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Key Points:
- Surface velocities of ice masses in the Queen Elizabeth Islands are mapped
- Dynamic discharge is estimated and its importance to total mass loss assessed
- Queen Elizabeth Islands dynamic discharge is compared with other Arctic regions

Supporting Information:
- Readme
- Supplementary Figure Captions
- Table S1
- Table S2
- Figure S1
- Figure S2
- Figure S3
- Figure S4
- Figure S5
- Figure S6
- Figure S7

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Abstract
Recent studies indicate an increase in glacier mass loss from the Canadian Arctic Archipelago as a result of warmer summer air temperatures. However, no complete assessment of dynamic ice discharge from this region exists. We present the first complete surface velocity mapping of all ice masses in the Queen Elizabeth Islands and show that these ice masses discharged ~2.6 ± 0.8 Gt a⁻¹ of ice to the oceans in winter 2012. Approximately 50% of the dynamic discharge was channeled through non surge-type Trinity and Wykeham Glaciers alone. Dynamic discharge of the surge-type Mittie Glacier varied from 0.90 ± 0.09 Gt a⁻¹ during its 2003 surge to 0.02 ± 0.02 Gt a⁻¹ during quiescence in 2012, highlighting the importance of surge-type glaciers for interannual variability in regional mass loss. Queen Elizabeth Islands glaciers currently account for ~7.5% of reported dynamic discharge from Arctic ice masses outside Greenland.

1. Introduction and Study Area
The Queen Elizabeth Islands (QEI; Devon, Ellesmere, and Axel Heiberg Islands; Figure 1 inset) contain ~104,000 km² of glacier ice, which represents 25% of Arctic glacier ice outside the Greenland Ice Sheet. Long-term surface mass balance records from this region indicate that prior to the late 1980s the ice masses were largely in balance. Since the late 1980s, however, and especially since 2005, changing summer atmospheric circulation patterns have increased advection of warm air from the northwest Atlantic to the Canadian High Arctic, leading to increased surface melt, longer melt seasons, and increased anticyclonic air circulation over the QEI in summer. As a consequence, ice loss from the QEI via surface mass balance has increased sharply in recent years. However, little is known about the importance of dynamic discharge as a mass loss mechanism in the QEI. This study provides the first complete mapping of surface velocities of all ice masses in the QEI and combines this with estimates of ice thickness near the termini of tidewater glaciers to estimate dynamic ice discharge to the ocean. Here we define dynamic ice discharge as the mass of ice passing through a terminus flux gate (also called iceberg flux in some studies); we do not account for the effects of terminus advance or retreat on dynamic ice discharge.

2. Methods
2.1. Determination of Surface Motion
Ice surface motion was determined using a custom-written MATLAB speckle-tracking algorithm applied to RADARSAT-2 fine (resolution: ~8 × 8 m) and ultrafine (resolution: ~3 × 3 m) imagery acquired in 24 day pairs between January and May 2012 (supporting information Table S1). This algorithm tracks the relative displacement of small image chips by applying a two-dimensional cross-correlation algorithm to accurately co-registered RADARSAT-2 image pairs (co-registration was completed using an area cross-correlation technique). This method has previously proved effective for measuring surface ice displacements at the ice cap scale within the Canadian High Arctic. Image acquisitions were restricted to middle to late winter due to the requirement for good coherence. Displacements were determined in both azimuth and range.
Figure 1
directions using chip sizes of \(-450\) m in azimuth and \(-350\) m in range for fine beam imagery and \(-150\) m in azimuth and \(-125\) m in range for ultrafine beam imagery. The 1:250,000 version of the Canadian Digital Elevation Dataset, Level 1, was used to remove the topographic component of the slant-range displacement. To remove systematic biases due to inaccuracies in the baseline or squint effects between image acquisitions, displacements were calibrated using manually selected areas of known zero motion, such as bedrock outcrops, to determine the local bias, which was then removed from the rest of the data set [Gray et al., 2001]. Final displacements were converted to annual velocities. Displacement editing and filtering of the displacements followed the criteria of Van Wychen et al. [2012]: (1) surface ice velocities should be faster along glacier centerlines than at glacier margins; (2) flow direction should follow topography and surface flow features; and (3) flow vectors should not deviate dramatically from adjacent vectors in direction or magnitude over short distances. Filtering was performed manually using ArcGIS™ 9.2, and incorrect matches were removed from the data set. The point velocity results were interpolated to 100 m grid spacing for the entire glaciated region of the QEI using an inverse distance-weighting method with a fixed 500 m search radius.

Output raster grids were mosaicked to produce a single velocity map for each major ice cap. In areas where rasters overlapped, the minimum overlapping value was used to provide a conservative estimate of ice velocity. To assess the consistency between results determined from different image pairs, we compared the absolute difference between overlapping raster pixels at \(-6\) million locations and determined a mean difference of 5.95 m a\(^{-1}\) and standard deviation of 15.5 m a\(^{-1}\). The final raster was clipped using glacier outlines provided in version 3.0 of the Randolph Glacier Inventory (RGI) [Arendt et al., 2012] for the Canadian High Arctic. Due to our use of winter imagery, which does not capture summer speedup events, we assume that velocity maps and calculated discharges represent minimum annual estimates. Comparison of continuous summer and winter differential Global Positioning System (dGPS) records from the terminus region of the tidewater-terminating Belcher Glacier, Devon Ice Cap, indicates that extrapolation of winter-only velocities may underestimate annual velocities by \(-10\)–\(-15\)% (B. Danielson, personal communication, 2013).

To establish motion errors, we assumed that the mean velocity calculated over marginal bedrock outcrops provides information about the error limits of the method. Based on a total of \(-11\) million measurements, mean displacement over bedrock was 5.9 m a\(^{-1}\) with a standard deviation of 6.3 m a\(^{-1}\) (error for individual ice masses is presented in Table S2). To determine error throughout the interior regions of ice caps, where bedrock control was not available, displacements were extracted along ice divides where surface motion should be nearly zero [Raymond, 1983]. Major ice divides were derived from version 3.0 of the RGI and are shown as white lines in Figure 1. Mean velocities along ice divides provided an overall mean error of 6.8 m a\(^{-1}\) and standard deviation of 3.9 m a\(^{-1}\) (Table S2).

### 2.2. Calculation of Dynamic Ice Discharge

This study utilizes airborne-radar measurements of ice thickness collected by NASA’s Operation Icebridge campaigns over the Canadian High Arctic in May 2012 (for 56% of glaciers) and 2006 (5% of glaciers) [Gogineni, 2012], the Scott Polar Research Institute in 2000 (32% of glaciers) [Dowdeswell et al., 2004], and by the University of British Columbia in 1981 (5% of glaciers) [Narod et al., 1988]. The fluxgate for the Good Friday Bay Glacier on Axel Heiberg Island is located \(-20\) km from the calving front and uses a centerline ice thickness estimated using an area-depth scaling scheme [Ommanney, 1969]. This is the only glacier that relies on this data set and this thickness is ascribed a 40% uncertainty based on comparison of ice thickness estimates provided by Ommanney [1969] for four different locations on Axel Heiberg Island with nearby (within \(1.5\) km) ice thickness measurements from NASA’s Operation Icebridge in 2006.

To calculate dynamic ice discharge, both cross-section (20 of 41 glaciers, primarily located on Devon Ice Cap) and centerline methods (21 of 41 glaciers) were used depending on the availability of ice thickness data sets (Table S1). For glaciers with depth measurements acquired perpendicular to ice flow, the cross-section method was used. In such cases, the terminus width was divided into a number of evenly spaced columns depending on the spacing between ice thickness data points unique to each acquisition and post-processing

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**Figure 1.** Surface velocities for (a) Northern Ellesmere Icefield; (b) Agassiz Ice Cap; (c) Prince of Wales Icefield; (d) Manson Icefield; (e) Sydkap Ice Cap; (f) Müller and Steacie Ice Caps; and (g) Devon Ice Cap. Major ice divides used to determine errors are denoted in white. Note the difference in velocity scale between images, although ice discharge scale remains constant (base image: Moderate Resolution Imaging Spectroradiometer, 4 July 2011). Inset map: study site.
method (typically 20–30 m). For each column, upper and lower estimates of ice flux were calculated by assuming different fractional contributions of internal ice deformation to the measured surface velocity. For the lower ice flux estimate \( Q_{\text{min}} \) we assume a depth-averaged velocity of 80% of the measured surface velocity (i.e., 20% of the overall motion is accounted for by internal deformation) [Paterson, 1994]. For the upper estimate \( Q_{\text{max}} \) we assume that the depth averaged velocity is equal to the surface velocity and that ice movement is by basal sliding alone. Discharge was then calculated for each column width \( W \) using

\[
Q_{\text{min}} = (0.8 \times (V - V_{\text{error}})) \times (H + H_{\text{error}} + (T \times E)) \times (W)
\]

\[
Q_{\text{max}} = (V + V_{\text{error}}) \times (H + H_{\text{error}} + (T \times E)) \times (W)
\]

where \( V \) is surface ice velocity, \( V_{\text{error}} \) is a constant error derived separately for each ice mass from the velocity errors computed along ice divides (i.e., highest measured error as indicated in Table S2), \( H \) is measured ice thickness, \( H_{\text{error}} \) is the error associated with each individual ice thickness data set (Table S1), \( T \) is the average annual terminus elevation change as determined from ICESat data in the QEI between 2003 and 2009 [Gardner et al., 2011], and \( E \) is the number of elapsed years between the acquisition of the ice thickness data and 2012. The ICESat data set is the most complete elevation change data set available for this region and enables a correction to be made for the change in ice thickness between the time an ice thickness measurement was made and 2012. Total \( Q_{\text{min}} \) and \( Q_{\text{max}} \) for a cross section were then calculated from the sum of the individual column widths and provide error limits on our estimates.

For glaciers for which only centerline measurements of ice thickness were available, the cross-sectional flux gate was assumed to have a “U” shape, based on the morphology of tidewater glaciers in the QEI for which cross sections are known [e.g. Gogineni, 2012] and modeling of valley form after long periods of erosion [Harbor, 1992]. This “U” shape was modeled based on

\[
H_{\text{interpolated}} = \left( \frac{(10-C)}{(D_2^2)} \right) \times (D_1^2) + C
\]

where \( H_{\text{interpolated}} \) is the assumed ice thickness using a parabolic interpolation from the measured ice thickness at the centerline to a marginal ice thickness of 10 m, \( C \) is the measured centerline ice thickness, \( D_1 \) is the distance from the centerline to the glacier margin, and \( D_2 \) is the distance from the centerline to the center of the interpolated ice column. Ice column thickness was interpolated at 20 m intervals from the centerline to the margins. To assess the quality of the modeled cross-sectional glacier geometries, we determined the difference between true glacier cross-sectional geometry and the cross-sectional geometry modeled from only a single point at the glacier centerline for all glaciers within the QEI with depth measurements collected perpendicular to ice flow. This analysis indicated that the centerline method of cross-sectional area modeling underestimated the true cross-sectional area by ~12%. We therefore multiply the modeled cross-sectional area by 1.12 to correct for this bias. Minimum and maximum discharges for each column were then calculated from the following equations:

\[
Q_{\text{min}} = (0.8 \times (V - V_{\text{error}})) \times (H_{\text{interpolated}} \times 1.12 - H_{\text{error}} + (T \times E)) \times (W)
\]

\[
Q_{\text{max}} = (V + V_{\text{error}}) \times (H_{\text{interpolated}} \times 1.12 + H_{\text{error}} + (T \times E)) \times (W)
\]

Total \( Q_{\text{min}} \) and \( Q_{\text{max}} \) for a cross section were calculated from the sum of the individual column fluxes and provide error limits on our estimates. Reported dynamic ice discharge values are the average of \( Q_{\text{min}} \) and \( Q_{\text{max}} \) with \( Q_{\text{min}} \) and \( Q_{\text{max}} \) providing the lower and upper uncertainty bounds, respectively.

3. Results and Discussion

Glacier velocities in the Queen Elizabeth Islands are generally relatively low (Figures 1 and S1–S7). In the interior regions of ice caps and ice fields, surface velocities are typically < 10 m a\(^{-1}\), indicative of ice frozen to its bed and moving by ice deformation alone. Most tidewater glaciers have velocities of ~30–90 m a\(^{-1}\) (interquartile range) along their main trunks, rising to < 300 m a\(^{-1}\) at the terminus, while land-terminating glaciers typically have velocities ~20–50 m a\(^{-1}\) (interquartile range) along their main trunk, rising to < 75 m a\(^{-1}\) near the glacier terminus. For land-terminating glaciers, this pattern is different to the peak in velocity that is typically expected near the equilibrium line altitude (ELA) and could be due to the fact that these glaciers tend to have large accumulation areas that drain into relatively narrow valleys and/or because they change from cold based to warm based between their accumulation and ablation areas [Burgess et al., 2005;
Table 1.  Pan-Arctic Comparison of Ice Discharge Estimates

<table>
<thead>
<tr>
<th>Region</th>
<th>Period</th>
<th>Ice Discharge (Gt a⁻¹)</th>
<th>% of Pan-Arctic Discharge (excluding Greenland)</th>
<th>Data Source</th>
</tr>
</thead>
<tbody>
<tr>
<td>CAA, Canadian Arctic Archipelago</td>
<td>Queen Elizabeth Islands</td>
<td>2012</td>
<td>2.59</td>
<td>7.5</td>
</tr>
<tr>
<td></td>
<td>Baffin and Bylot Islands</td>
<td>1999–2003</td>
<td>0.25</td>
<td>0.7</td>
</tr>
<tr>
<td>Alaska All Glaciers</td>
<td>2006–2010</td>
<td>17.1</td>
<td>49.7</td>
<td>Burgess et al. [2013]</td>
</tr>
<tr>
<td>Russia Academy of Sciences Ice Cap</td>
<td>2003–2009</td>
<td>1.4</td>
<td>4.1</td>
<td>Moholdt et al. [2012]</td>
</tr>
<tr>
<td></td>
<td>Franz-Josef Land</td>
<td>1952–2001</td>
<td>4.3</td>
<td>12.5</td>
</tr>
<tr>
<td></td>
<td>Novaya Zemlya</td>
<td>1952–2001</td>
<td>1.4</td>
<td>4.1</td>
</tr>
<tr>
<td></td>
<td>Severnaya Zemlya</td>
<td>1952–2001</td>
<td>0.75</td>
<td>2.2</td>
</tr>
<tr>
<td>Svalbard All Glaciers</td>
<td>2000–2006</td>
<td>6.75</td>
<td>19.6</td>
<td>Blaszczyk et al. [2009]</td>
</tr>
<tr>
<td>Pan-Arctic glaciers and ice caps (excluding Greenland)</td>
<td>34.54</td>
<td>100</td>
<td></td>
<td>Dowdeswell et al. [2008]</td>
</tr>
</tbody>
</table>

*aCAA, Canadian Arctic Archipelago.

Copland et al., 2003a). The only tidewater glaciers with observed velocities > 300 m a⁻¹ are the Belcher and Fitzroy Glaciers (located in the northeastern sector of Devon Ice Cap (DIC), and the Trinity and Wykeham Glaciers (located in the southeastern region of Prince of Wales (POW) Icefield, Ellesmere Island). Short and Gray [2005] determined velocities of ~800 m a⁻¹ at the terminus of Trinity Glacier and ~400 m a⁻¹ at the terminus of Wykeham Glacier in 2004, which compare to our velocities of ~1200 m a⁻¹ and ~500 m a⁻¹, respectively, in 2012. This suggests a recent flow acceleration, although the moderate velocity increase and lack of other evidence (e.g., looped moraines and new crevassing) suggests that this is unlikely due to surging. The higher velocities of these four glaciers may reflect relatively high local accumulation rates due to proximity to year-round moisture sources in the North Open Water Polynya and Baffin Bay [Mair et al., 2009; Koerner, 1979].

Copland et al. (2003b) determined velocities of 1000 m a⁻¹ along the whole 25 km length of Mittie Glacier in 1999, and Short and Gray [2005] determined flow speeds of up to ~1000 m a⁻¹ at the terminus of Mittie Glacier in winter 2003 and ~700 m a⁻¹ in 2004. By comparison, we measured velocities that were < 10 m a⁻¹ in 2012. This indicates that between 2004 and 2012 Mittie Glacier entered the quiescent phase of its surge cycle and that its surge cycle is at least 5 years in length (and possibly longer). Similarly, the main trunk of Iceberg Glacier, Axel Heiberg Island, was flowing ~100 m a⁻¹ in winter 2004 [Short and Gray, 2005] but was stagnant in 2012. Conversely, velocities determined in 2012 at the front of Good Friday Bay Glacier on southern Axel Heiberg Island are similar to those determined in the 1950s when the glacier was actively surging [Müller, 1969; Copland et al., 2003b], suggesting that this glacier is currently in an active surge phase (although it only has a small impact (0.09 Gt a⁻¹) on total regional dynamic ice discharge even when surging). Only two land-terminating glaciers in the QEI have surface velocities exceeding 75 m a⁻¹ (Chapman and Unnamed6 Glaciers), both of which have previously been identified as surge type [Copland et al., 2003b; Hattersley-Smith, 1969].

Fast-flowing glaciers in the QEI (including surge-type glaciers) typically occupy deep subglacial troughs that channel flow from the upper reaches of the accumulation area toward the margins [Burgess et al., 2005]. Faster-flowing glaciers tend to be located in areas of higher accumulation [Koerner, 1979] and occupy deeply incised terrain, such as is found on eastern DIC, eastern POW, and western Axel Heiberg Island. In contrast, slower-flowing ice occurs in areas of lower accumulation [Koerner, 1979] and where glaciers overlie plateau-like surfaces, such as western DIC, western POW, and eastern Axel Heiberg Island.

Total ice discharge from QEI glaciers is currently 2.6 ± 0.8 Gt a⁻¹ (Table S1). Ice discharge estimates reported here are similar to previous estimates from studies of selected glaciers within this region [Van Wychen et al., 2012; Short and Gray, 2005; Mair et al., 2009; Williamson et al., 2008; Burgess et al., 2005]. A limited number of glaciers account for most of the ice discharge from the QEI, and ~50% of total ice flux is channeled through just the Trinity and Wykeham Glaciers that drain POW Icefield (although this value likely varies temporally due to the frequent occurrence of tidewater glacier surging, or other forms of velocity variability, in the QEI).

Similarly, a small number of glaciers typically dominate the discharge to the ocean from individual ice masses: Belcher and Fitzroy Glaciers (~55% of DIC), Cahn Glacier (~55% from Agassiz Ice Cap), Otto and Yelvertone Bay Glaciers (~80% from Northern Ellesmere Icefields), and Good Friday Bay Glacier (~95% from Steacie and Müller Ice Caps). As a consequence, overall ice discharge to the ocean from the QEI may be highly sensitive to changes in the dynamics of only a few glaciers. To assess this, we combine the 2003 surge speeds of Mittie
Acknowledgments

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Glacier [Short and Gray, 2005] with our ice depth measurements to compute that in 2003 the Mittie Glacier would have contributed 0.90 ± 0.09 Gt a\(^{-1}\) of ice to the ocean, compared to 0.02 ± 0.02 Gt a\(^{-1}\) in 2012. This rapid change in ice flux from a single glacier signifies the importance of surge-type glaciers within the region, which may oscillate between being large and negligible contributors to dynamic ice discharge over relatively short periods (<10 years).

Comparison of our estimate of dynamic discharge with recent estimates of surface mass balance in the QEI provides insight into their relative importance to mass loss (assuming that the dynamic discharge in 2012 is representative of long-term values). The 2012 discharge rates are equivalent to ~3.1% (2.6 versus 83.3 Gt a\(^{-1}\)) of the average total annual runoff ((melt + rain) − refreezing) from the QEI from 2007 to 2009, the most recent period for which data are available [Gardner et al., 2011; A. Gardner, personal communication, 2013]. For the 2004–2006 period, the dynamic discharge would have been equivalent to ~5.2% (2.6 versus 49.6 Gt a\(^{-1}\)) of the average total annual runoff. This suggests that current mass loss in the QEI is dominated by surface melt and runoff, and that dynamic discharge has become a proportionally smaller component of mass loss in recent years. Determining how ice discharge rates within the region may respond to this increase in surface melt and runoff requires further study.

To put the ice discharge rates reported here within a broader context, we compare our estimates with those for other Arctic glaciers and ice caps (GIC) outside of Greenland (Table 1). The QEI currently contributes ~7.5% of total reported dynamic ice discharge from GIC in the Arctic, although the exclusion of Greenland GICs (due to lack of flux data) means that this is likely an overestimate. Nevertheless, Alaska and Svalbard account for the majority of mass loss via dynamic discharge currently reported for the circumpolar Arctic. The large differences in discharge are probably driven by regional variations in accumulation and climate, with more maritime regions experiencing higher rates of mass turnover and thus higher discharge rates (e.g., Alaska, precipitation ~1–5 m w.e. a\(^{-1}\) [Braithwaite, 2005]), while more continental climates experience lower rates of mass turnover and relatively lower discharge (e.g., Canadian High Arctic, precipitation <0.5 m w.e. a\(^{-1}\) [Braithwaite, 2005]).

4. Conclusions

The first complete surface velocity maps of the glaciated region of the QEI reveal a marked asymmetry in flow structure (Figure 1), with higher rates of motion in areas of higher accumulation and deep subglacial troughs, and lower rates in regions underlain by flatter topography and lower accumulation [Koerner, 1979]. Velocities of land-terminating glaciers are usually restricted to velocities of ~20–50 m a\(^{-1}\) (interquartile range) along their main trunk, except on surge-type glaciers in their active phase. Maximum velocities of tidewater-terminating glaciers are ~30–90 m a\(^{-1}\) (interquartile range) along their main trunks and typically rise to <300 m a\(^{-1}\) at their termini. Total dynamic mass losses are currently 2.6 ± 0.8 Gt a\(^{-1}\), about half of which is drained through the Trinity and Wykeham Glaciers of POW Icefield. Although the QEI contains ~25% of Arctic glacier ice outside the Greenland Ice Sheet, it contributes only ~7.5% of the total reported dynamic ice discharge from GIC in the circumpolar Arctic (excluding Greenland). This is largely a function of the QEI's continental climate and lower rate of mass turnover compared to other Arctic regions, together with the fact that large regions of many QEI ice masses terminate on land.

Our results suggest that mass loss by surface melt and runoff currently dominates the loss term in the mass balance of the QEI and that ice discharge has become a proportionally smaller component of that term in recent years. However, changes in tidewater glacier dynamics due to the termination or initiation of surges can rapidly change the rates at which ice is transported to the oceans from the QEI. For example, the 2003 surge of the Mittie Glacier likely increased total regional ice fluxes by ~35%. This oscillation between faster and slower flow necessitates continued annual mapping of ice motion in order to identify all surge-type glaciers in the QEI and quantify the impact of surge cycles on glacier mass balance. Further studies are also required to understand seasonal and long-term variations in motion on non surge-type glaciers in the QEI.

References

