner et al., 2011; Lenaerts et al., 2013; Colgan et al., 2015; Millan et al., 2017; Noel et al., 2018). However, recent studies have suggested that mass loss from selected tidewater glacier catchments has accelerated in recent years (van Wychen et al., 2016; Millan et al., 2017), but the origin and mechanisms of this loss are not well understood. It is vital to understand changes across the QEI and the factors controlling them so that we can accurately determine near-future mass loss from the region and its impact on sea level changes.

[Insert Figure 1 here]

In the Canadian Arctic, four glaciers (Trinity Glacier, Wykeham Glacier, Belcher Glacier and Yelverton Glacier) make a significant (>0.1 Gt yr^{-1}) contribution to regional mass loss via calving at their termini (van Wychen et al., 2014, 2016; Millan et al., 2017). Notably, ice discharge from the Trinity-Wykeham Glacier system (TWG), which has a total drainage-basin area of 3,046 km^2 and is part of the Prince of Wales Ice Field (POW) (see Figure 1), increased from 0.55 Gt yr^{-1} in 2000 to 1.43 Gt yr^{-1} in 2015, with the latter representing 63% of the total ice discharge from the QEI in 2015 (van Wychen et al., 2016). This is considerably larger than the average rate of ice discharge from other QEI outlet glaciers, largely because flow rates of many glaciers in the region are well below 1 km yr^{-1} (van Wychen et al., 2014, 2016). Millan et al. (2017) suggested that this increase in ice discharge may be driven by the transport of warmer ocean waters from the Nares Strait, but there is no oceanographic data from the TWG fjord to support this. In comparison, Cook et al. (2019) show a strong correlation between atmospheric warming and glacier retreat, but did not include the impact of velocity or surface elevation changes in their assessment. The impact of subglacial topography, fjord geometry or bathymetry was not included in any of these studies but may play an important role in controlling the rate of ice discharge from the TWG and its response to changes at the ice front.

3. Methodology

In this section we consider the datasets used to derive estimates of the subglacial topography, dh/dt and mean rates of horizontal glacier velocity change (du/dt). Ice front retreat rates and predicted areas of flotation are also derived.

3.1. Surface elevation change – ICESat and CryoSat-2

Rates of surface elevation change (dh/dt) over the TWG were measured using repeat ICESat tracks (2003-2009), and swath-mode processing of the European Space Agency's (ESA) CryoSat-2 data (2010-2016). ICESat tracks over 6 years (2003-2009) were obtained using the Geoscience Laser Altimeter System (GLAS), which acquired point elevation data from a 64 m diameter footprint on the ground, and at 170 m intervals along-track (Abshire et al., 2005). GLAS/ICESat L1b Elevation Data Version 34 were downloaded directly from the National Snow and Ice Data Centre (NSIDC) website (www.nsidc.org/data/gla06). Measuring dh/dt from repeat ICESat tracks is difficult because the tracks can be offset by as much as 300 m over the TWG. This drawback was overcome by expanding the measurement area across-track by constructing planar surfaces of surface elevation (m) (h) and acquisition date (days) (t) within 2-year epochs using Triangular Irregular Networks (TIN) (Pritchard et al., 2009). We firstly isolated point measurements of elevation (h) and time (t). We then interpolated these measurements by constructing planar TIN surfaces within a 300 m radius around the measurement points. Each ICESat track that overlaps with these surfaces was differenced from it to obtain estimates of elevation change (dh) and the associated time difference (dt). dh/dt were measured directly from this. Erroneous dh/dt values (>20 and <-20) were then removed and mean rates of dh/dt were calculated by averaging within a 300 m radius. This method achieved greater spatial coverage than previous methods (Felikson et al., 2017) and extended across-track measurements of dh/dt to 0.8 km from the original ICESat measurements. A detailed review of the method can be found in Pritchard et al. (2009) and Felikson et al. (2017). The uncertainty of dh/dt using this method was estimated as ± 0.1 m yr^{-1} along-track and ± 0.07 m yr^{-1} across-track (Pritchard et al., 2009).

CryoSat-2 L1b Interferometric Synthetic Aperture Radar (SARIn) data for the TWG were acquired from the ESA website (ftp://science-pds.cryosat.esa.int/). Surface elevation was extracted using swath-mode processing of the CryoSat-2 data. This approach utilised the full altimetric waveform across the satellite ground track to generate a dense set of elevation points across a swath of up to 5 km (Foresta et al., 2016; Gourmelen et al., 2018). Echoes from across the beam were combined via SAR processing, in which a global phase unwrapping procedure was applied to account for steep sloping glacial valleys, both across and along the valley slopes (Gourmelen et al., 2018). This technique led to an improvement in spatial sampling from conventional CryoSat-2 Point-Of-Closest-Approach (POCA) products by an order of magnitude over the QEI region, and also improved echo location accuracy over sloping terrain (Wingham et al., 2009). The generation of multiple elevation swaths was then used to measure d*h*/d*t* at greater spatial and temporal resolutions than was previously possible via POCA, and enhanced d*h*/d*t* mapping across the variable terrain of the TWG. The maximum error for d*h*/d*t* from CryoSat-2 swath altimetry was ± 1 m yr^{-1}, although values are frequently smaller (± 0.5 m yr^{-1}) (Foresta et al., 2016; Gourmelen et al., 2018).

3.2 Subglacial topography

Point measurements of ice thickness were acquired from two separate airborne radar surveys (see Figure 1) over the TWG; (1) a Scott Polar Research Institute and University of Texas Institute for Geophysics (SPRI-UTIG) Natural Environment Research Council (NERC) funded Canadian Arctic Geophysical Exploration (CAGE) flight on 3 May 2014, equipped with the High Capability Radar Sounder (HiCARS-2) at 60 MHz; and (2) a NASA Operation IceBridge (OIB) mission on 6 May 2014 (Leuschen et al., 2010), using an airborne ice-penetrating radar with a central frequency of 195 MHz for the Multichannel Coherent Radar Depth Sounder (MCoRDS). The uncertainty of the HiCARS-2 instrument was ~7 m over smooth surfaces (Peters et al., 2005; Blankenship et al., 2017) but can be as large as 50 m over rough terrain based on crossover analysis (Young et al., 2017). These values were adopted as an estimate of uncertainty for the ice thickness values. The ice penetrating radar measurements from the OIB and CAGE flights were used to derive bedrock elevation by subtracting measured ice thickness from an independent gridded surface elevation dataset. We used the 2 m resolution ArcticDEM, obtained from pairs of stereoscopic WorldView imagery (Noh and Howat, 2015; Morin et al., 2016), for this purpose.

3.3. Annual and seasonal ice surface velocity

Annual and seasonal ice flow estimates were determined using pairs of Landsat 7, Landsat 8 and ASTER imagery. Each ASTER image was orthorectified using the COSI-corr feature-tracking software (Leprince et al., 2007) and coregistered to the ASTER GDEM through iterative minimisation of tie-points generated between the image and the DEM. The image was then resampled onto this new grid and projected onto the Universal Transverse Mercator (UTM) coordinate system. Every image scene was clipped to a bounding box of the Global Land Ice Measurements from Space (GLIMS) (www.glims.org/maps/glims) (Bolch et al., 2014) TWG glacier catchment in order to reduce processing time. A high pass filter was then applied to each geo-rectified image in order to enhance the contrast between surface features.

Image pairs with a ~365-day separation were used to estimate annual velocity. Glacier surface features are likely to undergo significant alterations over the year due to melting, transient snow cover and changes in the pattern of stress and strain within the glacier, leading to errors in the feature-tracking result. To avoid such issues, we created velocity stacks from multiple pairs of satellite images across two epochs, T_0 and T_1, to obtain an annual velocity estimate for T_1. For example, if T_0 = 1999 and T_1 = 2000, the annual velocity estimate would be for 2000. Annual velocity was estimated by acquiring images between June and July for years 1999-2016 and pairing each image from T_0 with those from T_1. For years 2013-2016, spring velocity was estimated by setting T_0 to April and T_1 to June. Summer velocity was estimated by setting T_0 to July and T_1 to September (T_0 was set to June for the summer of 2013 due to the absence of Landsat 8 images in July of that year). Winter velocity (for years 2013-2016) was estimated by setting T_0 to October of the previous year and T_1 to March of the successive year. For years 2013 and 2014, T_0 was set to March and T_1 was set to May due to the absence of suitable images. Images with significant cloud cover were manually removed.

Glacier surface velocity for each image pair was calculated using the COSI-corr feature-tracking software (Leprince et al., 2007). Spurious data were removed from each velocity estimate based on the Signal-to-Noise ratio (SNR), the standard deviation of ice velocity and the standard deviation of flow direction calculated from the x and y components of velocity. We defined spurious data as those with a normalised SNR of less than 0.9, an ice velocity standard deviation of greater than 40 m yr^{-1} and a flow direction standard deviation of greater than 20°. For each annual, summer, and winter velocity estimate we then constructed a velocity stack and merged them together by taking the median of each cell. An error estimate for each velocity grid is calculated from the mean displacement over ice-free terrain (assumed to be static) and these are presented in Table 1.

Table 1. Error estimates for each velocity grid obtained by calculating the mean displacement across stable terrain.

Velocity Estimate	**Year**	**Error (m yr^{-1})**
Annual	2000	19.6
	2001	12.0
	2002	13.5
	2003	25.4
	2004	18.6
	2005	39.0
	2006	73.6
	2007	21.0
	2008	15.4

	2009	28.5
	2010	59.8
	2011	63.0
	2012	84.1
	2013	85.0
	2014	17.3
	2015	8.0
	2016	10.8
Spring	2013	46.6
	2014	25.7
	2015	68.6
	2016	90.7
Summer	2013	46.9
	2014	56.5
	2015	80.8
	2016	112.7
Winter	2013	92.8
	2014	56.8
	2015	24.7
	2016	28.8

3.4. Grounding-line and ice-front retreat

The hydrostatic flotation depth (P) measures the thickness of ice required to cause buoyancy (flotation) and can be used to estimate the position of the grounding-line (the point of transition between grounded and floating ice). If the terminus of a glacier is floating it is no longer subject to basal friction, enabling faster flow and higher rates of ice discharge. P can be estimated based upon the glacier freeboard elevation (h) (Le Meur et al., 2014), i.e. height above sea level (a.s.l.):

$$P = \frac{\rho_i h}{\rho_w - \rho_i} \quad (1)$$

where ρ_i is the ice density (890 kg m^{-3}) and ρ_w is the ocean density (1028 kg m^{-3}). We estimated h by using a 2010 CryoSat-2 surface elevation swath, which was referenced to the Earth Gravitational Model (EGM96) geoid. Regions of ice flotation were estimated by differencing the CAGE-OIB ice thickness grid (resampled to 500 m) from P. This produced a grid of values denoting floating ice ($P < 0$). This hydrostatic method assumes constant ocean and ice densities, as well as accurate surface elevation and ice thickness data. Because the hydrostatic method is simple and neglects internal stresses (Le Meur et al., 2014), the location of floating ice is denoted as a prediction rather than an observation.

The ice front of the TWG was digitised manually using pan-sharpened late-summer Landsat images for every year between 2000 and 2016, setting 31 July as a baseline. The new ice front positions were used to update the GLIMS catchment polygons for each year between 2000 and 2016, which we used to clip velocity data and estimates of ice flotation. We estimated the uncertainty in ice front position estimates to be half a pixel, in this case ± 8.5 m (pan-sharpened Landsat images have a spatial resolution of 15 m).

4. Results

4.1. Changes in surface geometry

Between 2003 and 2009, the lowermost 10 km of Trinity Glacier underwent thinning of 3-4 m yr^{-1}. Further up-glacier (17-20 km), rates of thinning were smaller at ~1 m yr^{-1}. This resulted in a decrease in thinning rates from the glacier snout to 20 km up-glacier of 0.15 m yr^{-1} km^{-1}. Between 2010 and 2016, sustained thinning at rates of 4-6 m yr^{-1} persisted across the lowermost 10 km of Trinity Glacier. Meanwhile, rates of thinning 17-20 km up-glacier were between 0 and 2 m yr^{-1} (Figure 2*b*). In comparison, Wykeham Glacier experienced an asymmetric pattern of thinning across its 6 km-wide terminus in 2009; its northern tongue thinned at a rate of 4 m yr^{-1}, while its southern tongue thinned at a rate of 2.5 m yr^{-1}. Thinning on the northern tongue of Wykeham Glacier decreased from 5 m yr^{-1} to 2 m yr^{-1} between the ICESat (2003-2009) and the CryoSat-2 (2010-2016) observation periods, while thinning of the southern tongue of Wykeham Glacier increased at the same rate as on Trinity Glacier. Overlapping measurements between ICESat and CryoSat-2 (Figure 2*c*) show that dh/dt has become more negative during the CryoSat-2 study period (2010-2016). Overall, the CryoSat-2 results suggest that rates of surface thinning at the termini of both Trinity Glacier and Wykeham Glacier increased by 1-2 m yr^{-1} compared to the ICESat results.

[Insert Figure 2 here]

4.2. Subglacial topography

The bedrock topography of Trinity and Wykeham glaciers remains below sea level for ~40 km and ~30 km inland, respectively (Figure 3). A 30 km-long trough lies beneath the northern margin of Trinity Glacier (Trough #2), with a similar feature running parallel to this for 8 km along its southern margin (Trough #1). In contrast, a set of three overdeepenings characterises the bed of Wykeham Glacier, interspersed with subglacial ridges that appear to have been eroded by the glacier due to their alignment perpendicular to ice flow. The most prominent set of these is present ~5 km from the ice front (Ridges #1) and rises to 200-300 m above the surrounding bed but remaining below sea level. The overdeepenings beneath Wykeham Glacier, and to a lesser extent Trinity Glacier, cause sections of the bed to become reverse sloping; that is, they slope downwards in an up-glacier direction. At the termini of both glaciers a small region of elevated topography is observed (the 'pinning point' in Figure 3) and acts as a barrier to ice flow.

[Insert Figure 3 here]

4.3. Annual changes in velocity and terminus position

Between 2000 and 2016, velocity at the terminus of Trinity Glacier doubled from ~500 m yr^{-1} to ~1,000 m yr^{-1} (Figures 4*a* and 4*b*). In contrast, the ice front of Wykeham Glacier showed more complex behaviour; its southern tongue doubled in speed, while its northern tongue stabilised (Figures 4*a* and 4*b*). An anomalous area of decelerating flow is observed at the terminus of Wykeham Glacier and had a velocity that remained constant at 50 m yr^{-1} between 2000 and 2016. The increase in glacier velocity that originated at the terminus of Trinity Glacier and the southern tongue of Wykeham Glacier propagated inland, leading to an almost doubling of flow speed between 2000 and 2016 up to 20 km up-glacier. Velocity at the terminus of Trinity Glacier increased by 150 m yr^{-1} between 2003 and 2009 and by a further 200 m yr^{-1} between 2010 and 2016 (Figures 4*c* and 4*d*). Propagation of the velocity increase inland was more coherent during the CryoSat-2 study period than in the ICESat observation period. A 30 m yr^{-1} velocity increase on the northern tongue of Wykeham Glacier is observed only during the ICESat period (Figure 4*c*). Sections of the lowermost 10 km of Wykeham Glacier showed a velocity decrease of ~70 m yr^{-1} between 2003 and 2009 and then a velocity increase of ~70 m yr^{-1} between 2010 and 2016. This led to a total change in du/dt of ~100 m yr^{-2} at those localities. Overall, du/dt appears to have increased over the CryoSat-2 study period relative to the ICESat observation period and the region affected by these changes appears to be spreading inland.

[Insert Figure 4 here]

Retreat of Trinity Glacier has led to its separation from Wykeham Glacier (Figure 4*b* and 4*e*), with the result that the flow regimes of the two glaciers have become independent of each other. Three regimes of ice front change have been identified (Figure 4*e*): Trinity Glacier (A-A'), the Trinity-Wykeham Glacier confluence (B-B'), and Wykeham Glacier (C-C') (Figure 4*b*). Between 2000 and 2016, Trinity Glacier retreated 3.56 km while Wykeham Glacier retreated 1.01 km. The Trinity-Wykeham Glacier confluence retreated 4.45 km which was primarily due to rapid retreat of the ice front between 2009 and 2012. Width averaged retreat of the ice front decreased from 1.38 km to 1.12 km between the ICESat (2003-2009) and CryoSat-2 (2010-2016) observational periods.

4.4. Seasonal changes in velocity

Seasonal (spring, summer and winter) changes in velocity between 2013 and 2016 are shown in Figure 5. The velocity of both Trinity and Wykeham glaciers was consistently highest during spring (Figures 5*d* and 5*h*). Within the lowermost 20 km of Trinity Glacier, winter velocity exceeded summer velocity. In comparison, summer velocity was higher than winter velocity across Wykeham Glacier until 2016, when the glacier exhibited a summer slowdown of ~100 m yr^{-1}. The deceleration anomaly observed at the terminus of Wykeham Glacier in Figure 4*a* is prominent and flows below 100 m yr^{-1} during all seasons. At Trinity Glacier, spring velocities increased by 300 m yr^{-1} between 2013 and 2016, whereas summer velocities decreased by 200-300 m yr^{-1}. At Wykeham Glacier, spring velocities increased by 100 m yr^{-1} between 2013 and 2016, while summer velocities decreased by 50-100 m yr^{-1}. Winter velocities across the lowermost 25 km of Wykeham Glacier increased by 100 m yr^{-1}. In comparison, they remained stable at Trinity Glacier.

[Insert Figure 5 here]

4.5. Grounding line

Estimates of the location of floating ice are shown in Figure 6, where P denotes the hydrostatic flotation depth; negative values indicate possible ice flotation (i.e. the ice is buoyant). We estimate that approximately 6 km^2 and 7.5 km^2 of the ice fronts of Trinity and Wykeham glaciers, respectively, were floating in 2014 (Figure 6*a*). The lowermost 5 km and 4 km of Trinity Glacier and Wykeham Glacier, respectively, have values of P < 500 m, which may be considered regions susceptible to ice flotation. Almost the entire calving front of Wykeham Glacier has values of P < 100 m suggesting the terminus is close to flotation. At the pinning point (see Figure 4*a*) of Wykeham Glacier P = 297 m, whereas the region immediately up-glacier is partially floating (P < 0). This distinction is also made clear by the surface morphology on both glaciers (Figure 6*b*). Regions of the ice front that are currently grounded (P > 0) are coincident with areas where the surface is fractured whereas floating regions have a flatter ice surface topography.

[Insert Figure 6 here]

5. Discussion

5.1. Changes between 2000 and 2016

At both Trinity and Wykeham glaciers, the rate of surface thinning, as measured by changes in surface elevation over time, increased from 2003-2009 to 2010-2016 at the same time as glacier velocity also increased (Table 2). Asynchronous retreat of the ice front between 2000 and 2016 has led to the separation of Trinity Glacier and Wykeham Glacier, but both continue to flow through their individual valleys. The simultaneous thinning, acceleration and ice front retreat at Trinity Glacier are indicative of a dynamic thinning mechanism for glacier change, which is likely to also be influenced by factors such as surface melting and subaqueous mass loss from any floating marginal areas. The surface mass balance of the POW as a whole remained stable until recently, when surface melt enhanced mass loss from the ice field (Mair et al., 2009; Noel et al., 2018). The increase in surface melt will also drive a component of the thinning observed here. While a similar mechanism is likely to be responsible for the changes at Wykeham Glacier, a bedrock pinning point at the terminus causes the glacier to redistribute ice into two separate flow units (Figures 3 and 4).

Table 2. Surface elevation change and TWG velocity for three separate time ranges, taken from independent estimates and this study.

Sensor	**Time Period**	**Max annual velocity at end of time range (m yr^{-1}) – Trinity Glacier**	**Max annual velocity at end of time range (m yr^{-1}) – Wykeham Glacier**	**Surface Elevation Change (m yr^{-1})**	**Reference**

Airborne	1995-2000	~ 600	~ 250	~ -0.48	Abdalati et al. (2004)
ICESat	2003-2009	~ 700	~ 300	~ -4	This Study
CryoSat-2	2010-2016	~ 850	~ 500	~ -6	This Study

Thinning of Trinity Glacier is broadly consistent with previous findings (Gardner et al., 2011; van Wychen et al., 2016), but our higher spatial and temporal sampling enables us to observe the spread of thinning inland along the lowermost 20 km of Trinity Glacier, as well as splitting of ice flow at the terminus of Wykeham Glacier. Swath-mode processing of SARIn CryoSat-2 data enables quantification of elevation changes at higher spatial resolution compared to real-beam radar altimeters (Foresta et al., 2016), whereas previous studies used overlapping airborne surveys and single DEMs to quantify d*h*/d*t* (van Wychen et al., 2016; Mortimer et al., 2018). For example, van Wychen et al. (2016) showed thinning rates at the TWG along OIB tracks (see Figure 1 for their location) by comparing them to a 2008 satellite-derived DEM which captures the broad pattern of thinning across the TWG terminus but does not observe the detailed spatial extent of d*h*/d*t* that CryoSat-2 swath processing provides.

The current study extends previous analyses of ice flow change at the TWG (van Wychen et al., 2016; Millan et al., 2017) by using Landsat 7, Landsat 8, and ASTER image pairs with a larger temporal baseline (~365 days). ASTER imagery fills data gaps between 2003 and 2012 due to the inability of COSI-corr to track features across Landsat 7 Scan Line Corrector (SLC) errors (Heid and Kaab, 2012). The total uncertainty, however, is greater than that associated with data from Landsat, most likely due to the lower accuracy of stereo imagery. The use of Landsat 8 data improves upon results derived using data from other optical sensors due to its 16-bit radiometric resolution (Fahnestock et al., 2016), resulting in reduced motion errors and the ability to better estimate seasonal velocity variations due to its shorter revisit time. Thus, the different methodologies used to measure surface velocity at the TWG in this study confirm previous observations of ice flow acceleration at the TWG (van Wychen et al., 2016; Millan et al., 2017) and the presence of rapid annual ice discharge into Nares Strait and Smith Sound.

5.2. Flotation of the TWG

A key result from this study is that both Trinity and Wykeham glaciers appear to be floating at their termini. At Trinity Glacier, most of the calving front is floating or near to floating, whereas the terminus of Wykeham Glacier displays a pattern analogous to a grounding zone (Fricker et al., 2009). At the front of Wykeham Glacier, a local pinning point is coincident with the position of a velocity minimum (<100 m yr^{-1}) and grounded ice. This suggests that as Wykeham Glacier flows into the region of elevated topography, the ice becomes compressed, decelerates and becomes grounded due to the reduction in water depth at this point. The flow of ice redistributes its mass around a bedrock bump as a result of mass conservation (Morlighem et al., 2011), causing the local reduction in ice thickness. The region immediately behind this grounded terminus is floating and is coincident with a flat ice surface indicative of low basal traction. The establishment of this grounding zone produces a backstress on the ice flowing into it and reduces ice discharge, which partially accounts for the different patterns of dh/dt and du/dt we observe at the termini of Trinity Glacier and Wykeham Glacier, respectively. Our measurements of du/dt at the deceleration anomaly differ from Millan et al. (2017) and van Wychen et al. (2014, 2016) as their early-ablation season imagery captures small velocity increases that are superimposed on the annual mean velocity estimate.

The exact cause of flotation is beyond the scope of the present study, but we suggest three possible influences. Firstly, retreat of Trinity Glacier began in 2005 as its northern margin became detached, and eventually separated, from Talbot Glacier (Figure 1). Such lateral disconnection would have reduced the local ice flux and caused a stress imbalance at the ice front. This reduction in ice flux reduces the ice thickness and may have induced buoyancy. Flotation of Trinity Glacier in 2014 may be a response to sustained thinning imposed on an ice front that has stabilised (Figure 4*e*) and thus cannot compensate for increases in ice discharge. Secondly, retreat of both glaciers into a region further below sea level (i.e. a reverse sloping bed) may again have led to an ice thickness that enables flotation. Thirdly, the relative effects of surface and submarine melting cannot be discounted, but their impact on ice thickness cannot be accurately determined here, although summer melting across the QEI appears to be high (Sharp et al., 2011; Mortimer et al., 2018). Flotation increases the area exposed to basal melting and reduces basal friction, both of which are likely to be dominant forcing mechanisms for current rates of thinning and acceleration at the TWG.

5.3. Subglacial controls on dynamic thinning

Our results suggest that the changes in glacier dynamics we have observed at both Trinity Glacier and Wykeham Glacier are strongly controlled by their subglacial topography. In particular, the bed of the TWG is grounded below sea level up to 40 km inland (Figure 3), suggesting changes at the front of both glaciers could propagate rapidly inland – this process appears to have begun (see Figures 2, 3, 4 and 5). For example, the disconnection of Trinity Glacier from the neighbouring Talbot Glacier (see

Section 4.2) was enhanced by the presence of a subglacial trough (Trough #2), acting to channelize the flow of ice to this northern region and further enhance the acceleration of ice flow. Thus, the initiation of retreat was caused by external forcing factors, but the subsequent changes appear to be driven by the subglacial topography.

The subglacial topography of Trinity Glacier appears more streamlined in comparison to the more irregular bed of Wykeham Glacier (Figure 3), where a set of subglacial ridges aligned perpendicular to ice flow reduces the ice flux. Ice becomes compressed when it flows into these ridges and causes local thickening upstream of these obstacles. Locally, this leads to a greater driving stress and faster flow on the down-glacier side of the ridge, causing extensional flow and thinning (i.e. dynamic thinning). However, the overall effect of these ridges is to increase the roughness of the bed and thus enhance the effect of basal friction on the flowing ice mass, hence the annual velocity is lower than Trinity Glacier. The pinning point at the front of Wykeham Glacier further complicates the pattern of glacier dynamics and has led to the establishment of two separate flow units. Rates of thinning on the northern tongue decreased from ~5-6 m yr^{-1} (2003-2009) to ~2-4 m yr^{-1} (2010-2016) due to the flow of ice on a bed that rises above sea level, causing a local reduction in ice thickness and flux. Sustained thinning south of the deceleration anomaly is most likely related to the divergence of ice flow southwards through the terminus overdeepening.

5.4 Seasonal changes at the TWG

Surface melting has been shown to have a strong influence on intra-annual ice flow variations at several glaciers in the QEI (Bingham et al., 2003; Pimentel et al., 2017) but no such influence has so far been detected at the TWG. Our new seasonal velocity results (Figure 5) suggest that the summer slowdown of both Trinity Glacier and

Wykeham Glacier is due to the effective drainage of subglacial meltwater in response to increased meltwater input from the surface. To investigate this effect further, we analysed the distribution of supraglacial lakes on the surface of the TWG to assess the timing of possible lake drainage events. The evolution and drainage of four lakes on the surfaces of Trinity and Wykeham glaciers is shown in Figure 7. The lakes highlighted here form across highly crevassed surfaces where subglacial ridges are present, suggesting lake drainage events are intimately linked to the bedrock topography. Meltwater that is present in crevasses can drain to the bed once a threshold of water pressure is passed (Benn et al., 2007); thus the absence of surface meltwater in the latter images of Figure 7 suggest they have drained to the bed. We do not find evidence for drained lakes before June which implies that these drainage events occur concurrently with summer velocity minima. This pattern is indicative of channelization of the subglacial hydrological system due to enhanced drainage of surface meltwater to the bed. Channelization of the subglacial hydrological system allows efficient evacuation of subglacial meltwater and reduced basal slip and this is likely driving the changes in ice flow during the summer at the TWG. This effect may be enhancing due to the slowdown of summer velocities from 2013 to 2016, although our short time series cannot confirm this.

[Insert Figure 7 here]

The influence of glacier hydrology has also been observed in other regions of the QEI. For example, Bingham et al. (2006) found that John Evans Glacier, to the north of the TWG on Ellesmere Island (see Figure 1 for its location), responded rapidly to supraglacial lake drainage events and enhanced its ice flux due to the storage of meltwater at its bed. Further, meltwater-induced acceleration events may occur at other tidewater glaciers in the QEI (Pimentel et al., 2017) but the effects of ice melange at the glacier terminus are also suggested to be important in modulating long-term seasonal ice flow changes. Meltwater that cannot be evacuated efficiently from the bed may be stored during winter (Chu et al., 2016) and could provide a mechanism for the enhanced winter velocity we observe at the TWG. In comparison, the mechanisms involved in enhancing the velocity of the TWG during spring are more difficult to explain. We suggest the most plausible mechanism is the reduction of backstress at the ice front, which may be induced by weakening of sea ice and melange or enhanced subaqueous melt between April and June when the ablation season begins (Wang et al., 2005).

5.5. Factors affecting future changes to the TWG

The strong dynamic thinning signal over Trinity Glacier compared to other glaciers in the QEI mirrors the pattern of enhanced low-elevation thinning in the ablation zone of the Greenland Ice Sheet (Pritchard et al., 2009), and the low-relief bed topography appears to enhance this effect. Rates of thinning at the TWG are an order of magnitude greater than the background rate of 0.38 m yr^{-1} for all glaciers across the QEI between 2003 and 2009 (Gardner et al., 2011), suggesting the dynamic behaviour of both glaciers has a strong influence on local thinning rates. Future changes at the TWG are likely to be influenced by (1) lateral and basal topography, (2) seasonal changes in melt and ice flow related to atmospheric forcing, and (3) enhanced submarine melting in response to an ungrounded terminus.

We have shown that subglacial topography strongly influences the current rate of dh/dt and du/dt. Insights from the recent pattern of velocity change at Trinity Glacier suggest it will continue to accelerate in the future, and the streamlined nature of its bed that lies below sea level is likely to intensify this effect further. The presence of ridges below Wykeham Glacier forms regions of overdeepened bedrock that can initiate rapid frontal retreat when the glacier retreats on a reverse bed slope, but equally can stabilise the glacier as it retreats on an uphill bed. Retreat of the deceleration anomaly towards the southern tongue of Wykeham Glacier may initiate this retreat pattern. Secondly, if summer warming continues in the coming decades (Serreze and Francis, 2006; Mortimer et al., 2016), enhanced surface meltwater production may influence seasonal velocity variability at the TWG. We have shown (see Figure 5) summer velocity minima which we infer to be a response to enhanced meltwater drainage to the bed of the TWG. If meltwater production increases, we may observe a lengthening of the ablation season which will enhance summer velocity minima but also lengthen spring velocity maxima. Thirdly, where the TWG is floating it is more susceptible to melt undercutting from both oceanic and freshwater sources. Rignot et al. (2015) showed that subglacial meltwater plumes can erode the base of a tidewater glacier and enhance sub-surface melting via subglacial meltwater extrusion. Melt undercutting can also occur due to the intrusion of warm ocean waters beneath the glacier, which Millan et al. (2017) suggest may have initiated the velocity increase at the TWG. Both of these effects remain unresolved and require additional data and analysis to constrain their effects.

6. Summary

This study utilizes near-concurrent airborne geophysical surveys in 2014 to accurately determine the subglacial topography of the Trinity-Wykeham Glacier system (TWG) on Ellesmere Island in Arctic Canada. Triangular interpolation of point elevation measurements from NASA's ICESat laser altimeter (2004-2009) (Pritchard et al. 2009) and swath-mode processing of ESA's CryoSat-2 SARIn mode (Foresta et al., 2016; Gourmelen et al., 2018) across the QEI's variable topography (2010-2016) were used to estimate rates of surface elevation change (dh/dt). Annual and seasonal ice flow changes were assessed by quantifying displacement between pairs of Landsat and ASTER satellite image pairs using the COSI-corr feature-tracking software (Leprince et al., 2007). Ice front change was measured by digitising Landsat images and comparing the locations of successive glacier terminus positions. Regions of glacier flotation were predicted using the principle of hydrostatic equilibrium.

Rates of thinning increased from 4 m yr^{-1} in 2009 to 6 m yr^{-1} in 2016 across the region of the TWG terminus which is grounded below sea level (40 km inland). Simultaneously, annual mean glacier velocities at Trinity Glacier and Wykeham Glacier doubled, which is likely due to an increase in peak flow rates during spring. The spatially coherent flow increase and thinning observed at Trinity Glacier is enhanced by a low relief bed topography, while a similar dynamic thinning effect at Wykeham Glacier is modulated by subglacial ridges that redistribute the flow of ice to the northern and southern sections of the terminus. We also suggest that both marine glaciers fronts are now floating, which could lead to enhanced dynamic thinning and retreat in the near-future. While the origin of these changes remains unresolved, comparisons with regional glacier changes suggest that elevated summer air temperatures have an important effect on rates of ice discharge. However, our results show that subglacial geometry exerts a first order control on the nature of the dynamic changes. The high-resolution bedrock topography presented here will be useful for modelling the TWG system in order to improve our understanding of how the bedrock topography will influence future ice dynamics.

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Conflicts of Interest

The authors declare no conflicts of interest

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Figures

Figure 1 Location of the (*a*) Queen Elizabeth Islands (QEI), (*b*) Prince of Wales Ice Field (POW) and (*c*) the Trinity-Wykeham Glacier system (TWG) which drains the POW. The TWG catchment covers 3,046 km^2 (taken from version 5.0 of the Global Land Ice Measurements from Space (GLIMS) (Bolch et al., 2014) and updated to the 2016 ice front position). Blue lines on (*c*) are the Operation IceBridge (OIB) flight lines and the red survey lines are from the Canadian Arctic Geophysical Exploration (CAGE) survey. Background image in panel (*c*) is a true colour Landsat image from 5 May 2014.

Figure 2 Surface elevation changes (d*h*/d*t*) from the lower parts of Trinity and Wykeham glaciers from (*a*) ICESat between 2003 and 2009 and (*b*) CryoSat-2 between 2010 and 2016. The northern and southern tongue of Wykeham Glacier are annotated on (*a*) but are also applicable to (*b*). (*c*) A graph showing dh/dt across transect A-A' (highlighted in (*b*)) shows rates of thinning increasing from the ICESat to the CryoSat-2 study period. Panel (*a*) is underlain by an ASTER image from 14 June 2009 and panel (*b*) is underlain by a Landsat 8 panchromatic image from 29 June 2016.

Figure 3 Bedrock topography derived from Natural Neighbour interpolation of the CAGE-OIB ice thickness measurements and subtracted from the ArcticDEM of ice surface elevation. Annotations describe key geomorphological features of the subglacial topography. Dashed lines show 100 m elevation contours and the bold line represents sea level (0 m). The CAGE-OIB flight lines are superimposed in light grey. The background image is a Landsat 8 natural colour image from 5 May 2014.

Figure 4 Annual velocity maps over the TWG in (*a*) 2000 and (b) 2016. The ice front position in 2000 (dashed black line) and 2016 (solid black line) are shown in (*b*). Velocity change (d*u*/d*t*) is shown for (*c*) the ICESat observation period (2003-2009) with a standard error of 3.56 m yr^{-1} and (*d*) the Cryosat-2 observation period (2010-2016) with a standard error of 4.97 m yr^{-1}. (*a*) is underlain by a Landsat 7 image from 16 June 2000, panel (*c*) is underlain by an ASTER image from 14 June 2009, and panels (*b*) and (*d*) are underlain with a panchromatic Landsat 8 image 29 June 2016. Ice front change for each year 2000-2016 relative to 30 July 2000 are shown in panel (*e*) for Trinity Glacier, Wykeham Glacier and their confluence (profiles indicated on panel (*b*)).

Figure 5 Seasonal velocity estimates between 2013 and 2016 for Trinity Glacier (*a-d*) and Wykeham Glacier (*e-h*). Spring (April to June) velocity estimates are shown in panels (*a*) and (*e*). Summer (July to September) velocity estimates are shown in panels (*b*) and (f). Winter velocity (October to March) estimates are shown in panels (*c*) and (*g*). Panels (*d*) and (*h*) shows seasonal velocity along each glacier averaged between 2013 and 2016.

Figure 6 (*a*) Gridded surface of the hydrostatic flotation depth (*P*) restricted to showing those areas most susceptible to flotation (i.e those with a value below 500 m). (*b*) Annotated diagram of the TWG ice front showing areas of high crevassing and those with a smooth surface, which may be related to flotation of the TWG terminus. Both figures are underlain with a true colour Landsat 8 image from 5 May 2018.

Figure 7 Supraglacial lake drainage events on the surface of both (*a-d*) Trinity Glacier and (*e-h*) Wykeham Glacier. Each image is a pan-sharpened true colour Landsat 8 image from 2016. The dates are shown for each panel. Blue regions of each true colour image are regions of surface meltwater accumulation.