

## Chapter 3    Glaciers, Ice Shelves and Ice Islands

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### Key messages

- Glacier and ice cap mass losses are accelerating across the Eastern Canadian Arctic, with losses over the past decade at least double that of previous decades.
- Most glacier and ice cap mass losses can be attributed to increasing summer air temperatures.
- Canadian Arctic ice shelves have decreased in area by approximately half over the past decade.
- Ice shelves and floating glacier tongues are producing large ice islands and icebergs that can provide significant hazards to offshore oil exploration and shipping companies.
- It is likely that all Arctic ice shelves will be lost by the 2040s or earlier.
- Continued losses for glaciers and ice caps are highly likely for the remainder of the 21st century, resulting in the complete disappearance of many small ice masses.
- Loss of all of these ice features will result in loss of biodiversity, and the complete extinction of globally unique ecosystems that depend on ice shelf, glacial ice and ice cap integrity.

### Abstract

The Eastern Canadian Arctic contains over a third of the world's Arctic glaciers and ice caps, and the last remaining ice shelves in the Northern Hemisphere. These components of the cryosphere provide an important part of the landscape diversity of Nunavut, act as important sentinels of climate change, and provide unique habitats for life living under extreme conditions. Glaciers and ice caps have been losing significant mass in recent decades, with current melt rates at their highest for at least the past 3000 years. Recent glacier mass loss rates in the northern Canadian Arctic Archipelago (CAA) have been five times greater than the 1963-2004 average, while mass loss rates in the southern CAA were more than twice as high in 2003-2011 compared to 1963-2006. These mass losses have been linked to increasing summer air temperatures, and suggest that glaciers and ice caps in the Eastern Canadian Arctic are far out of equilibrium with current climate. On northern Ellesmere Island, the number of ice shelves has halved since 2005, with their area decreasing from >1000 km<sup>2</sup> in 2005 to ~500 km<sup>2</sup> today. These losses are expected to continue in the future, making it likely that all ice shelves will disappear in the next few decades. When ice shelves and floating glacier tongues break up they produce icebergs and ice islands, which can be significant hazards to marine navigation and offshore oil and gas exploration. Ice islands in Baffin Bay typically originate from the breakup of floating glacier tongues in northwest Greenland, while those in the interior islands of the CAA and Beaufort Sea typically originate from the breakup of ice shelves on northern Ellesmere Island.

### 3.1 Introduction

Glaciers and ice caps in the Eastern Canadian Arctic comprise over a third of the ~400 000 km<sup>2</sup> terrestrial ice cover in the Arctic outside of the Greenland Ice Sheet, with ~42 000 km<sup>2</sup> in the southern Canadian Arctic Archipelago (CAA) (Baffin and Bylot Islands) and ~104 000 km<sup>2</sup> in the northern CAA (Queen Elizabeth Islands, consisting of primarily Axel Heiberg, Coburg, Devon, Ellesmere, Meighen and Melville islands) (Figure 1; Sharp et al. 2014). Ice masses are primarily concentrated along coastlines adjacent to moisture sources, of which Baffin Bay is the primary source and the Arctic Ocean a secondary source (Koerner 1979). Most ice caps range in elevation from 0 to ~2000 m above sea level, and flow slowly (<15 m a<sup>-1</sup>) in their interior regions where the ice is frozen to the bed (Van Wychen et al. 2014). In areas where the ice becomes confined in valleys it forms outlet glaciers which have unfrozen beds and flow towards the ocean at average speeds of ~20-200 m a<sup>-1</sup>, although speeds can reach >1 km a<sup>-1</sup> for the very largest glaciers and during surges (Van Wychen et al. 2014, 2016). On average, glacier speeds are higher in the northern CAA than the southern CAA, due to the larger sizes of glaciers in the north and the tendency for more of them to reach the ocean. Where glaciers reach the ocean they are called tidewater glaciers, and they typically produce icebergs. In the Canadian Arctic, ~98% of icebergs originate from the northern CAA and ~2% from the southern CAA (Van Wychen et al. 2015).

In the coldest, northernmost parts of the Canadian Arctic, some glaciers do not produce icebergs when they reach the ocean, but instead become preserved in fiords due to the protection provided by surrounding topography and/or sea ice. Over time, this glacier ice can fill the near-surface of fiords to produce ice shelves, which can be up to ~100 m thick (Mortimer et al. 2012). Ice shelves can also be produced by the *in situ* accumulation of very old sea ice and snow (Jeffries 2002). Studies indicate that ice shelves have occupied some of the fiords of northern Ellesmere Island for the past ~4000-5500 years (England et al. 2008, Antoniades et al. 2011). These ice shelves have undergone rapid disintegration in recent years, decreasing from a

total area of ~8900 km<sup>2</sup> in 1906 (Vincent et al. 2001) to ~500 km<sup>2</sup> in 2015. As ice shelves have disintegrated they have produced ice islands, which are large tabular icebergs with a typically rolling surface topography of troughs and ridges. They can reach many tens of km<sup>2</sup> in area, and most drift for periods of years to decades in the Arctic Ocean and interior islands of the CAA, before breaking up in more southerly waters. For example, the ice island that broke away from the Ayles Ice Shelf in 2005 was roughly the size of Manhattan, and rapidly drifted westwards in the Arctic Ocean after formation, along the coast of northern Ellesmere Island (Copland et al. 2007). These immense, drifting ice features are also created from the break-up of floating glacier tongues. Ice islands located in Eastern Canadian Arctic water bodies, such as Nares Strait and Baffin Bay, most often originate from the floating glacier tongues of northwest Greenland. Ice islands found in the Western Canadian Arctic and interior islands of the northern CAA are typically derived from the break-up of the ice shelves on northern Ellesmere Island, although a few have also been produced from the breakup of floating glacier tongues in this region.

In this report we outline the characteristics and recent changes of the glaciers, ice shelves and ice islands in the Eastern Canadian Arctic. Over the past decade, satellite observations have enabled the monitoring and measurement of changes to many of these features on a regional scale for the first time, and have provided new insight into how these components of the cryosphere are changing. This builds on earlier field-based observations, which were typically more local in nature and collected on the ground or by aircraft. We finish by providing an assessment of how these features are likely to evolve in the future given current and predicted climate change.

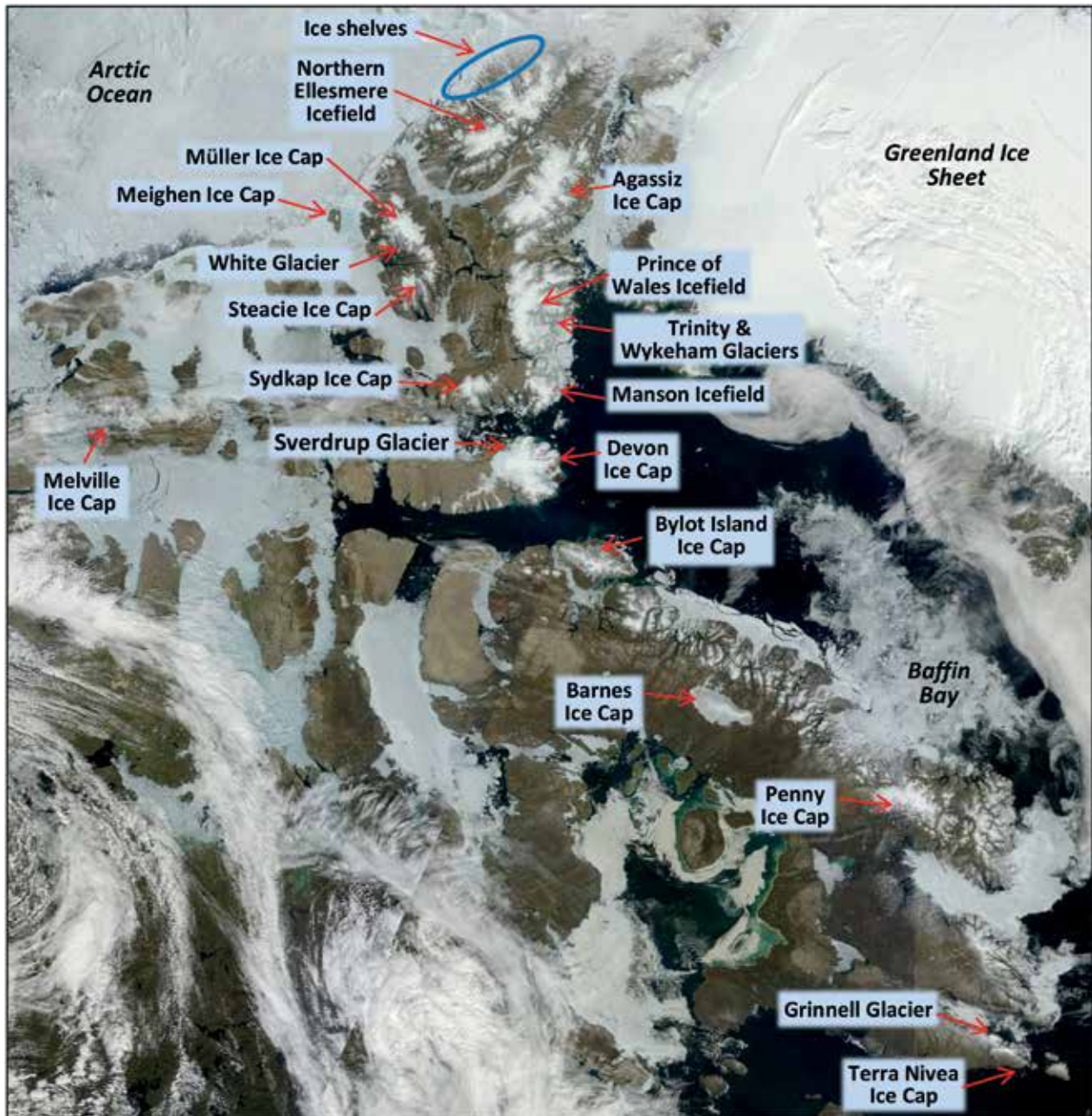


FIGURE 1. Location of the major glaciers, ice caps and icefields of the CAA. Source: MODIS Terra, July 21, 2015. The location of the ice shelves is marked by a blue oval (see Figure 5 for more detail).

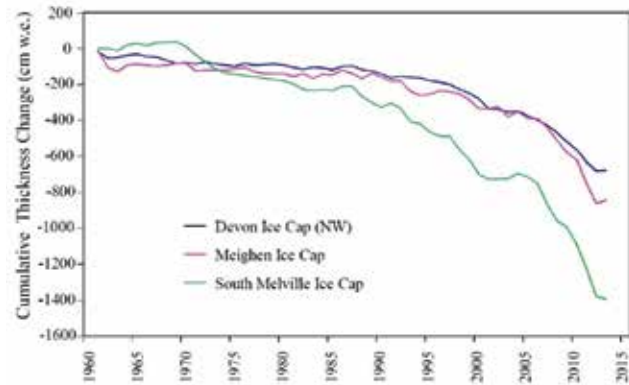


## 3.2 Glaciers and ice caps

### 3.2.1 *In situ* mass balance measurements

The traditional method of determining a glacier's health involves measuring the change in height of a network of mass balance stakes drilled into the glacier surface (Box A). Annual (or more frequent) measurements of the height of these stakes in relation to the glacier surface can then provide information on accumulation and ablation patterns along the glacier length. This 'glaciological' method has been used in the Canadian Arctic since the late 1950s to early-1960s at four monitoring sites (White Glacier, Axel Heiberg Island; Meighen Ice Cap, Meighen Island; Melville Ice Cap, Melville Island; Sverdrup Glacier, Devon Ice Cap, Devon Island) (Koerner 2005, Sharp et al. 2011). This record shows that prior to the late 1980s, the ice masses of the northern CAA were largely in balance, meaning that ablation (primarily melt) was balanced by accumulation (snowfall). However, glacier mass balances became negative in the mid-1990s, and have been acutely negative for most years since 2005. For example, mass loss from 2005-2009 was nearly five times greater than the 1963-2004 average (Koerner 2005, Mair et al. 2009, Sharp et al. 2011). On the northwest sector of Devon Ice Cap, surface mass balance has been negative since 1960 (Koerner 2005), but after 2005 surface melt rates have been ~4 times greater than the long-term average (Sharp et al. 2011). Between 1960 and 2013 there has been a cumulative mass loss of almost 12 m water equivalent (w.e.) on White Glacier (Box A), >6 m w.e. on Devon Ice Cap, >8 m on Meighen ice cap and >14 m on South Melville ice cap (Figure 2; Burgess 2014). Koerner (2005) argued that this negative trend has been driven primarily by warmer summers that have increased melt and the depth of meltwater percolation, rather than by changes in winter snow accumulation.

Ice core records reveal that melt rates on CAA ice caps over the last 25 years are at their highest level in millennia. On Agassiz Ice Cap, Ellesmere Island, ice core records acquired from the summit region indicate that melt rates since the early 1990s are comparable to those last experienced ~9000 years ago when conditions were warmer



**FIGURE 2.** Cumulative water equivalent (w.e.) surface elevation change on Devon, Meighen and South Melville ice caps over the period 1960-2013 derived from *in situ* mass balance measurements. Adapted from Burgess (2014), © Government of Canada..

during the Holocene climatic optimum (Fisher et al. 2012). A historical climate record derived from deep and shallow ice cores on Penny Ice Cap, Baffin Island, indicates that current melting there is unprecedented in magnitude and duration for the past ~3000 years (Fisher et al. 1998, 2012). Since the mid-1990s, near-surface temperatures recorded in shallow ice cores in the accumulation area of Penny Ice Cap have increased by ~10 °C due to increased amounts of summer meltwater percolating into the ice (Zdanowicz et al. 2012). On Devon Ice Cap, significant warming between 2007 and 2012 has resulted in an increase in the amount of ice in the higher elevation regions of the ice cap, replacing the previous firn (snow more than 1 year old), and resulting in reduced vertical percolation into deeper regions and reduced water storage potential of much of the firn reservoir (Gascon et al. 2013).

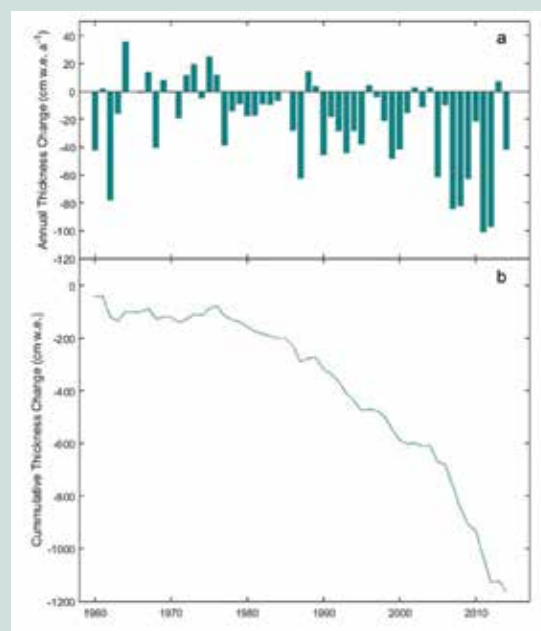
**BOX A.** Mass balance measurements at White Glacier

White Glacier (Figure A1) is a 14 km long mountain glacier, measuring approximately 40 km<sup>2</sup> in area, located on western Axel Heiberg Island, Nunavut (Figure 1). The mass balance monitoring program at White Glacier was initiated in 1959 by Fritz Müller, founder of the McGill Arctic Research Station (Müller 1963), and today it is one of 40 official reference glaciers within the Global Terrestrial Network for Glaciers through the United Nations Framework Convention on Climate Change. These observations, submitted annually to the World Glacier Monitoring Service ([www.wgms.ch](http://www.wgms.ch)), are used with others to calculate a worldwide glacier mass balance index that is regularly published in climate change assessment reports (WGMS 2008, 2015). As of 2015 the mass balance record (Figure A2) indicates that, on average, increased melt in recent years has not been offset by annual snowfall, resulting in mass loss and glacier retreat. This trend coincides with a raising of the equilibrium line altitude (ELA; the transition from regions of annual mass gain to regions of annual mass loss; Figure A1) by approximately 180 m over the past two decades (Thomson et al. 2017). The mean ELA over the period 2004–2014 was 1228 m above sea level (a.s.l.).

White Glacier's rich history of previous research provides valuable baseline data that enables us to observe in detail how Arctic glaciers are responding to climate change. Field studies are carried out in the spring (April and May) when we can access the glacier by snowmobile from our base at the nearby McGill Arctic Research Station. Our fieldwork involves measuring snow accumulation and ice melt along a transect of 40 mass balance stakes installed up the glacier centerline (Figure A1), spanning elevations from 110 to 1520 m a.s.l.. Being situated in a polar desert, annual snow depths rarely exceed 1.5 m at upper elevations (1500 m a.s.l.), and we make density measurements of the snowpack to calculate its water equivalence (Cogley et al. 1996). Summer ice melt near the terminus typically ranges between 2–3 m, but may exceed 5 m in high-melt summers (e.g., 2012). Recent results from a mapping campaign in 2014 indicate that glacier ice in the terminus region (lower 5 km) has thinned by 20 m on average over the past 55 years, with maximum losses exceeding 50 m (Thomson and Copland 2016).



**FIGURE A1.** View northwards up the White Glacier terminus showing approximate location of the mass balance stake network and contemporary equilibrium line.



**FIGURE A2.** Water equivalent (w.e.) surface elevation change on White Glacier for the period 1960–2014 derived from *in situ* mass balance measurements: (a) Annual measurements; (b) Cumulative changes.

### 3.2.2 Airborne and satellite mass balance measurements

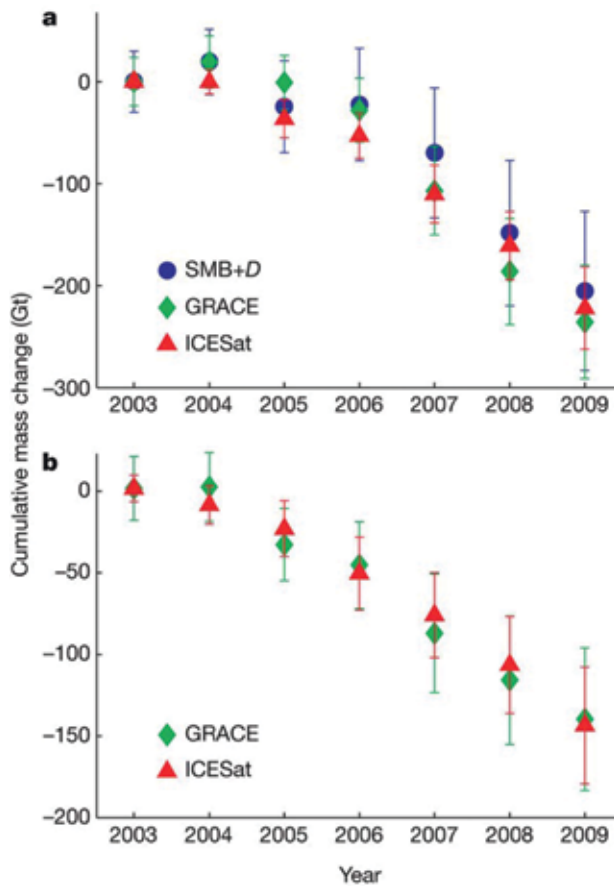
As an alternative to the ‘glaciological’ method of measuring glacier mass balance, recent studies have used airborne and satellite data to measure mass balance using the ‘geodetic’ method. This method relies on repeated measurements of the surface height of glaciers and ice caps, and can provide information on regional changes in mass balance that are not available from local *in situ* measurements. Some of the earliest kinds of these records are provided through repeated laser altimetry measurements made by the National Aeronautics and Space Administration (NASA), which started in 1995 over many CAA ice caps and have continued at approximately 5 year intervals to the present day. Based on thickness changes derived from repeat airborne surveys in 1995 and 2000, Abdalati et al. (2004) reported that most ice caps in the CAA were thinning at lower elevations (<1600 m), but showing thickening or little change at higher elevations. For ice caps in the northern CAA thinning was generally  $<0.5 \text{ m a}^{-1}$ , but on Barnes and Penny ice caps in the southern CAA, thinning was  $>1 \text{ m a}^{-1}$  at lower elevations. On Penny Ice Cap, more recent measurements (2007–2011) indicate thinning of  $3\text{--}4 \text{ m a}^{-1}$  near the ice cap margin, amongst the highest rates anywhere in the Canadian Arctic (Zdanowicz et al. 2012).

Gardner et al. (2012) updated the earlier measurements of glacier changes in the southern CAA by using a combination of repeat airborne and satellite altimetry measurements, satellite gravimetry measurements and digital elevation models derived from stereo air photos and satellite imagery. These measurements indicate that total mass loss from the glaciers of Baffin and Bylot Islands more than doubled from  $11.1 \pm 3.4 \text{ Gt a}^{-1}$  for the period 1963–2006 to  $23.8 \pm 6.1 \text{ Gt a}^{-1}$  for the period 2003–2011. These changes were primarily attributed to increases in summer temperature, with little change in precipitation over this period. Between 2003 and 2011 the glaciers of Baffin and Bylot islands contributed  $\sim 16\%$  ( $0.07 \pm 0.02 \text{ mm a}^{-1}$ ) of the total contribution to sea level rise from glaciers and ice caps outside of Greenland and Antarctica (Gardner et al. 2012). This aligns with observations from passive

satellite microwave records that the average melt season on Barnes Ice Cap lengthened by  $\sim 33\%$  between 1979–1987 and 2002–2010, and nearly doubled on Penny Ice Cap between 1979 and 2010 (Dupont et al. 2012).

For the CAA as a whole, Gardner et al. (2011) used repeat satellite gravimetry, satellite laser altimetry and modelling to show that mass loss rates almost tripled from  $31 \pm 8 \text{ Gt a}^{-1}$  in 2004–2006 to  $92 \pm 7 \text{ Gt a}^{-1}$  in 2007–2009. Losses were extensive across both the northern and southern CAA (Figure 3). Similar to the findings of Gardner et al. (2012), these losses were attributed primarily to warmer summer air temperatures in the lower troposphere, with mass loss rates highly sensitive to relatively small temperature changes, at a rate of  $64 \pm 14 \text{ Gt a}^{-1}$  per  $1^\circ\text{C}$  increase in temperature. These overall losses made the CAA the largest contributor to sea level rise outside of the Greenland and Antarctic ice sheets for the period 2007–2009. More recently, Harig and Simons (2016) used satellite gravimetry data to demonstrate that ice mass losses have accelerated across the CAA since 2003, with the exception of a positive mass anomaly in summer 2013. While the majority of losses has been in the form of surface melt and runoff, Van Wychen et al. (2014) determined that iceberg discharge from the CAA amounted to  $2.6 \pm 0.8 \text{ Gt a}^{-1}$  in 2012, equating to 7.5% of pan-Arctic iceberg discharge and  $\sim 3.1\%$  of the average total glacial runoff in the QEI from 2007–2009.

For glaciers and ice caps, Sharp et al. (2014) calculated area changes using aerial photographs and satellite imagery from  $\sim 1960$  to  $\sim 2000$ . Over this period, total ice-covered area in the northern CAA reduced from  $107\,071 \text{ km}^2$  to  $104\,186 \text{ km}^2$ . The greatest area losses occurred on the ice caps on Devon Island and southern Ellesmere Island, with reductions of 4.0% and 5.9%, respectively. The northern Ellesmere Island ice cap also lost significant area, shrinking by 3.4% over this period. Using a similar technique, Thomson et al. (2011) determined that the ice coverage on Axel Heiberg Island reduced by a total of less than 1% between 1958–59 and 1999–2000. However, this hides the fact that losses were particularly pronounced on small glaciers and ice caps, with retreat of  $\sim 50\text{--}80\%$  for independent ice masses less



**FIGURE 3.** Cumulative change in glacier mass between fall 2003 and fall 2009 for the: (a) northern CAA and (b) southern CAA, derived from modelling of surface mass balance and ice dynamics (SMB+D), satellite gravimetry (GRACE) and satellite altimetry (ICESat). Note the difference in scale between the two parts. Reprinted by permission from Macmillan Publishers Ltd: Nature (Gardner et al. 2011), © 2011.

than 25 km<sup>2</sup> in size, and the complete disappearance of 90% of ice masses smaller than 0.2 km<sup>2</sup>.

In the southern CAA, other studies have also shown that small ice caps and glaciers are particularly vulnerable to losses; e.g., Paul and Svoboda (2009) found total area losses of 12.5% between ~1920 and 2000 for a total of 264 glaciers located on the Cumberland Peninsula, southeast Baffin Island. These losses showed a strong dependence on glacier size, with glaciers 0.1–0.5 km<sup>2</sup> in size losing an

average of 45.6% of their area over this period, compared to mean losses of 6.4% for glaciers >50 km<sup>2</sup> in size. A recent comparison of digital elevation models derived from historical aerial photography and high resolution satellite imagery shows that there have also been dramatic reductions in the volume and area of the Grinnell and Terra Nivea ice caps on far southern Baffin Island since the 1950s (Papasodoro et al. 2015). On Terra Nivea the ice cap-wide mass balance increased from  $-0.30 \pm 0.19$  m w.e. a<sup>-1</sup> over the period 1958/59–2007 to  $-1.77 \pm 0.36$  m w.e. a<sup>-1</sup> over the period 2007–2014. Similarly, Way (2015) found that Terra Nivea Ice Cap has lost 22% of its area since the late 1950s, while Grinnell Ice Cap has lost a total of 18%. These rapid reductions in area have been linked to increasing summer air temperatures, and suggest that these ice masses are far out of equilibrium with current climate.

### 3.2.3 Glacier dynamics

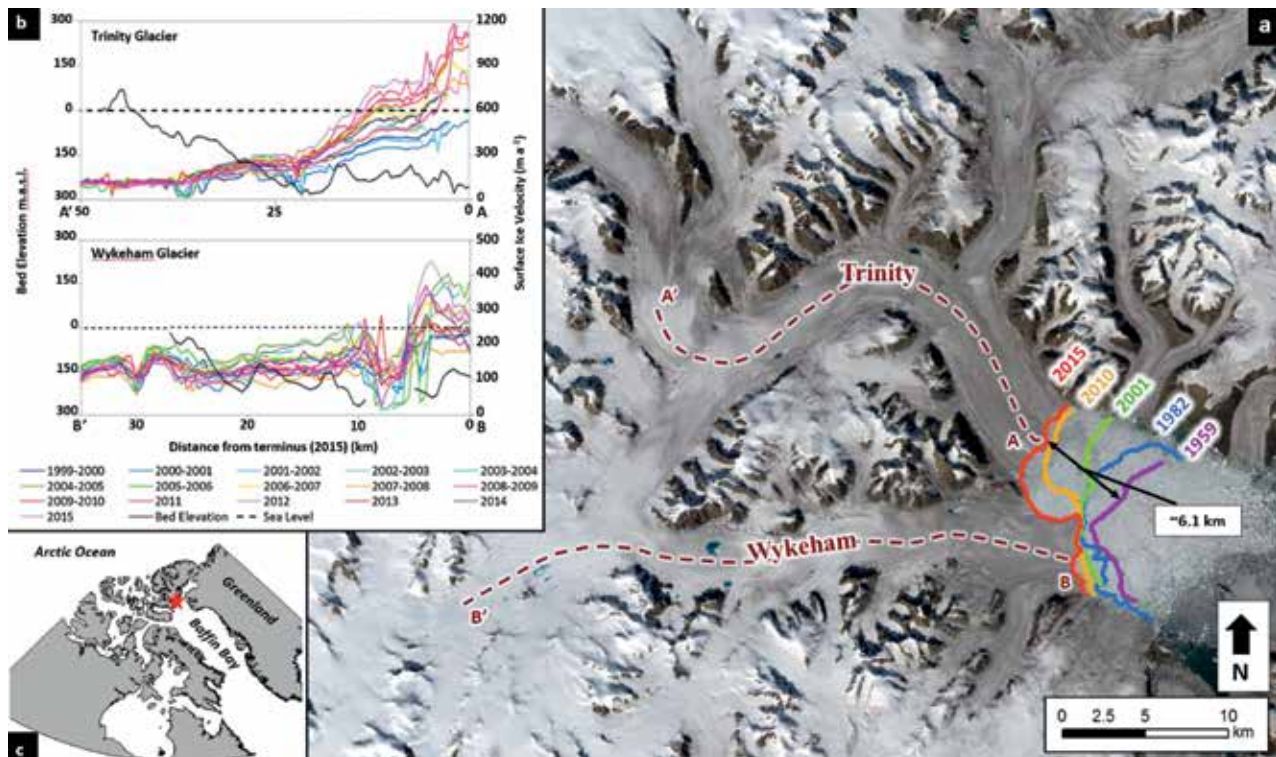
An important question, given the widespread and increasing recent mass losses from glaciers and ice caps across the CAA, is whether glacier motion is changing in response. To address this, Van Wychen et al. (2016) matched pairs of Radarsat and Landsat satellite imagery to determine the velocity structure of the ice masses of Axel Heiberg and Ellesmere Islands for the period 1999–2015. Out of 117 sampled glaciers the vast majority did not show significant velocity changes, but six had fluctuating velocities (both speed-up and slow-down), eight slowed progressively and two accelerated throughout the observation period. To investigate the causes of these velocity variations, Van Wychen et al. (2016) combined the observed glacier dynamics with a record of terminus positions and an inventory of optical satellite imagery used to identify glacier surface features that would indicate surge activity (e.g., looped surface moraines, extensive fresh crevassing). Glacier surging refers to a period of rapid glacier advance (over months to years) that can occur after a long period of near-stagnation (of decades to centuries), typically driven by the movement of excess ice mass from the top to bottom of a glacier as part of a glacier's internal dynamics (Meier and Post 1969). Pulse glaciers refer to ice masses that share many of the



characteristics of surging, such as significant velocity variability, but where the variations do not appear to encompass the entire glacier length (Van Wychen et al. 2016). Using this classification scheme, all of the glaciers that displayed velocity variability or a progressive slow-down could be explained by surge or pulse mechanisms. For these glaciers it appears that internal mechanisms, rather than external forcing (e.g., climatic or ocean warming), are responsible for the majority of observed dynamic changes.

The major exceptions to this behaviour are the two glaciers that underwent consistent acceleration, the Trinity and Wykeham glaciers of southeastern Prince of Wales Icefield, whose changes appear to be driven by external forcing (Figure 4). Over the period 1999–2015 the surface velocity of Wykeham Glacier nearly doubled (increasing from

$\sim 200 \text{ m a}^{-1}$  to  $\sim 400 \text{ m a}^{-1}$  in the lowermost terminus region), while the surface velocity of Trinity Glacier nearly tripled (increasing from  $\sim 400 \text{ m a}^{-1}$  to  $\sim 1200 \text{ m a}^{-1}$  in the lowermost terminus region). From 1959 to 2015, the terminus of Trinity Glacier retreated by  $\sim 6 \text{ km}$  while Wykeham Glacier retreated by  $\sim 2 \text{ km}$ . However, roughly half of the observed retreat for both glaciers has occurred since  $\sim 2001$ , coincident with the onset of faster motion (Van Wychen et al., 2016) and increases in summer air temperatures (Sharp et al. 2011). This pattern of terminus retreat coincident with acceleration is at odds with the expected behaviour of surge-type and pulse-type glaciers and suggests that a unique climatic and/or oceanic mechanism is driving the recent speed-up of these two glaciers. A comparison of glacier surface elevations in 2008 and 2014 indicates  $\sim 2\text{--}5 \text{ m a}^{-1}$  of surface lowering in the lowermost 10 km of both



**FIGURE 4.** (a) Frontal retreat (1959–2015) of Trinity and Wykeham Glaciers (background: Landsat 8, July 29, 2015); (b) 1999–2015 centreline surface velocities of Trinity and Wykeham glaciers, extracted along the red dashed lines shown in (a); (c) location of Trinity and Wykeham Glaciers (marked by red star; also see Figure 1).



glaciers, approximately double that expected from surface melting (Mair et al. 2009, Marshall et al. 2007). This implies that at least half of the observed thinning is driven by glacier acceleration. This pattern of unstable terminus acceleration, thinning and prolonged retreat has previously been reported in Greenland (e.g., Helheim Glacier; Howat et al., 2005), but not previously identified in the Canadian Arctic. Trinity and Wykeham glaciers accounted for ~22% of total CAA iceberg discharge to the ocean in 2000, but ~62% in 2015 (Van Wychen et al. 2016), indicating that changes in the dynamics of just a few glaciers can change iceberg production patterns for the entire Canadian Arctic. Given that these glaciers are grounded below sea level for ~30-45 km upglacier from their current calving fronts, they are particularly prone to further retreat and break-up before re-stabilization can occur.

### 3.3 Ice shelves

#### 3.3.1 Ice shelf area changes

There are currently three remaining major ice shelves on the northern coast of Ellesmere Island: the Petersen, Milne and Ward Hunt (Figure 5). These ice shelves are remnants of the Ellesmere Ice Shelf (unofficial name) that once fringed the entire northern coast of Ellesmere Island, and was first described by Lt. Aldrich of the British Arctic Expedition of 1875-76 (Nares 1878). Later observations by Peary (1907) of a 'broad glacial fringe' along the north-western coast of Ellesmere Island were used by Vincent et al. (2001) to estimate an ice shelf extent of 8900 km<sup>2</sup> and length of 500 km in 1906. In the studies of England et al. (2008) and Antoniadou et al. (2011), radiocarbon dating of driftwood collected in fiords behind ice shelves was used to infer when the north coast of Ellesmere Island was last ice-free. Based on samples collected inland of Ward Hunt Ice Shelf and five adjacent fiords, it is likely that the Ellesmere Ice Shelf began to form ~4000-5500 years ago.

Over the period 1906-1982, ~90% of the Ellesmere Ice Shelf was lost in a series of large calving events, particularly prior to the 1950s (Koenig et al. 1952). These events released many large ice islands into the Arctic Ocean and

left behind six remnant ice shelves located in fiords and bays along the coastline. From the 1950s to the end of the 20th century there were several calving and break-up (fracture development) events that further fragmented these ice shelves (Hattersley-Smith 1963, Jeffries and Serson 1983, Jeffries 1986a). During the 1960s the Ward Hunt Ice Shelf lost 50% of its area (596 km<sup>2</sup> from 1961-1962) (Hattersley-Smith 1963), and the Ayles Ice Shelf calved 15 km<sup>2</sup> (from 1962-1966; Jeffries 1986a). Between 1980 and 1983, the Ward Hunt Ice Shelf lost ~80 km<sup>2</sup> in two calving events (Jeffries and Serson 1983). From then until 2005, the ice shelves remained generally stable.

Starting in 2005, a renewed period of ice shelf losses has occurred. From 2005 to 2012, >50% of their total 2005 area of ~1043 km<sup>2</sup> was lost, with many ice islands drifting in the Beaufort Gyre as a consequence (Mueller et al. 2013). This period of ice shelf break-up began with the complete loss of the Ayles Ice Shelf (87 km<sup>2</sup>) and an ~8 km<sup>2</sup> loss from the Petersen Ice Shelf in summer 2005 (Copland et al. 2007, White et al. 2015a). In 2008, 42 km<sup>2</sup> of the Ward Hunt Ice Shelf calved away, along with 60% (122 km<sup>2</sup>) of the Serson Ice Shelf, a ~9 km<sup>2</sup> loss from the Petersen Ice Shelf, and the complete loss of the Markham Ice Shelf (50 km<sup>2</sup>; Mueller et al. 2008, White et al. 2015a). The year 2011 was also marked by substantial calving events, including a 39 km<sup>2</sup> loss from the Ward Hunt Ice Shelf as it broke into two, 5.5 km<sup>2</sup> loss from the Petersen Ice Shelf and a 32 km<sup>2</sup> loss

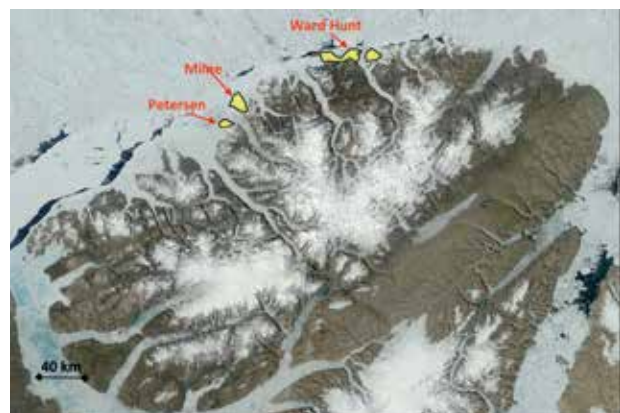


FIGURE 5. Summer 2015 location of ice shelves on northern Ellesmere Island. Base image: MODIS, July 12, 2015.

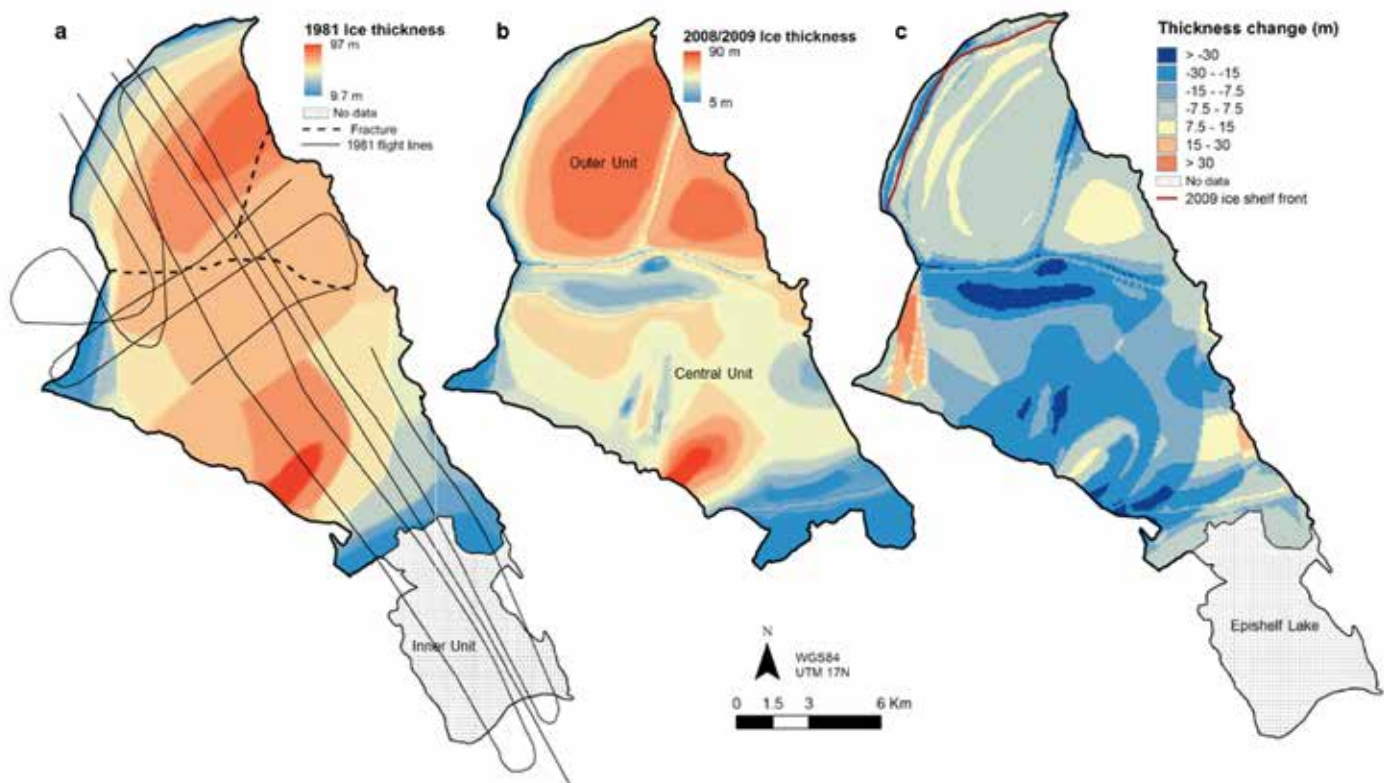
from the Serson Ice Shelf (White et al. 2015a). An additional 5.5 km<sup>2</sup> loss from the Petersen Ice Shelf occurred in summer 2012. Many of these ice shelf losses have occurred in conjunction with the loss of adjacent multi-year landfast sea ice, such as 690 km<sup>2</sup> of ~70 year old sea ice which broke out from the head of Yelverton Bay in August 2005 shortly before losses of parts of the Petersen Ice Shelf (Pope et al. 2012). Many ice shelf losses have also resulted in the loss of adjacent freshwater epishelf lakes that were once dammed behind them (Box B; White et al. 2015b).

### 3.3.2 Recent changes in ice shelf thickness

Recent ice shelf thickness measurements are limited to the Petersen and Milne ice shelves. In May 2011 measurements conducted with a 250 MHz ground penetrating radar (GPR) system on the Petersen Ice Shelf revealed a mean thickness

of 29 m, with a standard deviation of 24 m and maximum thickness of ~106 m for a tributary glacier (White et al. 2015a). GPR measurements on the Milne Ice Shelf in April 2008 and May 2009 revealed a mean thickness of 55 m, with a standard deviation of 22 m and a maximum of 94 m (Mortimer et al. 2012). Both of these ice shelves contain thicker ice where tributary glaciers provide input, and are thinner near their inland margins.

A comparison of the 2008/2009 ice thickness measurements on the Milne Ice Shelf with airborne measurements collected in 1981 (Prager 1983; Narod et al. 1988) provides the only direct quantification of ice shelf thickness change for the Canadian Arctic (Mortimer et al. 2012). These measurements indicate that the ice shelf reduced in volume by 13% over this period, with an average thinning of  $8.1 \pm 2.8$  m, equivalent to  $-0.26 \pm 0.09$  m w.e. a<sup>-1</sup> (Figure 6).



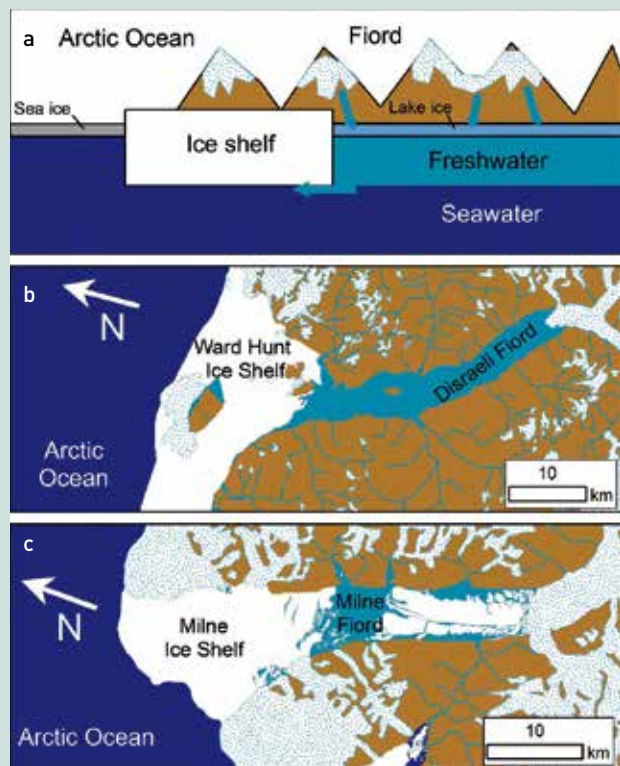
**FIGURE 6.** Thickness maps of the Milne Ice Shelf showing: (a) 1981 thicknesses interpolated from a contour map produced by Prager (1983), with superimposed flight lines; (b) interpolated thicknesses derived from 2008/2009 ground penetrating radar measurements; (c) changes in the thickness of the Milne Ice Shelf between 1981 and 2009. Source: Mortimer et al. (2012).

**BOX B.** Epishelf lakes

Ice shelves at the northern coast of Ellesmere Island are attached to the coast along the sides of fiords and embayments. However, there can also be areas at the rear of ice shelves where they are not attached to the land, which allows for the formation of a freshwater layer in the fiord that floats on the relatively high density sea water below. These are referred to as epishelf lakes and they can exert considerable influence on the ice shelves that dam them (Figure B1).

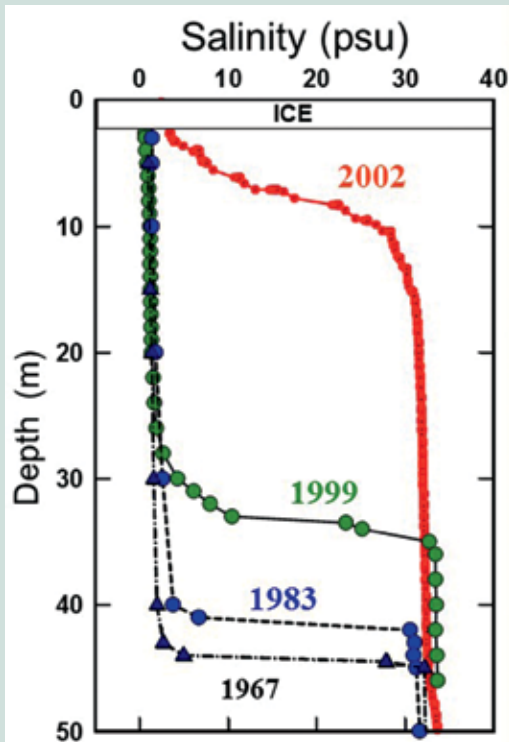
Epishelf lakes are supplied with meltwater that flows from the surrounding land in the brief summer period. This freshwater collects in the upper water column with minimal mixing with the ocean below, creating a sharp step-like change in salinity (Figure B2) at a depth that is equal to the minimum draft of the ice shelf. If more meltwater flows into the fiord, it is forced out under the ice shelf. If this water is cold enough, it can then freeze on the underside of the ice shelf (Jeffries et al. 1988). In the past, this process has increased the thickness of the ice shelf, which in turn can deepen the epishelf lake (Jeffries 2002).

In the current climate regime, with ice shelves thinning and with warmer meltwater inputs, epishelf lakes are reducing in size. For example, between 1999 and 2002 an epishelf lake in Disraeli Fiord drained through a fissure in the Ward Hunt Ice Shelf (Mueller et al. 2003). This was the largest epishelf lake (6.1 km<sup>3</sup> in 1967) in the Northern Hemisphere and the first time that the drainage of an epishelf lake had been documented (Figure B2). Since that time, the reduction in depth of Ellesmere Island's epishelf lakes has been examined by water column profiling and the reduction in area of epishelf lakes has been inferred using RADARSAT satellite imagery (Veillette et al. 2008, White et al. 2015b). In 2011, ArcticNet researchers installed a mooring in the last large epishelf lake found in Milne Fiord to monitor the changes in this unique environment.



**FIGURE B1.** (a) Cross sectional diagram of an epishelf lake showing the freshwater layer (light blue) that is impounded behind the ice shelf. Since the freshwater is less dense than the seawater, it rests above the ocean (dark blue), persisting in a layer between the coast and the ice shelf. The lake is covered by perennial freshwater ice that prevents mixing by the wind. More meltwater (light blue arrows) from snow and glaciers (stipple) is added to the lake in the summer, which pushes freshwater out under the ice shelf. (b) Map of the epishelf lake that persisted in Disraeli Fiord behind the Ward Hunt Ice Shelf until 2002. (c) Map of the epishelf lake in Milne Fiord as of spring 2015.





Epishelf lakes are also fascinating ecosystems that are highly structured. It is possible for organisms that live in fresh or brackish water to co-exist just above marine organisms in the same water column. Green algae dominate in the fresh upper water column, whereas marine diatoms are more commonly found in deeper waters (Veillette et al. 2011). Similar distributions can be found with copepods (tiny crustaceans), bacteria and viruses where certain types of these organisms are found either in the fresh or marine environments (Van Hove et al. 2001, Veillette et al. 2011). Since these stratified ecosystems are structured by the ice shelves that dam them, they are vulnerable to ice shelf break-up, and are likely to disappear completely from Nunavut over the next few decades.

FIGURE B2. Salinity profiles showing the loss of the freshwater epishelf lake in Disraeli Fiord. Over the years, researchers lowered instruments into the water column to measure salinity. In 1999 and before, the salinity was near zero in the upper water column, indicating freshwater, before transitioning rapidly to sea water below. In 2002, the profile shows a dramatic loss of most of the freshwater epishelf lake (due to flow through a crack in the Ward Hunt Ice Shelf) and replacement by high salinity sea water from below.

### 3.3.3 Recent changes in ice shelf inputs and mass balance

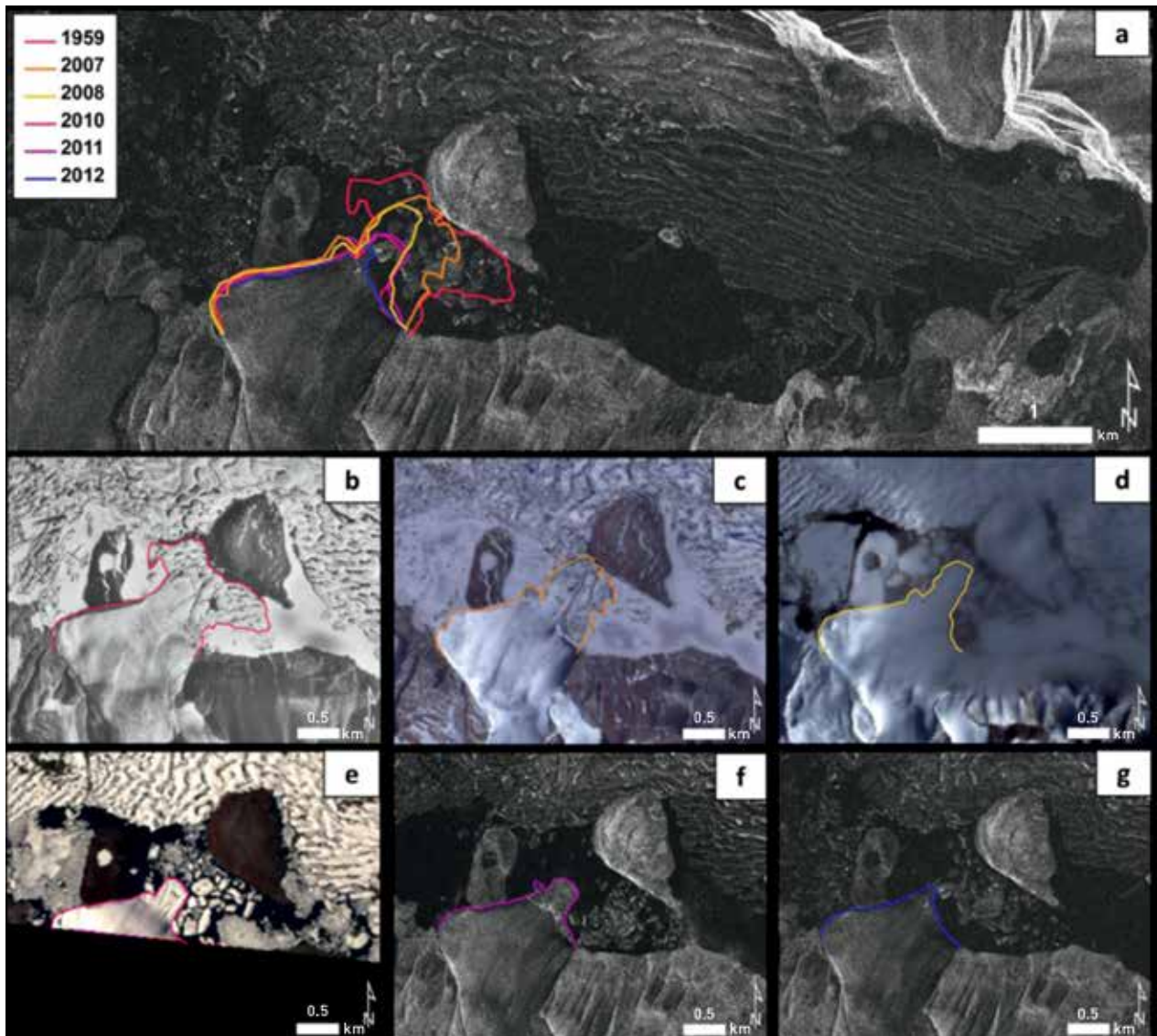
For some of the ice shelves along northern Ellesmere Island, glaciers have provided an important source of mass input (Jeffries 1986b). Of the three remaining ice shelves today, only the Milne and Petersen show evidence of glacier input, with the Ward Hunt Ice Shelf primarily formed from the *in situ* accumulation of sea ice and snow. For the Milne Ice Shelf, it is clear that there has been a marked reduction in the importance of glacier input over the past ~50 years, with aerial photographs indicating that five tributary glaciers terminated on the ice shelf in 1959 (Mortimer et al. 2012). Three of these glaciers extended between 1.5 and 4.5 km into the ice shelf interior, but by 2011 the glacier input was limited to a single glacier contributing an estimated 0.048 m w.e.  $a^{-1}$  to the ice shelf. This accounts for

<20% of the average thinning of 0.26 m w.e.  $a^{-1}$  measured between 1981 and 2008/9, indicating that the ice shelf is far out of balance with current climate conditions (Mortimer et al. 2012).

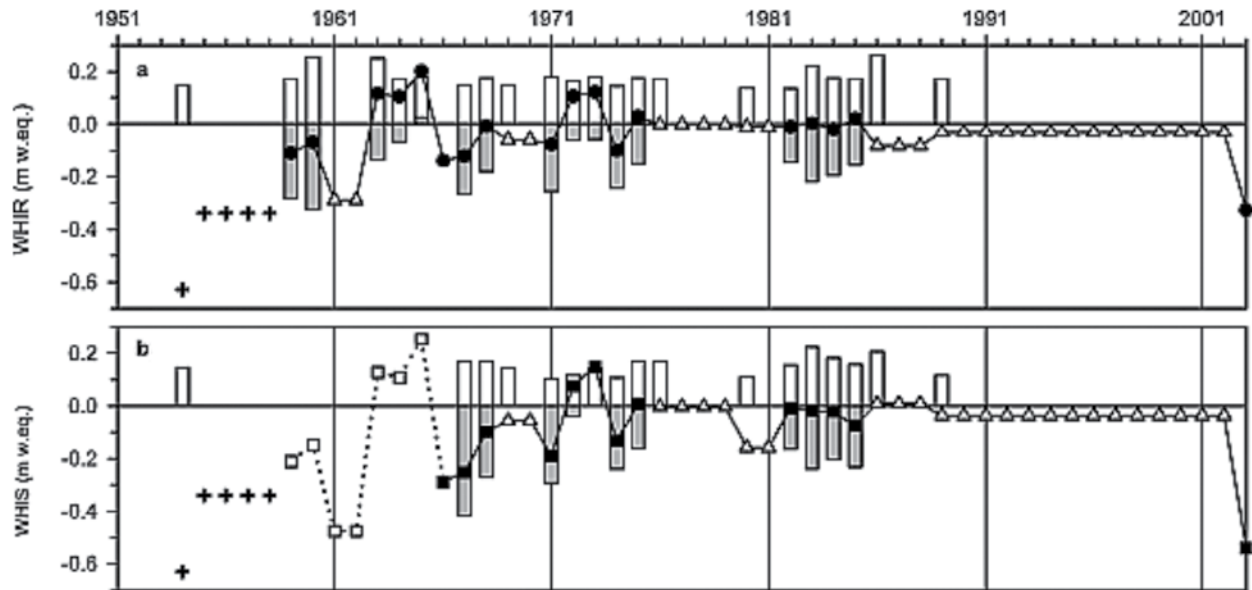
The Petersen Ice Shelf currently receives mass from two tributary glaciers along its remaining landfast margin. An additional tributary flowed into the southern margin of the ice shelf in 1959, but this became detached by 2007, therefore eliminating it as a source of mass to the ice shelf (Figure 7; White 2012). Based on velocity measurements made between 2011 and 2012, the remaining two glaciers contribute 1.19-5.65 Mt  $a^{-1}$  to the ice shelf (White et al. 2015a). This mass input is far exceeded by the current surface ablation on the Petersen Ice Shelf of 28.45 Mt  $a^{-1}$ , helping to explain why the ice shelf has weakened to the point of disintegration.

The longest mass balance record available for a Canadian ice shelf is for the Ward Hunt Ice Shelf and adjacent Ward Hunt Ice Rise (region of the ice shelf resting on the sea bed). Braun et al. (2004) compiled a continuous record of annual surface mass balance for these locations from 1954 to 2003 (Figure 8), which show that winter snow

accumulation remained relatively constant over time, but that summer ablation was much more variable. Positive balance years have been infrequent (only 1963–1965 and 1972–1973), and negative balance years dominate the record with total mass losses of 1.68 m w.e. ( $0.04 \text{ m w.e. a}^{-1}$ ) for Ward Hunt Ice Rise and 3.1 m w.e. ( $0.07 \text{ m w.e. a}^{-1}$ ) for Ward Hunt



**FIGURE 7.** Aerial photography and satellite imagery highlighting changes in the tributary glacier flowing from the southern coast into Petersen Bay and the dispersement of icebergs. (a) Radarsat-2, February 3, 2012; (b) Aerial photography, August 13, 1959; (c) ASTER, July 7, 2007; (d) ASTER, August 22, 2008; (e) ASTER, July 19, 2010; (f) Radarsat-2, July 19, 2011; (g) Radarsat-2, February 3, 2012. Source: White (2012).



**FIGURE 8.** Surface mass balance for: (a) Ward Hunt Ice Rise and (b) Ward Hunt Ice Shelf from 1954 to 2003 for: winter (open bar), summer (shaded bar), and annual balances (solid square). Plus symbols represent multiyear balance estimates based on limited measurements, open triangles represent annual values calculated as averages of multiyear balances, and open square/dots are annual balances calculated using a linear regression. Modified from Braun et al. (2004).

Ice Shelf. This contrast in surface mass balance has been associated with the ridge/trough topography of the ice shelf, where meltwater collected in troughs enhances ablation (Braun et al. 2004). According to ablation measurements, 2003 was the most negative year, when 50% of the mass loss from 1989–2002 occurred (0.54 m w.e. for Ward Hunt Ice Shelf, and 0.33 m w.e. for Ward Hunt Ice Rise).

The mass balance measurements presented by Braun et al. (2004) are limited to the surface and therefore do not account for mass fluctuations at the base of the ice shelf. However, the thickness change measurements presented by Mortimer et al. (2012) provide a means to determine the state of the basal ice. Mortimer et al. (2012) applied the rate of surface mass loss for the Ward Hunt Ice Shelf to the Milne Ice Shelf and combined this rate with the total thinning determined for the 1989–2003 period. Based on this calculation, thinning due to basal melt would account for ~73% (0.19 m w.e. a<sup>-1</sup>) of mass loss, suggesting that it can be a dominant loss mechanism.

### 3.4 Ice islands

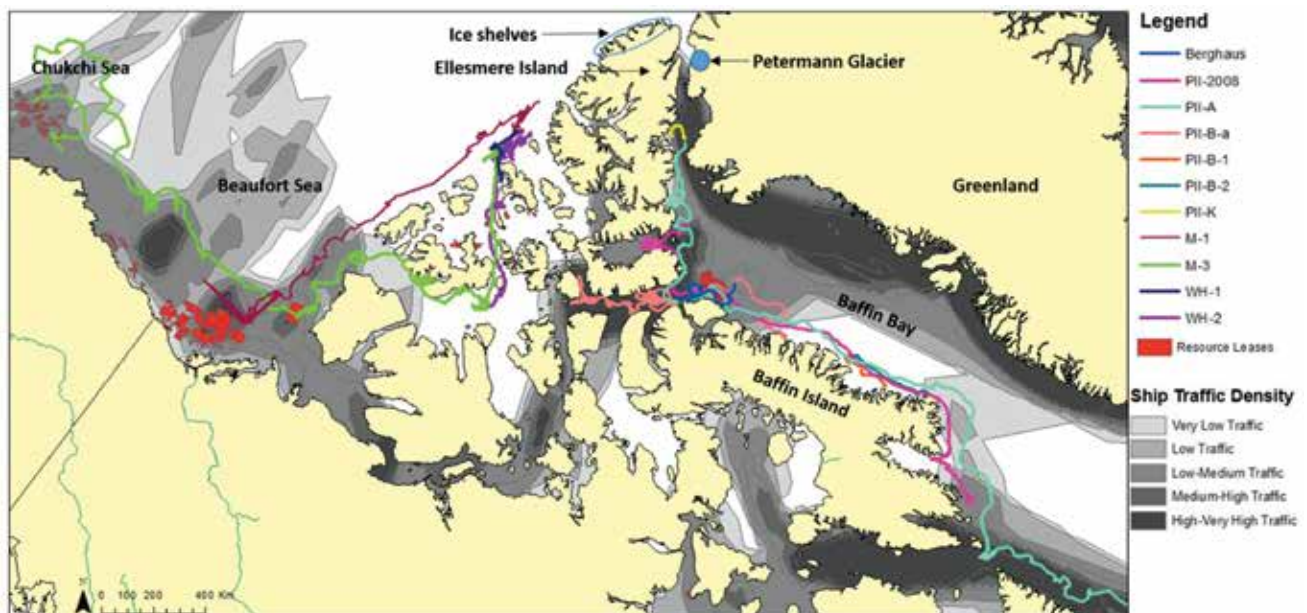
Ice islands have been described as the “most massive ice features known in the Arctic Ocean” (Jeffries et al. 1987), with the largest observed in the CAA covering an area of ~1,000 km<sup>2</sup> (Jeffries 1992, Koenig et al. 1952). These large, tabular icebergs were first observed in the late 1940s (Fletcher 1950) and were originally used for military and scientific purposes as aircraft landing and research platforms (Althoff 2007, Jeffries 1992, Mueller et al. 2013). These ice features have become subjects of interest recently due to the link between climate change and calving events of ice shelves and floating glacial tongues (Copland et al. 2007, Derksen et al. 2012). Shipping and offshore exploration companies are also concerned due to the risk of a collision between an ice island and a vessel or infrastructure (McGonigal et al. 2011, Peterson 2011). This risk is illustrated in Figure 9, which shows ice island drift trajectories overlapping regions of shipping and natural resource extraction activity.



Ice islands drifting in the CAA typically originate from one of two locations: northwest Greenland's floating glacial tongues or the ice shelves of northern Ellesmere Island (Higgins 1989, Jeffries 2002). Ice islands from Greenland generally drift south through Eastern Canadian Arctic waters, at times reaching the Labrador Sea or the Grand Banks of Newfoundland (Newell 1993). Ice islands of Ellesmere Island origin commonly drift clockwise in the Beaufort Gyre toward the Beaufort and Chukchi seas, though some have been observed within the channels of the CAA (Jeffries et al. 1987, Van Wychen and Copland 2017), or drifting east into Nares Strait (Nutt 1966) (Figure 9).

Both of these source regions have experienced multiple, substantial calving events since the start of the 21st century (Mueller et al. 2013, Peterson 2011). The Petermann Glacier of northwest Greenland calved large ice islands in 2001 (71 km<sup>2</sup>; Johannessen et al. 2011), 2008 (31 km<sup>2</sup>; Johannessen et al. 2011), 2010 (253 km<sup>2</sup>; Falkner et al. 2011) and 2012 (32 km<sup>2</sup>, NASA 2012). Johannessen et al.

(2011) note the absence of sea ice in the fiord prior to the 2010 calving, a condition also observed in MODIS satellite imagery which captured the 2012 event (NASA 2012). This situation likely promotes calving by removing the floating glacial tongue's protection from ocean and wind-induced waves (Falkner et al. 2011). Rignot and Steffen (2008) suggest that enhanced channel erosion on the subsurface of the Petermann Glacier's floating tongue caused by warming ocean waters is another probable contributor. It cannot be conclusively stated that these events are linked to climate warming, although Falkner et al. (2011) propose that global atmospheric pressure and temperature changes have an impact on both of the aforementioned calving precursors. Copland et al. (2007) also argue that these conditions, namely increasing atmospheric temperatures and the absence of protective sea ice, led to the weakening and eventual calving of the Ayles Ice Shelf in August 2005. This event created a 66 km<sup>2</sup> ice island and 21 km<sup>2</sup> of smaller fragments.



**FIGURE 9.** Overlap between ice island drift trajectories, historic U.S. and Canadian oil and gas exploration lease blocks and ship traffic density. Legend abbreviations are: PIL = Petermann Ice Island, M = Markham, WH = Ward Hunt. Shapefiles of oil and gas leases are courtesy of the U.S. Bureau of Ocean Energy Management (BOEM 2013) and Aboriginal Affairs and Northern Development Canada (AANDC 2013). Shipping density contours represent a snapshot of marine traffic for August 2013 (exactEarth 2013). Major sources of ice hazards annotated.

The drift and deterioration of five ice islands from the Petermann Glacier calvings were monitored between 2010 and 2015 by GPS tracking beacons and RADARSAT-1 and 2 satellite images (Crawford 2013, Crawford et al. 2015, Halliday et al. 2012, Hamilton et al. 2013, Wagner et al. 2014). These ice islands followed the common drift pattern through Nares Strait and into the cyclonic current of Baffin Bay (Newell 1993, Tang et al. 2004). Some fragments, such as Petermann Ice Island (PII)-B-a of the 2010 Petermann Glacier calving event, diverted into Lancaster Sound due to the intrusion of strong ocean currents at the north entrance of the sound (Hamilton et al. 2013, Peterson et al. 2009, Tang et al. 2004) (Figure 9).

The time it takes for an ice island to drift through the Eastern Canadian Arctic can vary considerably due to these diversions and/or grounding events. This is illustrated by comparing the 8 month and 3 year transit times of PII-A and PII-B-a, respectively, from the Petermann Glacier to the Labrador Sea (Crocker et al. 2013). The latter ice island's transit was delayed due to the aforementioned diversion and becoming grounded 130 km northeast of Clyde River, Nunavut, for a 12 month period (Crawford 2013, L. Desjardins pers. comm.). Offshore oil and gas operators will need to consider that ice islands have frequently grounded on Baffin Island's continental shelf (Crawford 2013, L. Desjardins and S. Tremblay-Therrien, pers. comm.) if natural resource extraction is to be conducted in the Eastern Canadian Arctic since sub-seafloor infrastructure will be at risk from ice island scouring (King et al. 2003).

Ice islands in the Eastern Canadian Arctic have drafts (distance between the waterline and ice bottom) ranging from approximately 60 to 140 m, as determined from GPR, multibeam sonar and autonomous underwater vehicle surveys (Crawford 2013, Forrest et al. 2012, Hamilton et al. 2013, Wendleder et al. 2015). An ice island's draft determines its possible grounding locations and, in combination with its sail, provides a dominant control on the ice feature's drift. The deterioration of this vertical dimension is relatively more important for ice islands than icebergs due to the former's extensive horizontal surfaces (Ballicater 2012).

Surface melt and micro-meteorological data indicate that the ice surface can melt at an average rate of 3.3 cm d<sup>-1</sup> in August in Lancaster Sound (Crawford et al. 2015), while average basal and surface deterioration rates of 5.0 cm d<sup>-1</sup> and 2.4 cm d<sup>-1</sup> were observed further south in the Labrador Sea (Halliday et al. 2012).

Ice islands also deteriorate through calving processes such as the 'footloose mechanism', which is induced by buoyancy stresses generated by an ice ram (an underwater protrusion at the ice island's edge) (Wagner et al. 2014). These deterioration processes result in ice island fragments, as well as 'bergy-bits' and 'growlers', which are 5-15 m and <5 m in length, respectively. These are relatively small ice features yet they are still capable of damaging vessels or infrastructure and are particularly dangerous due to the difficulty of detecting them remotely (Crocker et al. 2004, Saper 2011). Interestingly, Stern et al. (2015) observed that the temperature and salinity stratification of the water column adjacent to a large (~41 km<sup>2</sup>) grounded ice island in Baffin Bay was altered by wind-induced up- and down-welling, or Ekman transport. This process may lead to enhanced deterioration of ice island sidewalls (when grounded) to the left of the predominant wind direction, and possibly even contribute to fracture mechanisms such as the footloose mechanism.

Ice islands which have originated from the Ellesmere Island ice shelves over the past decade drifted south along the western coast of the CAA, with 'Markham-3 (M-3)' and 'Ward Hunt-2 (WH-2)' diverting into the Sverdrup Basin (Crawford 2013, McGonigal et al. 2011). M-3 drifted approximately 2200 km and was tracked as far as 161°W in the southern Beaufort Sea (Crawford 2013, Hochheim et al. 2012), reinforcing the concern of offshore exploration companies regarding these ice hazards (Mueller et al. 2013, Sackinger et al. 1991).

### 3.5 Summary and outlook

In summary it is clear that many parts of the cryosphere are undergoing rapid, and increasing, changes in the Eastern Canadian Arctic. All available evidence, whether from *in situ* measurements, satellite remote sensing, or modelling, indicates that glaciers and ice caps have been in a strongly negative mass balance condition since at least the start of the 21st century. Losses have accelerated over the past decade, with glacier and ice cap thinning particularly prominent at low elevations and in more southerly regions. Smaller ice masses are also experiencing high rates of mass loss, with the complete loss of some small ice bodies already recorded on Axel Heiberg Island (Thomson et al. 2011). Way (2015) predicts that if the current ice declines on Grinnell and Terra Nivea ice caps, Baffin Island, continue to AD 2100, the total ice-covered area will be reduced by >57% compared to present.

The study of Lenaerts et al. (2013) provides one of the few detailed assessments of how glaciers and ice caps across the CAA will respond to future climate warming. Using the moderate AR5 RCP4.5 warming scenario and predictions to the year 2100, Lenaerts et al. (2013) demonstrate that enhanced meltwater runoff from CAA glaciers will not be sufficiently compensated by increased snowfall. This results in sustained 21st century mass losses for glaciers across the CAA in >99% of model runs. This makes it highly likely that the recently observed mass losses of these glaciers will continue, resulting in the complete disappearance of many small glaciers and ice caps over the next century.

Both early and recent studies on the ice shelves of northern Ellesmere Island suggest that these features are unlikely to persist long into the future if current climate conditions persist (Hattersley-Smith 1955, Braun et al. 2004, Copland et al. 2007, Mortimer et al. 2012, White et al. 2015a, 2015b). This prediction is based on the inability of past ice shelves to regenerate, and gradual mass losses over the last ~100 years which have weakened ice shelves by causing thinning and fracture development. It has also been observed that increasing air temperatures and record

warm summers combined with the loss of multi-year land-fast sea ice, open water conditions along the northern coast of Ellesmere Island and offshore/along-shore winds are providing an environment conducive to ice shelf break-up that will eventually lead to the demise of the remaining ice shelves (Copland et al. 2007, Pope et al. 2012, White et al. 2015a). Hattersley-Smith (1955) predicted that the Ward Hunt Ice Shelf would disintegrate by 2035 if the summer conditions of 1954 continued, while White et al. (2015a) estimated that, based on the rate of surface melt and lack of glacial input, the Petersen Ice Shelf would disappear by 2041-2044. However, recent calving and open water events suggest that the complete loss of the Petersen Ice Shelf will likely occur well before this. For the Milne Ice Shelf, Mortimer et al. (2012) concluded that it is not in equilibrium with current climate as it formed under past colder conditions. Given the recent thinning of this ice mass and reduction in glacier inputs, it is likely that over the next few decades the Milne Ice Shelf will follow the pattern of breakups recently observed for all other northern Ellesmere Island ice shelves. The loss of ice shelves is also likely to lead to the loss of a unique assemblage of microbial life (Box C).

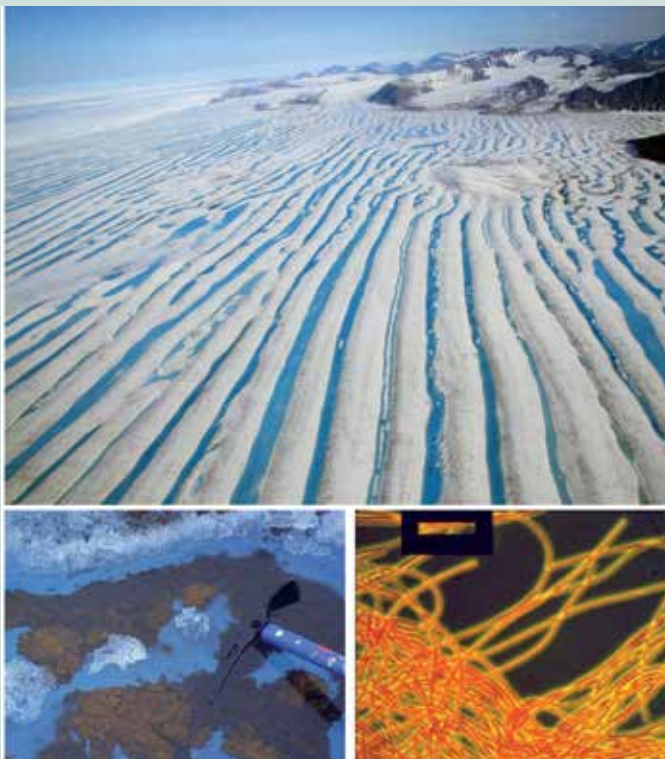
Since the Ellesmere Island ice shelves show no sign of regeneration they will eventually cease to become a source of ice islands. However, calving from floating glacier tongues on Ellesmere Island and northwest Greenland will persist for the foreseeable future, leading to continuing risks for marine shipping and offshore exploration in the Eastern Canadian Arctic.



**BOX C.** Life on ice

Snow fields, glaciers, ice shelves and ice islands are changing rapidly in Canada's north. However, the consequences of this are not restricted to the physical environment. These ice masses are also habitats for diverse organisms that survive, and even thrive, in the "cold biosphere", the ensemble of habitats over planet Earth that experience prolonged cold and freezing. These life forms are mostly microscopic, and they have captured the interest of microbiologists in their research on how cells cope with extreme environments and how life may have persisted and evolved during periods of glaciation on Earth (Vincent and Howard-Williams 2000). These observations have also led scientists to speculate about the potential for microbial life in icy environments elsewhere in the solar system, for example on planet Mars and certain moons of Jupiter. Biotechnologists have developed a keen interest in these microscopic 'extremophiles' because they are the potential source of unusual enzymes, antifreeze compounds and other biomolecules that could have applications in medicine, food production, bioremediation and industrial processes (e.g., Christner, 2010).

Snow was once thought of as a sterile substance, but is now known to contain many types of bacteria and a variety of more advanced, nucleus-containing cells ('microbial eukaryotes'). Studies along the northern coast of Ellesmere Island have shown that the snow contains diverse microbiota, including species likely transported in the atmosphere from the Arctic Ocean and from lakes and streams. Some of the other species were similar to those found outside Canada but from alpine regions, Antarctica, and other parts of the Arctic, indicating the global distribution of these genetically distinct cells throughout the cold biosphere (Harding et al. 2011).



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Cryoconite holes were first discovered on the Greenland Ice Sheet, and are now the focus of international attention as 'extreme ecosystems' with diverse microscopic life. These appear as small cylindrical melt holes at lower elevations on the surface of glaciers, and they are common features of the glacial environments of Nunavut. The holes form because small

**FIGURE C1.** Elongate melt pools on the surface of the Ward Hunt Shelf, northernmost Ellesmere Island, Nunavut, 23 July 2007 (see Figures 1 and 5). Most of this ancient ice shelf melted, fractured and disappeared over the last decade due to rapid climate warming. At some locations, brightly coloured microbial mats coat the bottom sediments of the pools (bottom left). These mats are composed of filamentous cyanobacteria (fluorescence photomicrograph, bottom right) and many associated microorganisms.

amounts of sediment (called cryoconite) heat up in the sun and melt down into the ice. Many small cryoconite holes can coalesce into larger ones, and studies on the White Glacier, Axel Heiberg Island, and on the Canada Glacier, Antarctica, have shown that they harbour microbial consortia composed of green algae, cyanobacteria, and diatoms as well as other microorganisms (Mueller et al. 2001). These cryoconite features can modify the reflective properties of ice, and thereby hasten the local melting of glacier surfaces.

The richest communities of microscopic life on ice in Nunavut have been discovered in meltwater lakes on the ice shelves along the northern coast of Ellesmere Island (Jungblut et al. 2017; Figure C1). At some locations on these ice shelves, the base of the lakes is coated with a bright orange ‘microbial mat’ composed of cyanobacterial filaments that bind together the sediments. The coloration is the result of red-orange carotenoids such as canthaxanthin that protect the cells from harsh UV radiation that penetrates to bottom of these clear waters. Genetic comparisons of these Nunavut mat communities with similar communities in Antarctica have shown that they contain hundreds of species, with an impressive genetic capacity to cope with the extreme stresses imposed in these environments (Varin et al. 2012). Climate warming in Nunavut is leading to the rapid loss of these ice-dependent habitats and the imminent extinction of these unique biological communities (Vincent et al. 2011).

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