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Forty-seven Years of Research on the Devon Island Ice Cap, Arctic Canada

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ABSTRACT. The Devon Island ice cap has been the subject of scientific study for almost half a century, beginning with the first mass balance measurements in 1961. Research on the ice cap was the first to investigate (1) the role of meltwater in seasonal ice-velocity variations on a polythermal Arctic ice cap, (2) the use of air temperature rather than net radiation as a proxy for the energy driving surface melt, and (3) the influence of the changing frequency of specific synoptic weather configurations on glacier melt and mass balance. Other research has included investigations of ice cap geometry, flow dynamics, and mass balance; ice core analyses for records of past climate and contaminant deposition; and studies of changes in ice cap area and volume and their relationship to surface mass balance and ice dynamics. Current research includes ground studies connected to efforts to calibrate and validate the radar altimeter that will be carried by the European Space Agency's (ESA) CryoSat2 satellite, and a major collaborative Canadian International Polar Year (IPY) project focused on the Belcher Glacier, on the northeast side of the ice cap, that examines hydrodynamics of large tidewater glaciers. This paper summarizes our current knowledge of the Devon Island ice cap and identifies some of the outstanding questions that continue to limit our understanding of climate-ice cap interactions in Arctic regions.

Key words: Devon Island ice cap, Canadian Arctic, glaciology, ice dynamics, mass balance, climate change, tidewater, ice cores

RESUMÉ. La calotte glaciaire de l'île Devon fait l'objet d'une étude scientifique depuis près d'un demi-siècle, les premières mesures du bilan massique remontant à 1961. C'est la première fois que des travaux de recherche sur la calotte glaciaire permettent de faire enquête sur 1) le rôle de l'eau de fonte dans les variations caractérisant la vélocité de la glace d'une calotte glaciaire polytherme de l'Arctique; 2) l'utilisation de la température de l'air au lieu du bilan radiatif en surface en guise d'approximation pour la fonte superficielle conductrice d'énergie, et 3) l'influence exercée par la fréquence changeante de configurations climatiques synoptiques spécifiques sur la fonte du glacier et le bilan massique. Parmi les autres travaux de recherche, notons des enquêtes sur la géométrie de la calotte glaciaire, la dynamique des débits d'écoulement et le bilan massique; l'analyse des enregistrements relatifs aux carottes glaciaires en ce qui a trait à d'anciens dépôts climatiques et dépôts de contaminants; et l'étude des changements caractérisant l'aire et le volume de la calotte glaciaire de même que leur relation par rapport au bilan massique en surface et à la dynamique des glaces. Par ailleurs, les travaux de recherche actuels prennent la forme d'études sur le terrain se rapportant aux efforts visant à calibrer et à valider l'altimètre radar, études qui seront effectuées par le satellite CryoSat2 de l'Agence spatiale européenne (ASE), et un projet d'envergure en collaboration avec l'Année polaire internationale (API) au Canada portant sur le glacier Belcher, du côté nord-est de la carotte glaciaire, projet qui examine l'hydrodynamique des gros glaciers de marée. La présente communication résume nos connaissances actuelles de la calotte glaciaire de l'île Devon de même que certaines des questions en suspens qui continuent de restreindre la façon dont nous comprenons les interactions entre le climat et la calotte glaciaire dans les régions arctiques.

Mots clés : calotte glaciaire de l'île Devon, Arctique canadien, glaciologie, dynamique des glaces, bilan massique, changement climatique, marée, carottes glaciaires

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INTRODUCTION

Anthropogenic climate change is predicted to be greatest in Arctic regions (Overpeck et al., 1997; Serreze et al., 2000; Johannessen et al., 2004) and will significantly affect the mass balance of Arctic glaciers and their contributions to sea level rise (Lemke et al., 2007). Changes in the Arctic have cascading effects on global systems, making it critical to understand both the changes that are occurring and their impacts. Long-term measurements are required to assess

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the health of Arctic glaciers, but these measurements are time-consuming and costly to obtain, and are therefore limited. The Devon Island ice cap has been the subject of scientific study since 1961, and now has a mass balance record spanning 46 years (Koerner, 2005). This long record provides a unique opportunity to assess variability in surface mass balance and its relationship to climatic variability, and it provides important data that can assist in interpreting the causes of changes in ice cap area and volume.

The Devon Island ice cap is one of the largest ice masses in the Canadian Arctic, the region that contains the greatest total area of land ice in the Northern Hemisphere outside the Greenland ice sheet (Dowdeswell et al., 2004). Devon Island is located in the southeast Queen Elizabeth Islands, Nunavut, Canada (Fig. 1). The ice cap (74˚30′ to 75˚50′ N; 80˚00′ to 86˚00′ W) covers approximately 14 400 km^2. The current ice cap volume is 3962 ± 140 km^3, with a maximum ice thickness of ~880 m (Dowdeswell et al., 2004). The ice cap summit is at 1901 m above sea level (asl). The eastern margin of the ice cap faces the North Open Water (NOW) polynya at the head of Baffin Bay, and large outlet glaciers descend to sea level in this region. Smaller tidewater glaciers drain the northern and southern ice cap margins. By contrast, the western margin terminates entirely on land at 300 – 500 m asl. An arm extending approximately 80 km to the southwest of the main ice cap consists largely of near-stagnant, ablating ice that lies almost entirely below the equilibrium line altitude (ELA) as defined by mass balance data from the northwest sector of the ice cap compiled by Koerner (2005), which averaged 1166 ± 240 m asl from 1961 to 2003.

Devon Island ice cap studies began with researchers from the Arctic Institute of North America (AINA) and their colleagues. That three-year research program (1961 – 63), financed by American and Canadian agencies, was designed to "understand the interrelationships between the glacier-ice of Devon Island, the ocean in Jones Sound, and the encompassing atmosphere" (Apollonio, 1961:252). The expedition's glaciological research focused mainly on the Sverdrup Glacier and the northwest side of the ice cap, where two summer meteorological stations devoted to micro-meteorological studies were established at 1300 m (1961 – 63) and 1800 m asl (1963) (Holmgren, 1971a). Glacier movement studies were conducted in 1961 and 1963, mainly at elevations below 500 m asl on the Sverdrup Glacier (Cress and Wyness, 1961). Glacier mass balance measurements were initiated by Koerner during this expedition and were continued during his tenure at the University of London (1964 – 67), the Polar Continental Shelf Project (1969 – 79), and the Geological Survey of Canada (GSC) (1979 – 2008).

Various others have worked on the ice cap in collaboration with either the AINA group or a GSC group. More recently, researchers from the University of Alberta have combined remote sensing techniques with field studies to investigate large-scale ice cap dynamics, as well as changes in ice cap geometry that are currently taking place. Work on the Devon Island ice cap has also explored mass balance, climate, paleoclimate records in ice cores, and aerosol and anthropogenic contaminant deposition. The Devon Island ice cap is now one of the most intensively studied large ice masses in the circumpolar Arctic.

This paper synthesizes and evaluates research conducted to date on the Devon Island ice cap and identifies the outstanding questions that continue to limit our understanding of the response of Arctic ice caps to climate variability and change. It is also a tribute to the legacy of the late Roy "Fritz" Koerner, who pioneered and led much of the work on the ice cap and who passed away in May 2008 shortly after returning from his last field season on the ice cap.

ICE CAP GEOMETRY

Knowledge of ice cap geometry (area, surface and bed topography, ice thickness) at specific points in time provides vital data from which temporal changes in areal extent and ice thickness can be assessed. It also provides baseline data from which water flow paths and ice dynamics can be calculated. Initial estimates of the geometry of the Devon Island ice cap were based on point or transect measurements. Current methods are able to produce more complete maps and digital elevation models of the ice cap, but are based on old (1959 – 60) aerial photography or satellite imagery (ERS and RadarSat SAR [Synthetic Aperture Radar]; SPOT) with limited ground control points. Elevation errors are large at some locations, particularly in the ice cap interior. The increasing use of airborne and ground-based radio-echo sounding, GPS techniques, and airborne and satellite laser altimetry has greatly improved the accuracy with which ice thicknesses and surface elevations can be mapped. These new methods allow the assessment of temporal changes in ice thickness and surface elevation and the mapping of subglacial topography, which are prerequisites for detailed studies of ice cap dynamics and change.

Ice Area

An early estimate of ice cap area from 1959–60 aerial photography (~15 600 km^2; Koerner, 1970b; Paterson, 1976a) has been more accurately calculated from the same data as 14 738 ± 40 km^2 (Burgess and Sharp, 2004). LandSat 7 imagery from 1999 – 2000 suggests that the current ice cap area is 14 400 km^2 (Burgess and Sharp, 2004). Data managed by Natural Resources Canada give an ice cap area of 12 794 km^2 (Abdalati et al., 2004), which excludes the southwest arm of the ice cap.

Ice Surface Topography

Measurements of ice surface elevation across the entire ice cap were first derived photogrammetrically around 1962 from 1:60 000 aerial photography obtained during the National Air Photo Mapping Program in 1959 – 60. These data provide useful background information for delineating

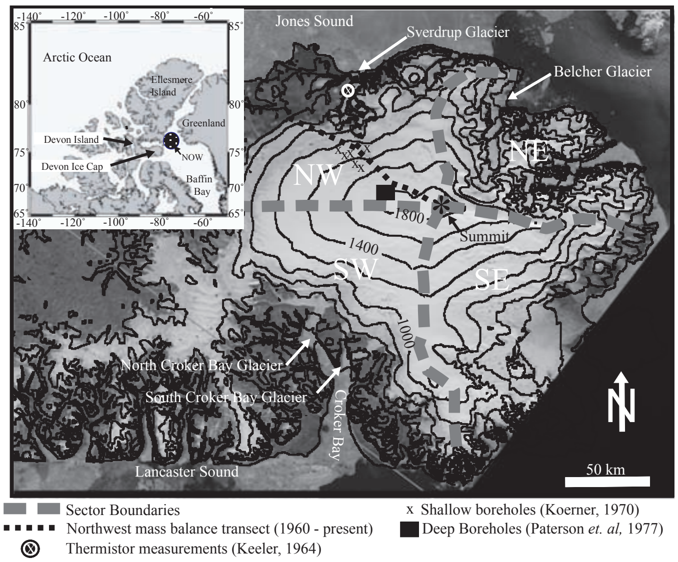

FIG. 1. Location of the Devon Island ice cap in the Canadian Arctic. NOW = North Open Water polynya. Contour lines are in meters above mean sea level. Mass balance sectors are labeled (from Mair et al., 2005). Locations of field-based mass balance measurement transects, ice core boreholes, shallow cores in the superimposed ice zone, and englacial thermistor string measurements are shown (from Keeler, 1964; Koerner, 1970a, b; Paterson et al., 1977).

drainage basins and calculating area-elevation distributions for mass balance studies. However, their poor accuracy (± 30 m near margins and ± 100 m in the interior regions; A. Gagne, pers. comm. 2002) limits their usefulness as baseline measurements against which to monitor changes in ice elevation.

Additional surface elevation measurements with ± 7 – 8 m accuracy were acquired along 3370 km of aircraft track during an airborne radio-echo sounding (RES) campaign over the Devon Island ice cap in 2000 (Dowdeswell et al., 2004). These measurements were based on differentially corrected GPS measurements of aircraft position and altitude, together with aircraft terrain clearance data from the RES system. A 1 km grid cell digital elevation model (DEM) was constructed from these data (Fig. 2). Since the accuracy of these elevation measurements over the snow-covered interior regions of the ice cap is much better than that of the photogrammetric measurements, this dataset provides the most accurate map of the ice cap's surface topography currently available. More recently, ice cap DEMs have been developed from 2007 SPOT 5 high-resolution stereo imagery (± 4 – 9 m vertical error near the ice cap margins); however, problems remain with the ice cap interior (N. Barrand, pers. comm. 2008).

The Devon Island ice cap has a relatively simple dome-like shape, with a maximum elevation of 1921 m asl. Ice divides run east, north, and south from this high point. The western side of the ice cap slopes gently down to a terrestrial margin. To the north and south of the main east-west divide, mountain ranges protrude through the surface of the dome. Bedrock is exposed along the steep valley walls of the major outlet glaciers that drain much of the eastern half of the ice cap and usually terminate in tidewater. These outlets occupy canyon-like valleys similar to those that can be seen on unglaciated terrain beyond the western ice cap margin. The southeastern sector of the ice cap consists of an extensive, gently sloping, piedmont area that is not well constrained by mountain topography and which terminates in the ocean. The area-elevation distribution of the ice cap is rather uniform, although there is a minor peak at 1000–1100 m asl (Dowdeswell et al., 2004).

Ice Thickness

The earliest ice thickness measurements on the ice cap were made using gravity techniques (Hyndman, 1965) and DC resistivity soundings (Vogtli, 1967). The areas surveyed were in the northwest region, in the general area of the Sverdrup Glacier and surrounds, and along the southern margin near Croker Bay (Fig. 3). Ice thicknesses of up to 700 m were measured, and the two techniques produced comparable results when applied in the same area. Subsequently, 440 MHz RES measurements made over an 80 km^2 area at the ice cap crest detected ice thicknesses of 300–750+ m (Paterson and Koerner, 1974), while 620 MHz measurements along two traverses (Fig. 3) revealed that the mean ice thickness was greater in the southeast than in the northeast (568 vs. 426 m) (Koerner, 1977a).

The most comprehensive ice thickness measurements available were obtained in 2000 by airborne 100 MHz ice-penetrating radar (Dowdeswell et al., 2004). A maximum ice thickness of ~880 m was measured at the head of the large outlet glaciers that drain the southeast sector of the ice cap. Ice thickness generally increases rapidly away from the ice cap margins and more slowly and monotonically towards the main ice east-west trending divide. Overall, ice is thickest (700 – 880 m) in the southwest region of the ice cap (Fig. 3). Ice thicknesses are less than 350 m in the large (400 km^2) piedmont area of the ice cap's southeast region and at the northeast margin.

Bed Conditions and Topography

Dowdeswell et al. (2004) used surface elevation and ice thickness data derived from airborne RES to map bed topography beneath the ice cap at 1 km spatial resolution (Fig. 4). The bed consists largely of an upland plateau dissected by a number of steep-sided valleys that control the locations of major outlet glaciers, including the large glaciers that drain into the piedmont lobes in the southeast. About 8% of the bed lies below sea level. A number of the larger outlet glaciers are grounded between 200 and 400 m below sea level for as much as 20 km upstream from their margins; much of the southeast piedmont lobe is also grounded below sea level.

The glacier bed in these regions may be composed of unlithified marine sediments deposited during past periods of reduced ice extent. Recent mapping by the Canadian Coast Guard Ship (CCGS) *Amundsen* in front of the Belcher Glacier terminus, which drains the northeast part of the ice

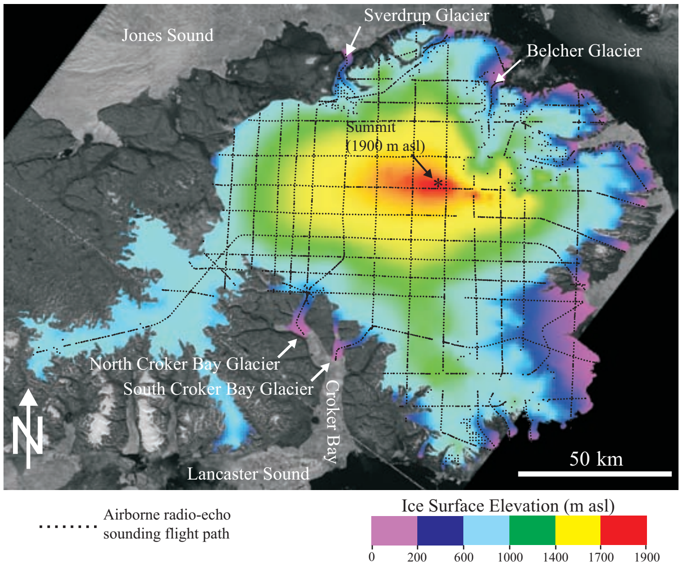

FIG. 2. Most recent ice surface elevation model of the Devon Island ice cap. This digital elevation model (DEM) is derived from airborne radar surveys completed in 2000 (from Dowdeswell et al., 2004).

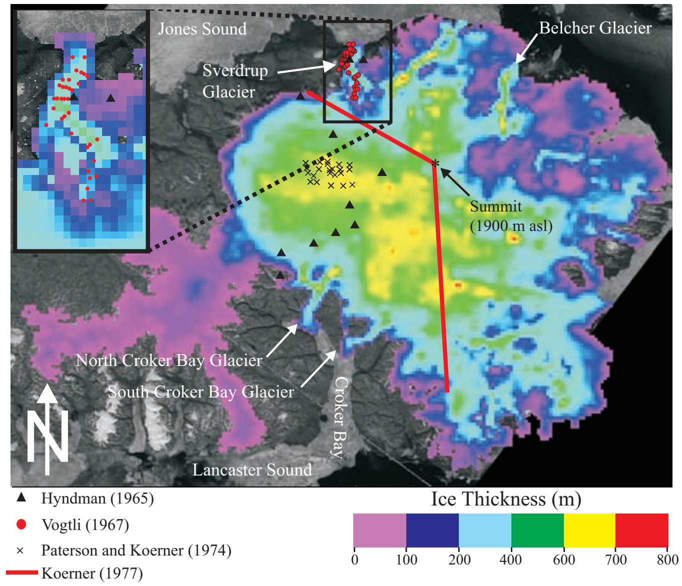

FIG. 3. Ice thickness distribution across the Devon Island ice cap from airborne radar surveys completed in 2000 (from Dowdeswell et al., 2004) and locations of previous ice thickness surveys (from Hyndman, 1965; Vogtli, 1967; Paterson and Koerner, 1974; Koerner, 1977a).

cap, shows a seafloor that is fluted and crisscrossed by a series of deglacial ice marginal moraine ridges (T. Bell and J. Hughes-Clark, pers. comm. 2006).

The picture of subglacial topography that emerges from the recent airborne radar surveys is consistent with what was deduced on a more local basis by earlier work (Hyndman, 1965; Paterson and Koerner, 1974). Some of this early work also provided information on bedrock geology. Hyndman (1965) interpreted the results of his gravity measurements to suggest that the bed consisted of PreCambrian metamorphic rock in the northwest and sedimentary limestone in the southwest, while Oswald (1975) noted that RES data indicated a decrease in bedrock permittivity from west to east across the north-central part of the ice cap. The topographic high detected with both gravity and RES techniques controls the ice divide location (Fisher, 1979).

Thermal Regime and Ice Rheology

In the summit region, the Devon Island ice cap is cold-based, as determined from temperature measurements in boreholes (Paterson, 1976b) (Figs. 1 and 5). However, the fast-flowing outlet glaciers that form in bedrock valleys are likely warm-based (Dowdeswell et al., 2004). Analysis of flow regimes using surface velocities derived from SAR interferometry suggests that the central region of the ice cap is frozen to the bed, while the velocity increases toward the ice cap margin are likely a function of basal lubrication and, near outlet glacier termini, of basal sediment deformation (Burgess et al., 2005).

Ice near the glacier surface is at temperatures well below the freezing point. Thermistors installed in shallow boreholes in the ablation zone of the Sverdrup Glacier measured ice temperatures of -15˚C at 15 m depth (Keeler, 1964), and 12 m ice temperatures of -18˚C were measured at 1300 m asl by Koerner (1970b). An ice temperature profile from a borehole at the summit showed surface temperatures 1.5˚C warmer than predicted with a temperature model, a finding possibly related to climate warming (Paterson, 1976b) (Fig. 5). Further study revealed that latent heat release due to refreezing in the top layers is the more likely driver of the unexpected temperature difference, as it can increase the 10 m ice temperature by ~2.8˚C relative to the current surface temperature (Paterson and Clarke, 1978).

An ice core extracted at the summit contained a 1.7 m section of ice that ended 2.6 m from the bed and is late Wisconsinan in age (Koerner, 1977a). This ice is softer and more deformable than the overlying Holocene ice, largely because of its greater microparticle and impurity content (Paterson, 1976a; Paterson et al., 1977; Koerner and Fisher, 1979; Fisher and Koerner, 1986). Flow models applied to the ice divide region of the Devon Island ice cap successfully predicted the surface profile using a formulation of glacier-bed interaction that included softer Wisconsinan ice near the bed, even though a firn-free vertical profile was assumed (Reeh and Paterson, 1988).

MASS BALANCE

Mass balance data, particularly knowledge of mass balance–elevation gradients, assist in determining the causes of changes in ice cap geometry. Mass balance is closely linked to climatic conditions, which control rates of accumulation and ablation. It is also linked to glacier surface meltwater production, which has implications for ice dynamics. Fully spatially distributed mass balance measurements on the Devon Island ice cap are limited by its spatial scale, and the long-term record relates only to the northwest sector of

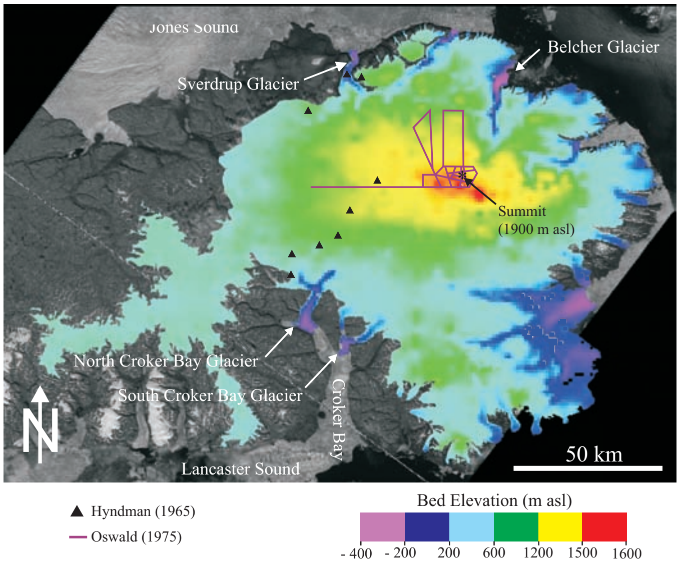

FIG. 4. Bed elevation DEM of the Devon Island ice cap derived from airborne radar surveys completed in 2000 (from Dowdeswell et al., 2004) and locations of previous bed condition measurements (from Oswald, 1975; Hyndman, 1965).

the ice cap. However, these stake measurements and additional estimates derived from ice core measurements can be used to validate the results of numerical models of ice cap–wide mass balance. Both direct measurements and model-derived estimates of glacier mass change are valuable for assessing the ice cap's contribution to sea level change over recent decades.

Local Climatology and Meteorology

The Devon Island ice cap is considered to be in a polar desert, as mean annual precipitation is under 200 mm (Gardner and Sharp, 2007). Mean monthly air temperatures are well below 0˚C in all months, with maximum mean monthly air temperature in July. Mean air temperature lapse rates between 15 and 1320 m asl vary at both daily and monthly scales (0.015–0.096˚C km^{-1}); thus, calculation of lapse rates across the ice cap requires knowledge of local meteorological conditions (Holmgren, 1971a). Snowmelt begins between early June and early July, and the ablation season is usually characterized by a series of short melt periods punctuated by episodes of melt cessation under inclement conditions.

Meteorological data from the summit indicate that air temperature inversions and katabatic winds are significant features of the regional climatology that also affect accumulation and ablation. For example, the melting of snow from surrounding land surfaces causes a temperature differential between ice and land, creating an "edge effect": a ring of cumulus clouds around the edge of the ice cap (Holmgren, 1971a; Lamoureux et al., 2002). While clear skies can persist above 1000 m asl, overcast conditions and rain may occur below this elevation.

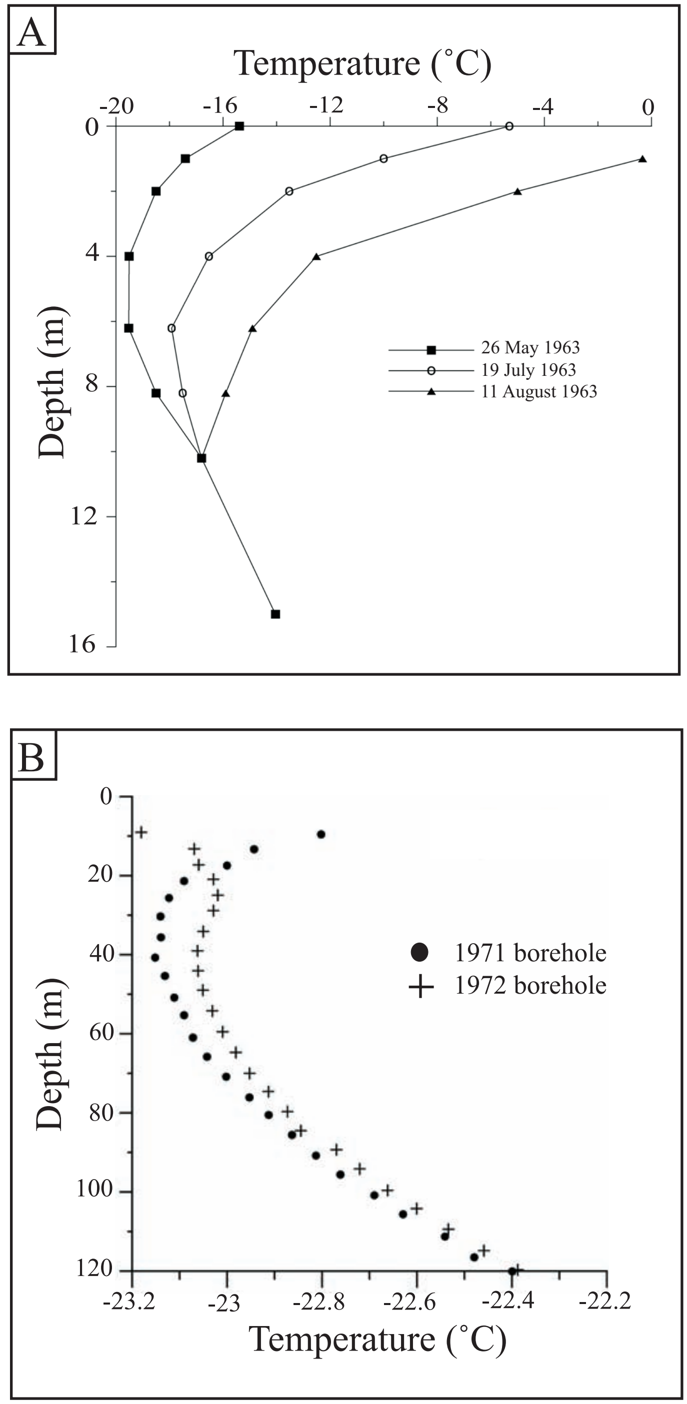

FIG. 5. Englacial temperature profiles: (A) in the top 14 m of the Sverdrup Glacier (from Keeler, 1964) and (B) in the top 120 m of a summit borehole (from Paterson and Clarke, 1978).

Historical Perspective

The margin of the retreating Innuitian ice sheet reached a position close to the northern, eastern, and southern margins of the current Devon Island ice cap by about 10 ^{14}C ka BP, but retreat to the current western margins occurred significantly later, between ~9 and 8 ^{14}C ka BP (Dyke, 1999; England et al., 2006). The pattern of retreat after that time is not well known, but analyses of sediment cores from a proglacial lake northwest of the ice cap suggest that the ice cap retreated from its northwest margin during the early Holocene because of regional warming, then re-advanced after the mid-Holocene climatic optimum (Lamoureux et al., 2002). Ice core studies confirm that the ice cap survived the Holocene period, that the warmest period was in the early Holocene, and that mass balances (as inferred from correlations with the abundance of melt features in ice cores) were largely negative until approximately 4 ka BP (Koerner, 2005). Ice core–derived accumulation rates have remained within 5% of current values for the past 1500 years; for the 3500 years prior to that, the accumulation rate was ~10% greater than at present (Paterson and Waddington, 1984). Thus, regrowth of the ice cap likely occurred within the last 4000 years. The southwestern arm of the ice cap cuts across the early Holocene ice retreat pattern, suggesting that this area was reglaciated during the late Holocene (Dyke, 1999).

Air temperatures declined by 2.7 – 3.5˚C after 5 ka BP, a trend that is largely related to changes in the solar constant (i.e., a 2500-year periodicity related to changes in solar activity). Cloudiness and precipitation from marine sources also decreased over the same period (Fisher and Koerner, 1981). There was extensive reduction of perennial snow and ice cover in the area west of the main ice cap between the Little Ice Age (1680–1730 in this area) and 1960, associated with a rise of more than 500 m in equilibrium line altitudes to elevations greater than those of the current glacier surfaces (Wolken et al., 2008a). These changes suggest a climate warming of around 2˚C over the same interval (Wolken et al., 2008b). Studies indicate significant climate warming over the past half-century in this region (Serreze et al., 2000).

Stratigraphic studies suggest minimal melt during the Little Ice Age, with high summer melt in the more recent past (Paterson et al., 1977). Ice cap mass balances inferred from melt features in ice cores were near zero from 1300 to 1860 AD, but have been mainly negative since then (Koerner, 1977b, 2005).

Annual melt feature thicknesses are positively correlated with annual summer maximum open water area in the Queen Elizabeth Channels, suggesting that warm summer air temperatures drive both ice cap and sea ice ablation. This view is strongly supported by data from 1962, which had the greatest percentage of open water, the greatest melt feature thickness, and the most negative mass balance of the 1961 – 77 period (Koerner, 1977b).

Accumulation

Measurement of winter and summer accumulation patterns on the Devon Island ice cap began in 1961 (Koerner, 1966b). Stake-based data from the Sverdrup Glacier in the northwest sector of the ice cap showed little variation in accumulation with elevation (Koerner, 1961). From 1962 to 1966, measurements were expanded to include the southeast, northeast, and southwest sectors of the ice cap (Fig. 1). Long-term measurements, however, have been collected only in the northwest sector. At higher elevations, winter snow redistribution due to katabatic winds results in an increase in accumulation from the centre of the ice cap towards the margins, with the minimum accumulation encircling the ice cap at elevations 100 – 200 m below the summit (Koerner, 1966b). The 1960s firn line elevation also varied across the ice cap: 1490 m asl in the northwest sector, 1400 m asl in the southwest sector, and 1370 m asl in the southeast sector, which is a location that sees maximum accumulation and minimal ablation (Koerner, 1966b).

Net accumulation (the difference between summer melt and the sum of annual snowfall, superimposed ice, and internal accumulation) increases with elevation on the ice cap (Koerner, 1966b). Superimposed ice formation is a major contributor to net accumulation. In the 1960s it was most significant from 1300 to 1500 m asl, but given its dependence on snow depth and the altitude of the annual snow line, it is highly variable across the ice cap (Koerner, 1970b). Since 1998, the elevation of superimposed ice formation has likely increased (Colgan and Sharp, 2008). Summer accumulation is a highly variable contributor to net accumulation on the ice cap; it is derived either directly from snowfall or from refreezing of rainfall in firn (Barr et al., 1967). Summer snowfall amounts are positively correlated with elevation and are greatest in the southwest sector of the ice cap (Koerner, 1966b).

Maximum accumulation occurs in the summer and early fall (July–October), and analysis of shallow firn cores suggests that interannual variability in accumulation rates is linked to changes in sea ice extent and concentration in the region surrounding the Queen Elizabeth Islands, as well as to changes in the trajectory of air masses reaching the ice cap during this period (Colgan and Sharp, 2008). These trajectory shifts are driven by changes in the strength of the quasi-stationary low in Baffin Bay.

Although average total accumulation over the whole ice cap shows minimal interannual variability (Koerner, 1966b), variability is greater in the west and northwest sectors. Average annual snow accumulation is ~10 cm water equivalent (w.e.) in the northwest sector, but increases to ~50 cm w.e. at the southeast margin (Koerner, 1970a). Accumulation in the southeast sector is enhanced by the proximity to the NOW, which brings frequent cyclonic systems and substantial snowfall in both winter and summer. $\delta^{18}O$ measurements taken along a southeast-northwest transect suggest that 30% of the snow at sea level in the southeast sector of the ice cap is derived from evaporation sources in the Baffin Bay region, a proportion that decreases inland (Koerner and Russell, 1979).

Net accumulation rates have been calculated for various sectors of the ice cap using shallow firn cores and deep ice cores (Table 1). Results from these studies cover from 11 to 40 years and range from 0.13 to 0.27 m w.e. a^{-1}. Values at the lower end of the range are largely in the superimposed ice zone, where melt reduces net accumulation more significantly than at the summit.

TABLE 1. Net accumulation rates determined for different ice cap regions.

Region	Net Accumulation Rate (m w.e. a^{-1})	Time Period	Method	# of Sites	Reference
> 1200 m asl	0.13 – 0.27	1963 – 2003	Shallow firn cores	13	Colgan et al., 2008
1400 – 1800 m asl	0.17 – 0.25 ± 0.06	1963 – 2003	Shallow firn cores	5	Colgan and Sharp, 2008
W sector	0.127 – 0.241	1963 – 2000	Shallow firn cores	7	Mair et al., 2005
E sector	0.217 – 0.241	1963 – 2003	Shallow firn cores	2	Mair et al., 2005
Summit	0.245	1963 – 98	Ice core	1	Kinnard et al., 2006
Summit	0.23	1963 – 98	Ice core (^{137}Cs)	1	Pinglot et al., 2003
Summit	0.24	1963 – 98	Ice core (^{3}H)	1	Pinglot et al., 2003
Summit	0.22	1963 – 74	Ice core	1	Koerner and Taniguchi, 1976
1300 m asl (SI zone)	0.14	1963 – 72	Ice core	1	Koerner and Taniguchi, 1976

Ablation

Summer melt conditions are the main driver of interannual variability in annual net mass balance on the Devon Island ice cap (Koerner, 2005; Gardner and Sharp, 2007), although above 1100 m asl mass balance variability is driven mainly by accumulation variability (Colgan and Sharp, 2008). Ablation is negatively correlated with ice cap elevation above sea level, and also with distance from Baffin Bay. The temperature regimes of the high- and low-elevation regions of the ice cap can become partially decoupled in years of low melt, and more closely related in years of high melt: years of low melt correspond with steeper air temperature lapse rates across the ice cap, while high melt years produce shallower rates (Wang et al., 2005). Drainage basins on the east side of the ice cap lose more mass to surface ablation and iceberg calving than basins on the west side, as a larger proportion of their surface area lies at lower elevations (Burgess and Sharp, 2004; Burgess et al., 2005).

Calculations of energy balance contributions to melt over the 1961 – 63 period indicated that, since absolute amounts of annual accumulation and ablation are both small, days with high melt rates are of major importance for mass balance. Such days are generally characterized by high winds and warm air advection over the ice cap, with clear skies and high incoming radiation (Koerner, 1961; Holmgren, 1971b). Summer rainfall can also enhance ablation by saturating the surface snow layer and promoting runoff and slush movement (Barr et al., 1967). The role of radiation in melt production decreases, however, as the melt season progresses and contributions from sensible and latent heat increase (Keeler, 1964) (Fig. 6). While upglacier winds can bring fog from Jones Sound and decrease melt overall, they can also increase the contribution of latent heat to melt production because of the high vapour pressure gradient (Keeler, 1964; Alt, 1978; Koerner, 2005). Air temperature is considered a better proxy for ablation than radiation, as days with high net radiation can also be cool, thereby reducing melt (Keeler, 1964).

Synoptic Weather Influences on Ice Cap Mass Balance Variability

Local meteorology is influenced by the tendency of cyclonic systems to move from south to north along western Greenland and stall over Baffin Bay. In winter, these storms—combined with the open water of the NOW—raise air temperatures and shift isotherms to run north-south across Devon Island (Holmgren, 1971a). Summer cyclonic activity is highly variable. In late summer (August–September), minimum sea-ice extent causes moist, cloudy conditions, which reduce diurnal air temperature variations by decreasing net radiation losses during the overnight period (Holmgren, 1971a). In the fall (October–November), katabatic winds on the northwest sector of the ice cap are generally from the southeast and often cause a temperature inversion.

Given the role of specific air masses in contributing to summer melt and the importance of summer balance variability for annual net balance variability, regional synoptic weather systems were classified to determine their impact on summer conditions across the ice cap during 1961–74 (Alt, 1978). Strong Baffin Bay cyclones decreased melt and increased summer snowfalls, while increased anticyclonic activity enhanced melt on the outlet glaciers but not over the ice cap interior. In contrast, major anticyclonic blocking in combination with warm air advection substantially decreased mass balance, removing in a single year positive mass balances accumulated over periods as long as five years. Interannual mass balance variations are thus controlled largely by the occurrence of very warm summers (e.g., 1960, 2001, 2005, 2007, 2008), which can eliminate the effect of positive balances caused by several cold summers.

Remote sensing studies of the extent and duration of melt on the Devon Island ice cap over five summer seasons support these results. Analysis of QuikSCAT data shows that melt duration depends on elevation and whether or not the region faces Baffin Bay. Interannual variations in melt duration are closely related to July 500 mb heights over the Queen Elizabeth Islands because of their impact on near-surface air temperature gradients: high mean geopotential heights result in shallow surface air temperature lapse rates and increased melt extent and duration, while low heights produce steep lapse rates and reduced melt extent and duration (Wang et al., 2005).

Alt (1987) defined three regional synoptic weather configurations associated with extreme mass balance years: (1) high melt due to a high pressure ridge from the south at all levels in the troposphere; (2) suppressed melt due to a deep, cold trough across Ellesmere Island; and, (3) summer accumulation due to cold polar lows tracking south/southeast from the Arctic Ocean (Fig. 7). The strength and position of the July 500 mb circumpolar vortex was observed to drive mass balance variability because of its impact both

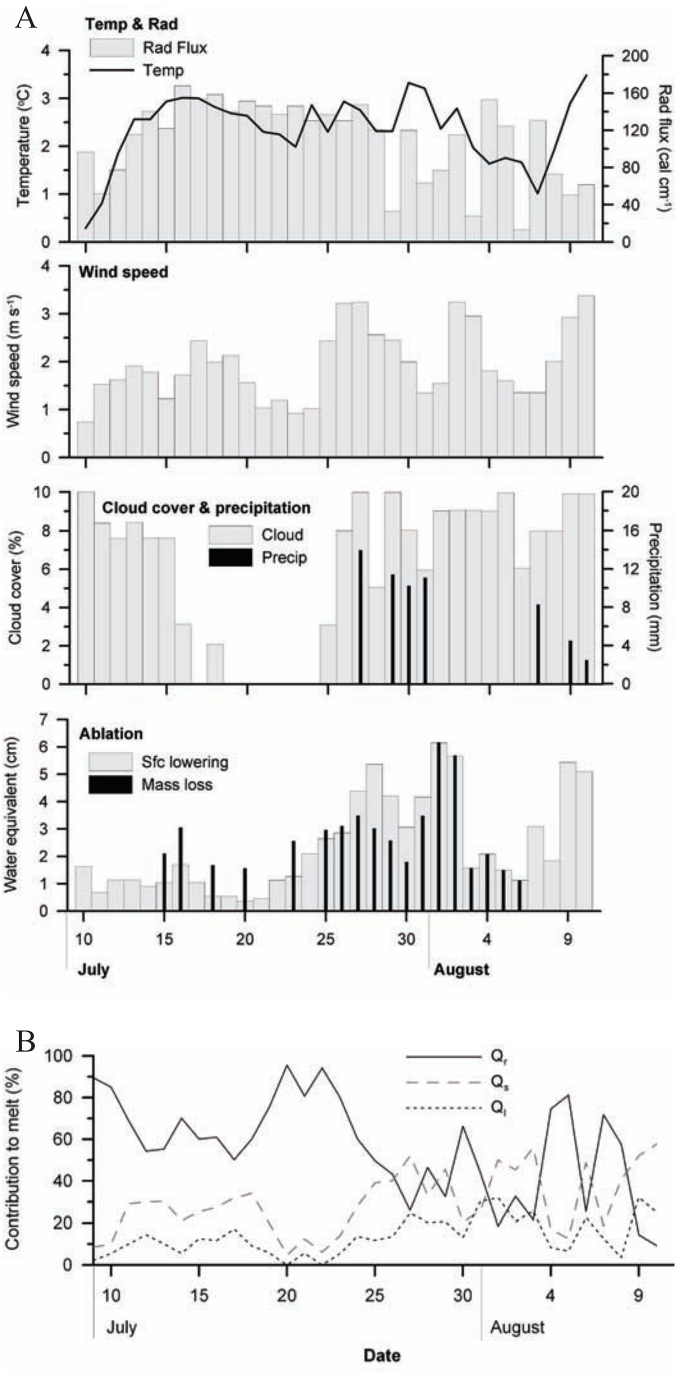

FIG. 6. Measurements from the Sverdrup Glacier, summer 1963: (A) net radiation and air temperature, wind speed, cloud cover and precipitation, mass loss, and surface lowering; and, (B) declining contribution of net radiation (Q_r) to melt over time, compared to sensible (Q_s) and latent (Q_l) fluxes (Keeler, 1964). Note that mass loss was measured from core samples of the surface weathering rind, while surface lowering ablation was calculated from ablation stake measurements multiplied by a constant density value.

on summer ablation and on the occurrence of high pressure ridging over the southern Queen Elizabeth Islands. Since 1987, the summer circumpolar vortex has shifted to the east and weakened, increasing ridging and surface air temperatures over the Queen Elizabeth Islands and resulting in more extreme melt years and more negative mass balances (Gardner and Sharp, 2007).

ICE DYNAMICS

Knowledge of the velocity field of an ice mass is required to understand the distribution of ice temperature and age within an ice cap and the provenance of ice in different regions of the ice cap, as well as to compute rates of mass loss by iceberg calving. When combined with knowledge of the surface mass balance distribution, information about the velocity field helps us to understand the important influence of these factors on patterns of ice thickness change across an ice cap. Knowledge of ice flow rates is also necessary for elucidating spatial and temporal variability in the mechanisms of ice flow, knowledge of which is critical to developing an ability to forecast how the ice cap will respond to changes in forcing from the atmosphere and ocean. Recent advances in radar remote sensing technology (SAR interferometry and speckle tracking techniques) have made it possible to map surface velocities across large areas of glaciated terrain in the polar regions (e.g., Joughin et al., 1995; Rignot et al., 1995), including the Devon Island ice cap.

Spatial Variations

SAR interferograms generated from ERS1 and 2 and RadarSat1 data collected during the 1990s provide a clear picture of the flow regime of the ice cap (Dowdeswell et al., 2004; Burgess et al., 2005; Shepherd et al., 2007) (Fig. 8). Low flow rates (0–10 m a^{-1}) are characteristic of the western part of the ice cap, much of the interior high-elevation regions, and the piedmont region in the southeast. Elsewhere, low flow rates are confined to ridges separating linear regions where flow is 7 to 10 times more rapid (the major outlet glaciers). In some areas (southeast, northeast), the regions of fast flow extend almost to the ice divides (and may be causing ice divide migration), while in others (southwest), they stop well short of the divides. Bedrock pinning points at the heads of glaciers draining the southwestern part of the ice cap appear to control their inland extent, whereas the lack of such controls along the eastward flowing glaciers means there is no such restriction (Dowdeswell et al., 2004; Burgess et al., 2005). In the southeast, the glaciers draining into the piedmont region are fast-flowing in their upper reaches, but fast flow terminates well short of the glacier margin. LandSat 7 imagery shows evidence of past fast flow beyond the current region of fast flow in the form of degraded flow stripes, former marginal shear zones, and folded medial moraines. This evidence suggests that the fast flow may be episodic in nature (i.e., surging), and that it ceased at some point in the past either in the piedmont area alone or over the entire flow system. If the latter is the case, it suggests that fast flow has restarted in the upper regions and may be propagating downglacier, a suggestion that is supported by preliminary results from IceSat altimetry data (G. Moholdt, pers. comm. 2009).

Driving stresses calculated from radar-derived ice thicknesses and surface elevation data are lowest close to the ice divides and in the piedmont lobes to the southeast, but are

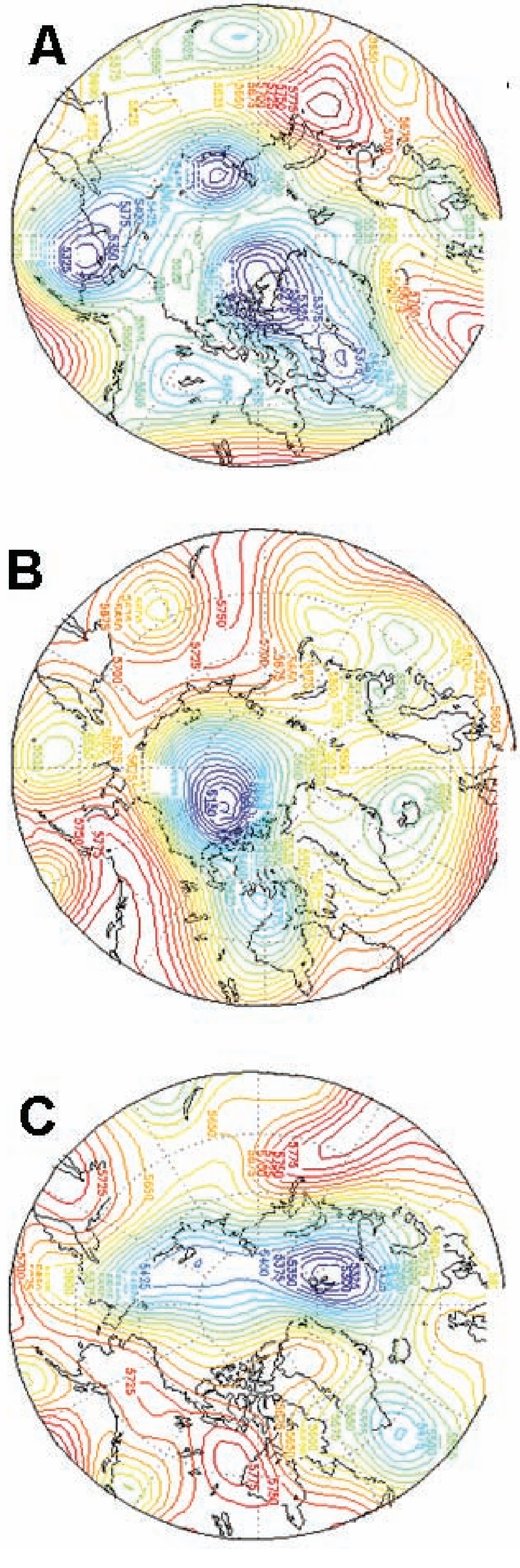

FIG. 7. 500 mb polar synoptic conditions observed to drive ice cap mass balance during 1961–74: (A) Type Ia, 27 June 1963; (B) Type IIa, 30 July 1962; and (C) Type IIIa, 17 July 1962 (from Alt, 1978). Images provided by the NOAA/ESRL Physical Sciences Division, Boulder, Colorado from http://www.cdc.noaa.gov/. Contours show 500 mbar geopotential heights.

often high around the heads of major outlet glaciers (Dowdeswell et al., 2004). Four main flow regime types (FR1–FR4) have been characterized on the basis of patterns of co-variability between the ratio of surface velocity to ice thickness and the local driving stress (Burgess et al., 2005) (Fig. 8). Modes of ice motion change from the interior of the ice cap, through the heads of the outlet glaciers, to their termini. In FR1, ice is suggested to be frozen to the bed, and ice motion is by internal deformation alone. This regime applies to 50% of the ice cap: the interior region above 1000 m asl, the western lobe, and part of the southeast sector below 300 m asl. In FR2, there is probably a small contribution from basal sliding. This regime applies to 22% of the eastern part of the ice cap and 8% of the western part, including the heads of major outlet glaciers (Croker Bay, Southeast), and to small areas along the western margin. In FR3, the contribution from basal sliding is increased, perhaps because surface meltwater is reaching the glacier bed. This regime applies to the lower reaches of major outlet glaciers (Croker Bay, Belcher). Finally, in FR4, basal motion may include a significant contribution from deformation of basal sediments. This regime applies only to the termini of the North Croker Bay, East 3, East Central 1, and South Cunningham glaciers, and part of the Southeast 2 glacier (Burgess et al., 2005). The distribution of FR4 is consistent with Dowdeswell et al.'s (2004) suggestion that marine sediments underlie the lower reaches of these outlet glaciers.

Temporal Variations

Measurements of seasonal velocity variations are limited; they are difficult to acquire with current remote sensing technology and, prior to the advent of GPS technologies, required intensive field study to measure directly. Early measurements of ice surface velocity across the Devon Island ice cap, however, detected significant velocity increases on the Sverdrup Glacier during the melt season (Cress and Wyness, 1961). In 1961, velocity measurements derived from conventional optical surveying techniques revealed flow rates of up to 30 m a^{-1} along four cross-glacier profiles. Maximum vertical movement was measured at 25 cm over the course of the 1961 melt season, with the glacier edges rising and the centre sinking slightly. Cress repeated these measurements in 1963 and also took measurements several times a day in July at a single location at 300 m asl on the Sverdrup Glacier. Ice velocities were observed to increase from 80 to 160 mm d^{-1} over a 12 h period (Cress and Wyness, 1961).

Glacier hydrology may have a major impact on ice dynamics if surface meltwater is able to penetrate the ice mass and reach the glacier bed (Boon and Sharp, 2003). As a result of the low ice surface gradient, meltwater collects at the surface during the snowmelt period, forming large slush fields. Only around mid-July, once the snow cover is removed and the ice surface is exposed, does meltwater channelize into surface drainage (Vogtli, 1967; Holmgren, 1971b). Snow cover removal is then a function of melt rate and slope. Melt rates must exceed the rate of superimposed

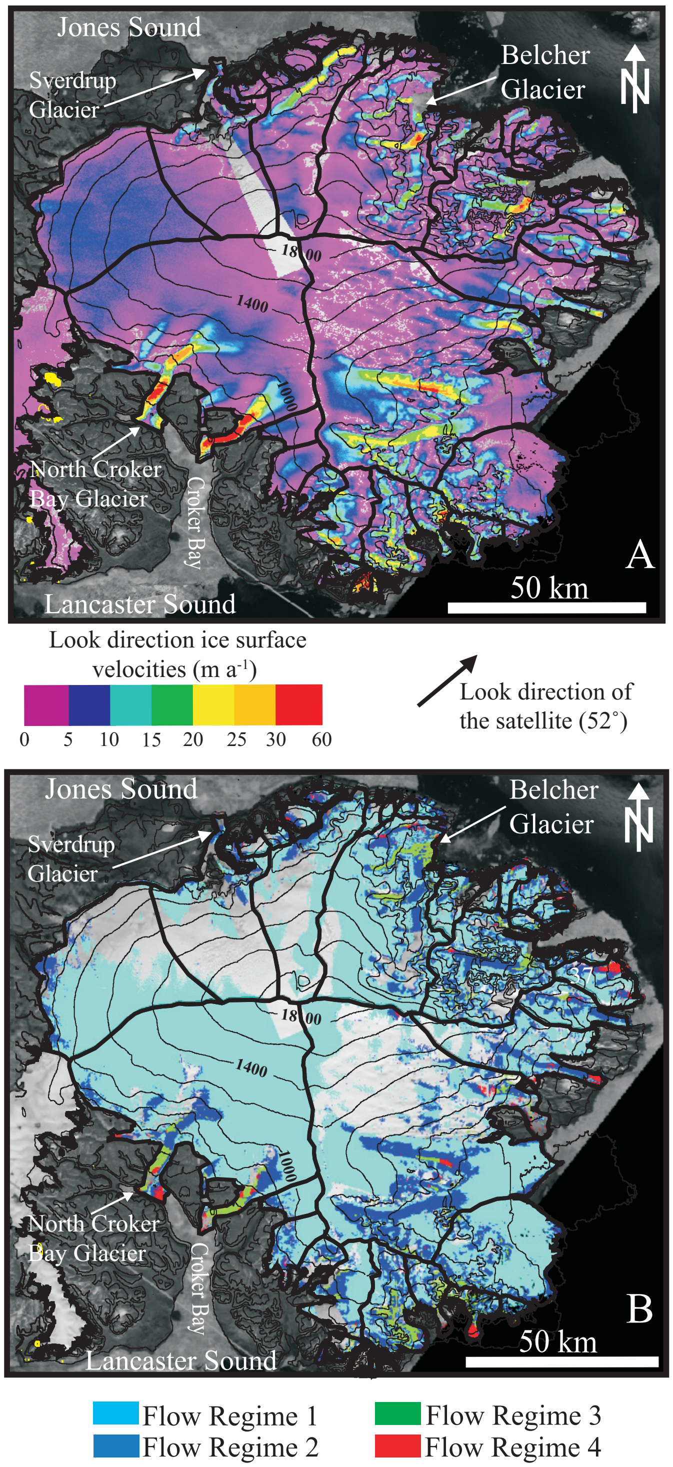

FIG. 8. (A) SAR-derived ice flow velocities across the Devon Island ice cap; and (B) distribution of four flow regime classes derived from surface velocity, ice thickness, and driving stress (from Burgess et al., 2005).

ice formation and internal accumulation to induce runoff, while greater slopes allow for faster drainage of slush into defined surface channels. Drainage on the Sverdrup Glacier was observed to flow mainly ice-marginally, but those surface streams that did develop had a cross-sectional area of 1–30 m^2 and terminated in moulins 400 m from the glacier snout (Koerner, 1961; Keeler, 1964).

Landsat 7 ETM+ imagery shows a dense network of supraglacial channels on the piedmont lobes in the southeast sector of the ice cap. These streams often end abruptly, suggesting that meltwater is entering the glacier via moulins and potentially reaching the glacier bed (Dowdeswell et al., 2004). Meltwater channels also terminate in major crevasse fields on the Belcher and Croker Bay glaciers (Burgess et al., 2005). Observations in the northwest sector of the ice cap indicate that the ice cap is drained supraglacially in this region, as no subglacial meltwater sources were identified (Lamoureux et al., 2002). However, moulins have been observed in this region at ~1250 m asl, where they direct water flow towards the Sverdrup Glacier (Koerner, 1966a). Supraglacial lakes tend to be concentrated on the major outlet glaciers and are rarely found in the ice cap interior or on the western lobe. Some of these lakes have been observed to drain either supraglacially or via crevasses on the lake floor (A. Gardner, pers. comm. 2009). In some cases, collapsed and fractured lake ice cover is observed on basin floors in the spring, suggesting that drainage occurred during the winter after the lakes had frozen over at the end of the previous summer.

ICE CAP CHANGES

Changes in ice cap area, surface elevation, and mass balance can be assessed as a function of initial geometry and ice cap–climate interactions. Ice cap changes result in sea level change and provide a measure of ice cap response to climate forcing. While areal change is simplest to calculate, determining changes in surface elevation and volume is more problematic given the nature of available data.

Area

Ice cap area decreased by 338 ± 40 km^2 (2.4%) between 1959–60 and 1999–2000 (Burgess and Sharp, 2004). This net decrease reflects the combined effects of margin retreat in the southwest sector (greatest contribution), tidewater glacier retreat, increases in exposed bedrock in the interior of the ice cap, and advances along the northwest margin and in the Sverdrup Glacier and Croker Bay areas—the latter of which is the most important calving region in the western sector of the ice cap.

Considering area change basin by basin, the largest percentage reductions in glacier area occurred in basins with a hypsometry such that a 100 m rise in the equilibrium line altitude would have resulted in a large change in the accumulation area ratio (proportion of the basin lying in the accumulation area). These basins, located on the southwest arm and in the south and southeast sectors of the ice cap, are characterized by generally low mean elevations. The net mass balance of these basins would be most strongly affected by a uniform rise in air temperature. Basins with short response times (less than 380 years, as defined by the ratio of the maximum ice thickness in the basin to the mass balance at the glacier terminus) tended to show greater proportional reductions in area than basins with longer response times. Basins with short response times include those on the southwest arm and on the east and south margins that are peripheral to the main ice cap and have accumulation areas that do not extend to the major ice divides. By contrast, basins with longer response times have experienced less area reduction (or even area expansion) and are located along the western margins of the main ice cap.

Ice Surface Elevation

Borehole measurements at the summit between 1971 and 1973 indicated that the glacier was thickening slightly (Paterson, 1976a). However, measurements of sufficient accuracy to quantify changes in surface elevation have so far been restricted to linear profiles. The first such measurements were performed along a 6 km transect up the northwest margin in 1970, using geodetic leveling techniques (I. Whillans, unpubl. data) (Fig. 9). Elevation measurements along this transect were repeated using similar methods in 1973, and again in 2007 using kinematic differential GPS techniques (Burgess and Koerner, unpubl. data), and these data are currently under analysis.

High-accuracy surface elevation measurements (RMS error < 10 cm) along two transects across the Devon Island ice cap were acquired with NASA's Airborne Topographic Mapper (ATM) in 1995 and 2000 (Fig. 9). These measurements on the ice cap reveal a broad pattern of thinning at low elevations and thickening at high elevations between 1995 and 2000 (Abdalati et al., 2004) (Fig. 9); however, this pattern is biased, since measurements were taken only near the main ice divide in western regions of the ice cap, and not near the major outlet glaciers (Colgan et al., 2008). Comparisons of annual net accumulation and measured ice outflow from all major drainage basins above 1200 m asl suggest that thickening of high-elevation regions of the ice cap may be confined to the area of the ice cap in which laser altimetry surveys were concentrated, while thinning may be characteristic of high-elevation regions to the south of the main east-west ice divide (Colgan et al., 2008).

Thickening in the central portion of the ice cap from 1995 to 2000 may be a function of a short period of above-average accumulation (Abdalati et al., 2004), a longer-term imbalance between accumulation rates and vertical strain rates (Koerner, 2005), or a reduction in ice flow from the centre of the ice cap as basal ice stiffens in response to the penetration of Neoglacial cooling to the bottom of the ice cap (Colgan et al., 2008). Both stake and ice core measurements from the high-elevation region of the ice cap indicate that rates of accumulation in the 1995–2000 period were not anomalously high relative to the post-1961 mean (Colgan et al., 2008). Ice core evidence suggests that firnification rates on the ice cap have increased in recent years and were especially high in the 1995–2000 period, so it is unlikely that the observed thickening is due to a reduction in mean firn density.

Since 2004, repeat elevation measurements using kinematic differential GPS methods and airborne laser altimetry have also been made along the CryoSat2 Line, a 48 km long north-south transect on the south side of the ice cap starting slightly west of the ice cap summit and ending at the southern margin of the ice cap (Fig. 9). These surveys show no measurable elevation changes within 10 km of the summit. However, nearly 2 m of thinning were measured downglacier of that point, near the ice cap margin, between 2004 and 2007. Repeat kinematic differential GPS measurements made along an east-west transect across the main north-south ice divide in the southeast region of the ice cap (Fig. 9) show thinning in the upper reaches of the fast flowing outlet glaciers that drain both east and west from the divide, at rates greater than can be explained solely by surface melting (J. Davis, unpubl. data).

Ice Cap Mass Balance

The mass balance of the Devon Island ice cap has been calculated using a range of techniques, including surface balances, estimates of calving only, volume-area scaling, flux divergence techniques, and a combination of remote sensing and in situ data.

The 46-year record of in situ glacier surface mass balance measurements from the northwest side of the ice cap shows generally negative balances with no long-term trend for the first 25 years of measurements, followed by a significant trend towards increasingly negative balances to the present day (Fig. 10; Koerner, 2005). In situ measurements indicate that cumulative mass balance of the Devon Island ice cap (1961–2003) was -3.54 m (Koerner, 2005). A combination of shallow borehole measurements and degree-day modeling indicates that the ice cap lost ~1.6 km^3 of water per year from 1963 to 2000, equivalent to a net mass balance of -0.13 m w.e. a^{-1} (Mair et al., 2005). Volume losses from iceberg calving alone were calculated as 20.5 ± 4.7 km^3 from 1960 to 1999. Of these losses, 67% originated from the northeast sector, with ~50% of total ice cap calving losses coming from the Belcher Glacier alone (Burgess et al., 2005). Volume losses of -72 km^3 (1960–99), calculated from measured area change using volume-area scaling techniques, are equivalent to a ~0.005 mm a^{-1} contribution to sea level (Burgess and Sharp, 2004). More recently, Burgess and Sharp (2008), using flux divergence and net surface mass balance techniques, calculated volume losses for the same period as -76.8 ± 7 km^3, also equivalent to a 0.005 ± 0.0005 mm a^{-1} increase in sea level. Volume change from 1995 to 2000 was calculated as -0.81 km^3 a^{-1} over the ice cap from laser altimetry and climate data (Abdalati et al., 2004), a ~0.002 mm a^{-1} contribution to sea level. An alternative 40-year calculation of 0.003 mm a^{-1} sea level rise from Devon Ice cap melt was made from GPS measurements at on-ice stakes, ice flux calculations, and surface mass balance (Shepherd et al., 2007).

To make direct comparisons between these values, all were converted to m w.e. a^{-1} (Table 2). Where initial values

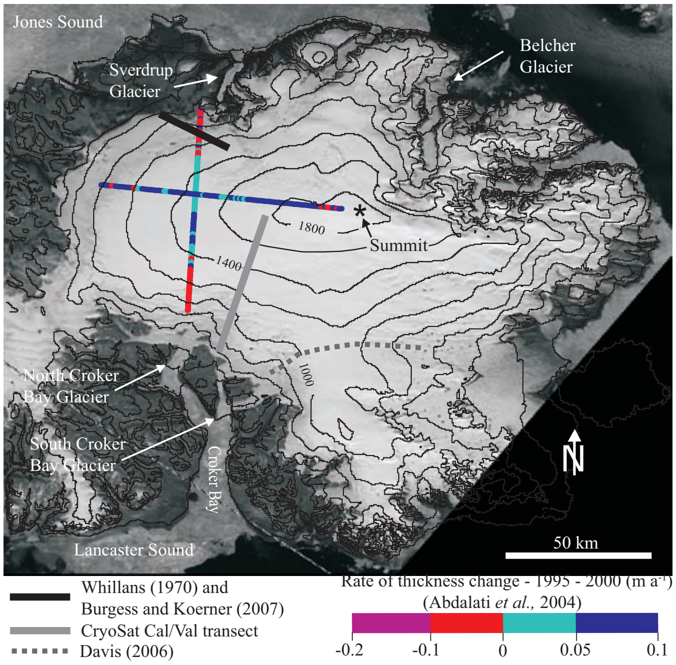

FIG. 9. Surface elevation change across the Devon Island ice cap derived from altimetry data (from Abdalati, 2004) and locations of field and remote sensing measurement transects from which changes in glacier surface elevation can be calculated.

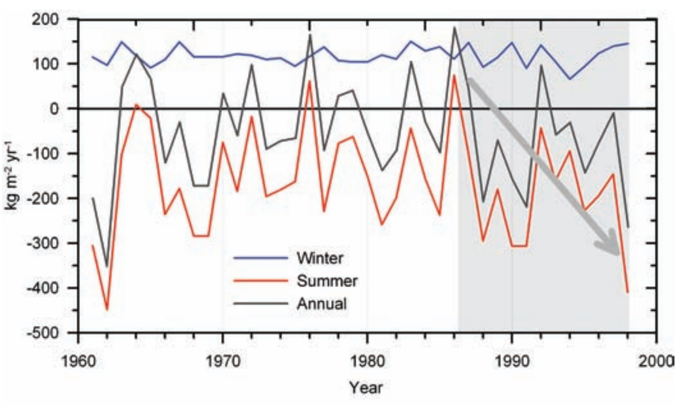

FIG. 10. Long-term (1961–2006) measured mass balance from the Devon Island ice cap (northwest sector), with the post-1987 shift in the circumpolar vortex highlighted.

referred to changes in volume, they were converted to two separate values of water equivalent thickness using assumed densities for firn (600 kg m^{-3}) and ice (900 kg m^{-3}). Original references were used to ascertain whether results cover the whole ice cap, or the ice cap minus the southwestern arm. Mass balance estimates range from -0.038 m w.e. a^{-1} (firn; Abdalati et al., 2004) to -0.17 m w.e. a^{-1} (ice; combined Mair et al., 2005 and Burgess et al., 2005). The surface mass balance, volume area scaling, and flux divergence methods all yield similar results when ice-based calculations are compared (-0.12 to -0.17 m w.e. a^{-1}). The combination methods, however, give significantly less negative mass balances (-0.058 to -0.088 m w.e. a^{-1}). The Abdalati et al. (2004) values are lower because laser altimetry measurements are

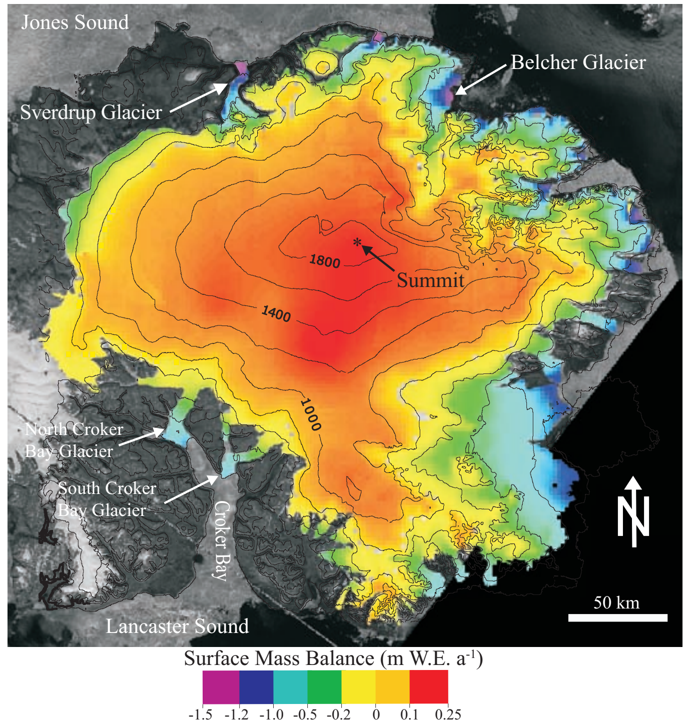

FIG. 11. Spatial pattern of mass balance across the Devon Island ice cap derived from field measurements and numerical modeling (from Mair et al., 2005).

biased away from outlet glaciers and low-elevation regions, which are known to contribute significantly to mass loss. The Shepherd et al. (2007) calculations, however, are lower because balance was simulated for 1996, which was not a very negative balance year (Gardner and Sharp, 2007). Additionally, using a fixed lapse rate to extrapolate air temperatures from ocean buoys to higher elevations underestimates melt at higher elevations (Gardner and Sharp, 2009). Melt at these locations occurs mainly on days with relatively shallow lapse rates (Gardner et al., 2009), which cannot be simulated using this technique. It is also likely that the southwest arm of the ice cap has a greater thinning rate than the main ice cap since it is stagnant and thus not maintained by flow from the centre of the ice cap. Estimates that do not include this arm will therefore be lower than those that do include it.

The distribution of mass balance change varies over the ice cap (Fig. 11). The largest drainage basins cover 86% of the ice cap but contribute only 68% to surface volume loss, while the remaining 14% of the ice cap contributes 32% to surface volume loss (Mair et al., 2005). These findings confirm the results of area change studies, which show that small basins at the periphery of the ice cap, with relatively low mean elevations and accumulation areas that do not extend to the main ice divides, are more sensitive to climatic change (Burgess and Sharp, 2004). Given that surface mass balance estimates from several studies range between -0.12 and -0.17 m w.e. a^{-1} (Table 2), iceberg calving may account for almost 30% of total mass loss over the past 40 years (Burgess and Sharp, 2004; Burgess et al., 2005).

TABLE 2. Ice cap mass balance (MB) estimated using various techniques.

Reference	Technique	Mass Balance (m w.e. a^{-1})	Ice cap area (km^2)
Mair et al., 2005	Surface MB	-0.13 ± 0.06	12 794
Burgess and Sharp, 2004	Volume-area scaling	-0.077 (firn) -0.12 (ice)	14 400
Burgess and Sharp, 2008	Flux divergence	-0.16 ± 0.015	12 794
Burgess et al., 2005	Calving only	-0.042 ± 0.01 (ice)	14 400
Mair et al., 2005 + Burgess et al., 2005	Surface MB + calving	-0.17 ± 0.06	14 400
Abdalati et al., 2004	Combination of methods	-0.038 (firn) -0.058 (ice)	12 794
Shepherd et al., 2007	Combination of methods	-0.088 ± 0.056	12 489

DISCUSSION

Baseline datasets of the bed topography, ice surface topography, and ice thickness of the Devon Island ice cap are now available. Long-term data collection has also provided a lengthy surface mass balance record for the northwest sector of the ice cap, which has been extended to the entire ice cap by a combination of ice core analyses, remote sensing, and numerical modeling. Various studies have assessed changes in ice area and volume since 1959–60, and the surface velocity field of the ice cap has been mapped using SAR interferometry. In combination with radar-derived measurements of ice thickness, these data have allowed an estimate of the rate of mass loss by iceberg calving to be made. The causes of observed spatial patterns of ice thickness and volume change have been investigated by comparing spatial patterns of surface mass balance with spatial patterns of ice flow divergences across the ice cap.

The increased use of remote sensing in studies of ice cap dynamics and change requires ground-truthing, as well as precise determination of what variables are recorded by remote sensing instruments. Calibration and validation experiments conducted for the CryoSat2 campaign have begun to address this issue by collecting concurrent ground-based and airborne measurements of ice surface elevation, along with detailed measurements of the physical properties and stratigraphy of near-surface snow and firn, which can be used to investigate the causes of differences in measurements of surface elevation made using GPS, laser altimetry, and radar altimetry. Remote sensing has expanded knowledge of ice cap behaviour from field measurements in the northwest sector of the icecap to both the southwest and northeast sectors, as well as to the region around the north/south ice divide in the southeast sector. Field data are still required, however, to describe key processes (ice-atmosphere interactions, ice-ocean interactions, ice dynamics, accumulation, superimposed ice formation, internal accumulation, melt, and ice cap drainage) and determine how to best represent them in numerical models of ice cap–climate interactions (Bell et al., 2008).

Although many techniques have been used to estimate ice cap mass balance and its spatiotemporal variability (Koerner, 1977b; Abdalati et al., 2004; Burgess et al., 2005; Mair et al., 2005; Shepherd et al., 2007), and detailed measurements of surface balance exist for the northwest sector (Koerner 1970a, 2005), a more precise mass balance estimate for the entire ice cap is required to more clearly determine the drivers of mass balance change. Research has been limited by the quality of photogrammetric elevation measurements from 1959–60, the lack of detailed field measurements of ice velocity to characterize its temporal variability, and the difficulty of modeling climate, mass balance, and ice flow at high spatial resolution over topographically complex terrain. Additional ground control points are required to improve photogrammetric elevation measurements and determine precise ice cap topography, especially in the ice cap interior.

The 2005 laser altimetry surveys included the acquisition of additional elevation data along the centerlines of seven major outlet glaciers that drain the main portion of the ice cap (W. Abdalati et al., unpubl. data). Repeat surveys along these glaciers are planned to search for evidence of dynamically driven thickness changes, and these should provide insight into the stability of ice flow from the interior regions to the ice cap margins. Analysis of transect-based surface elevation data from 1970 and 1973 (I. Whillans, unpubl. data) and 2007 (Burgess and Koerner, unpubl. data) will provide a useful measure of long-term changes in ice thickness that is independent of the in situ mass balance data collected from this region since 1960. While iceberg calving is an important source of mass loss, calving mechanisms and the factors controlling calving rates are less well known. The influence of tidal forcing, changes in fjord water column properties, and the seasonal removal of sea ice from fjords on ice flow, calving rates, and processes at the Devon Island ice cap have yet to be investigated. Seafloor bathymetry beyond the termini of calving tidewater outlet glaciers can control their rates of advance and retreat, and the extremely limited knowledge of seafloor terrain is a significant constraint on ice cap modeling activities.

The long-term mass balance dataset from the Devon Island ice cap has revealed the increasing impacts of climate change in Arctic regions, reinforcing the value of long-term studies for distinguishing natural climate variability from climate change. Several datasets suggest that over the 46-year record, mass balance has been negative overall, with a trend towards more negative balances since 1985 (Koerner, 2005; Gardner and Sharp, 2007; Colgan and Sharp, 2008). Precise quantification of this trend is a focus of ongoing research that aims to solve the problem of quantifying rates of internal accumulation in a period of warming summer temperatures (Koerner, 2005).

Superimposed ice (SI) formation and internal accumulation are major contributors to annual accumulation, but remain difficult to measure over large areas (Koerner, 1970b, 2005). It is likely that observed changes in both synoptic and long-term climate conditions will alter the locations, rates, and amounts of internal accumulation, and hence the mass balance. Long-term observations indicate that glacier facies zones have already shifted, with upglacier movement of all zones and increased ice content in firn at all elevations (Colgan and Sharp, 2008). Changes in the location and amount of meltwater refreezing within the surface layers of the ice cap will affect ice temperature and viscosity, with eventual consequences for glacier flow rates; more recent studies, therefore, include spatially distributed predictions of the annual amount of internal refreezing from 1980 to 2006 (Gardner and Sharp, 2009).

Cress and Wyness (1961) were perhaps the first to detect a rapid change in ice velocity on an Arctic glacier associated with increased surface meltwater reaching the glacier bed. This phenomenon remains the focus of much glaciological study, as researchers seek to understand the potential impact of climate warming on the stability of tidewater glaciers (e.g., Zwally et al., 2002; Alley et al., 2005). However, measurements of seasonal velocity variations on the Devon Island ice cap are limited. Field-based studies are intensive and costly, and remote sensing studies often resolve velocity variations at annual, rather than seasonal, scales. Short-term field measurements of vertical velocity using differential GPS and surveying would be particularly useful, as they would also help to determine whether meltwater is reaching the bed and driving changes in ice surface elevation. Also absent are studies of how removal of the sea ice cover from fjords in which tidewater glaciers terminate affects ice flow, how increased meltwater outflow into fjords affects the advection of warm ocean waters towards glacier fronts, and the consequences of these effects for rates and mechanisms of iceberg calving.

Accurate modeling of glacier flow also requires some understanding of the distribution of subglacial bed material. It has been suggested that marine sediments underlie some of the outlet glaciers, as their beds are below sea level, and seafloor mapping indicates a till cover on recently deglaciated seafloors (Dowdeswell et al., 2004; Bell and Hughes-Clark, pers. comm. 2006). Local geology has also been assessed by examining bedrock at the ice cap margins (Hyndman, 1965). However, the distribution of rock types beneath the remainder of the ice cap is largely unknown. Bedrock geology has implications for both ice flow and the structure of the subglacial drainage system in areas where water can exist at the bed (i.e., sheet flow, cavities, channels, etc.).

Understanding of the causes of potential seasonal velocity variations is thus limited by a lack of information regarding both the surface and subglacial hydrology of the ice cap and the connections between these two systems. While flow regimes across the ice cap have been derived from InSAR analysis (Burgess and Sharp, 2004), the controls on these flow regimes (such as ice temperature, glacier drainage, and bed properties) are poorly quantified. For example, if the shift from FR2 to FR3 is linked to increased meltwater inputs to the glacier bed, it should be possible to link the transition to sink points for supraglacial streams and to determine whether or not the location of the transition is temporally stable.

To determine the existence of a link between meltwater production, runoff, and ice dynamics, one must identify locations where surface meltwater reaches the subglacial system and establish how these locations relate to the thermal regime and dynamics of the ice cap. There are currently no data on the processes of transition from surface slush fields to channelized supraglacial flow, or on the volume of meltwater flowing on, within, or beneath the ice cap. Similarly, the role of supraglacial and ice marginal lakes in storing surface meltwater and eventually releasing it, either to the glacier bed or to downstream surface drainage systems, has yet to be explored. While surface DEMs are available, they are either too patchy or at too coarse a resolution to identify hydrologic features that may allow surface meltwater to access the glacier bed. Research is currently underway to assess the quality of SPOT5-derived DEMs for hydrological applications.

Subglacial hydraulic potential can now be calculated using maps of ice thickness and bed topography (e.g., Flowers and Clarke, 1999), but interpretation of results requires knowledge of the basal thermal regime. Unfortunately, current knowledge of ice thermal regimes is limited to point borehole measurements and assumptions derived from ice flow patterns (Paterson, 1976b; Burgess et al., 2005). Analysis of radar bed reflection characteristics could help determine where subglacial water is present (Copland and Sharp, 2001), provide a test of inferences regarding the basal thermal regime based on analyses of ice dynamics (i.e., FR1–4), and assist in modeling ice dynamics.

Potential links between dynamics and hydrology require knowledge of rates and patterns of surface melt production, knowledge which is also essential for mass balance studies. Keeler's (1964) work on air temperature and melt rates has led to the development and widespread use of degree-day models (e.g., Braithwaite, 1984; Hock, 1999). However, local meteorological controls on summer meltwater production must be explored in more detail, as existing studies are limited to the Sverdrup Glacier in the northwest sector of the ice cap. Understanding of meteorological controls is especially important given the increasing area of open water in regions such as Jones Sound, which may reduce glacier melt by enhancing fog and reducing air temperature below altitudes of ~350 m asl (Koerner, 2005). Meteorological variables have been monitored on the ice cap since the mid-1960s, and automated weather station data are available from five on-ice locations. These data have provided input to modeling studies (e.g., Mair et al., 2005), and they have recently been analyzed to assess variations in and the role of air temperature lapse rates across the ice cap (Gardner et al., 2009). Further analysis of these datasets is required to assess interannual variability in air temperature, accumulation, and melt. A systematic investigation of changes in the surface energy balance under varying synoptic weather settings is also needed to better understand the link between synoptic weather conditions and melt. This link is particularly important given that the relative frequencies of occurrence of the synoptic weather types recognized by Alt (1978, 1987) have shifted in the period since 1987, contributing to enhanced glacier ablation (Colgan and Sharp, 2008).

More research remains to be done, therefore, to fully quantify surface processes and ice dynamics across the ice cap and their relation with and response to climate and climate change. An integrated program is thus underway to assess summer conditions, glacier hydrology, ice dynamics (seasonal, spatial variability), and bed conditions (bed material, thermal regime).

FUTURE DIRECTIONS

Research on the Devon Island ice cap is now at the stage when data are available to begin examining interrelationships between ice cap geometry, dynamics, mass balance, and atmosphere-ocean conditions. This examination will require the development of whole ice cap models within which higher-resolution models of fast-flowing outlets can be nested, and the use of these models to explore how the response of ice caps to climate change is mediated by changes in meltwater production, glacier drainage, ice dynamics, and calving processes. Many of these issues have been explored as part of an International Polar Year study undertaken by a team of researchers as a contribution to the international Glaciodyn project. This project aimed to examine the interactions between changes in the atmospheric and ocean climates around the ice cap and the hydrology, flow, and iceberg calving fluxes of one of its major tidewater outlets, the Belcher Glacier.

CONCLUSION

Forty-seven years of research on the Devon Island ice cap have contributed to our understanding of the influence of surface meltwater on ice dynamics, the use of air temperature—rather than net radiation—as a proxy for surface melt, and the relationship between summer ablation and regional synoptic weather systems. The length of the mass balance record from the Devon Island ice cap has allowed researchers to observe the effects of recent climate change on the ice cap. These effects include shifts in the synoptic weather systems responsible for summer melt, increased open water extent in Jones Sound, a trend towards increasingly negative mass balance, and changes in the location and extent of glacier facies zones. The comprehensive dataset created by this research permits more detailed modeling studies of glacier response to climate change and potential contributions to sea level rise. These data will be used to further our understanding of ice dynamics – particularly on tidewater glaciers, where high calving rates contribute significantly to annual ablation.

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