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Baffin Island and West Greenland Current Systems in northern Baffin Bay

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ABSTRACT

Temperature, salinity, and direct velocity observations from northern Baffin Bay are presented from a summer 2003 survey. The data reveal interactions between fresh and cold Arctic waters advected southward along Baffin Island and salty and warm Atlantic waters advected northward along western Greenland. Geostrophic currents estimated from hydrography are compared to measured ocean currents above 600 m depth. The Baffin Island Current is well constrained by the geostrophic thermal wind relation, but the West Greenland Current is not. Furthermore, both currents are better described as current systems that contain multiple velocity cores and eddies. We describe a surface-intensified Baffin Island Current seaward of the continental slope off Canada and a bottom-intensified West Greenland Current over the continental slope off Greenland. Acoustic Doppler current profiler observations suggest that the West Greenland Current System advected about 3.8 ± 0.27 Sv (Sv = 10^6 m^3 s^{-1}) towards the northwest at this time. The most prominent features were a surface intensified coastal current advecting 0.5 Sv and a bottom intensified slope current advecting about 2.5 Sv in the same direction. Most of this north-westward circulation turned southward in the Baffin Island Current System. The Baffin Island system was transporting 5.1 ± 0.24 Sv to the south-east at the time that includes additional contributions from Nares Strait to the north (1.0 ± 0.2 Sv) and Lancaster Sound to the east (1.0 ± 0.2 Sv). Net freshwater fluxes were 72 and 187 mSv for the West Greenland and Baffin Island Currents, respectively. Empirical uncertainty arises from unknown temporal variations at weekly time scales and pertubations introduced by unresolved eddies. Eddies with 10 km horizontal and 400 m vertical scales were common and recirculated up to 1 Sv. Our 2003 observations represent conditions when the North-Atlantic Oscillation index (NAO) was close to zero. Analysis of historical hydrographic data averaged along isobaths during NAO-positive years reveals a baroclinic circulation in Baffin Bay more intense than 2003 with stronger southward flow of fresher Arctic waters along Baffin Island and stronger northward inflow of saltier Atlantic waters along Greenland. During negative NAO years this cyclonic circulation weakens as evidenced by a 1979 synoptic survey of the hydrography along Baffin Island.

Introduction

Climate change over the North-Atlantic Ocean causes rising coastal sea level along the US eastern seaboard (Sallenger et al., 2012) and more varied weather (Francis and Vavrus, 2012) which combine to increase the risk of extreme flooding (Lin et al., 2012). Enhanced Arctic freshwater discharge (Serreze et al., 2006), melting of polar ice sheets (Shepherd et al., 2012), thinning and retreating glaciers (Münchow et al., 2014), and the dramatic decline of Arctic summer sea ice (Kwok and Rothrock, 2009) all provide evidence of change and positive feedbacks. We here focus on the flux of relative fresh ocean waters from the polar ocean to the south. We use the North-Atlantic Oscillation (NAO) index of Hurrell and Deser (2009) as a metric to place detailed observations from 2003 into a larger climatological context. First, however, we introduce our study area to the west of Greenland via a historical review of available data that relates to circulation.

On April 30, 1873 the sealer *Tigress* working off coastal Labrador plucked 12 men, 4 children, and 2 women off an ice floe. Fed by two Inuit hunters they had floated on ice for 6 months after the *USS Polaris* abandoned them in Nares Strait to the north of Baffin Bay (Berton, 1988). Inadvertently, they also mapped the surface circulation of western Baffin Bay, traveling on ice floes almost 3000 km at an average speed of about 0.2 m/s. Less fortunate were the 1502 passengers who perished aboard the *RMS Titanic* on April

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15, 1912 when she was sunk by an iceberg off Newfoundland. Most likely, this iceberg originated from Greenland or northern Canada taking a path similar to that of the Polaris survivors. The dramatic loss of life in 1912 led to the formation of the International Ice Patrol that was charged with monitoring and predicting the location of ice and icebergs as they enter the busy sea lanes of the North Atlantic Ocean.

Starting with the 1928 Marion expedition, LCDR Eward H. ''Iceberg'' Smith of the US Coast Guard conducted pioneering studies of the frigid waters between Canada and Greenland that established the generally southward discharge of ice, icebergs, and buoyant surface waters from Baffin Bay via Davis Strait into the North Atlantic. Early hydrographic observations such as those taken during the Marion (Smith, 1931) and Gothaab (Kiilerich, 1939) expeditions in 1928 mapped water temperature and salinity of Baffin Bay, Davis Strait, and the Labrador Sea. Smith (1931) used these data to estimate circulation via geostrophy to predict iceberg motions. Furthermore, Smith (1931) developed a proxy for the North Atlantic Oscillation (NAO) to predict the number of icebergs emanating from Baffin Bay to impact shipping south of Newfoundland via a regression of past observations. He discovered that years of positive NAO correspond to higher iceberg counts off Newfoundland the following year. Dunbar (1951) collated early Canadian survey data to map water properties of Baffin and Hudson Bay, Labrador, and western Greenland.

Two main circulation features emerge from past hydrographic, modeling, and mooring studies of Baffin Bay. A cold and buoyant near-surface Baffin Island Current advects Arctic ice, waters, and properties southward towards Davis Strait (LeBlond, 1980; Fissel et al., 1982; Tang et al., 2004) and a warm and salty subsurface West Greenland Current advects Atlantic water northward towards Cape York in northern Baffin Bay (Bourke et al., 1989; Muench, 1971). A summary and synthesis of mostly Canadian mooring and hydrographic efforts in Baffin Bay from 1978 through 1989 is given by Tang et al. (2004) while Cuny et al. (2005) provides a similar synthesis for Davis Strait. The net volume flux out of Davis Strait is given as 2.6 ± 1.0 Sv by Cuny et al. (2005) who use current meter mooring records below 150-m and geostrophically estimated velocity shear above this depth. Measurements from a year-long 2004/05 deployment resulted in 2.3 ± 0.7 Sv which includes directly measured currents both in the upper 100-m of the water column and on the shelves (Curry et al., 2011). Using only hydrographic observations, Muench (1971) estimate the net transport across a section of northern Baffin Bay to vary between 1.5 and 2.7 Sv which agrees with the Davis Strait estimate. Ingram et al. (2002) reviews earlier work in northern Baffin Bay in relation to the North Water polynya (Dumont et al., 2009) and references Addison (1987) who distinguishes Baffin Island Current volume flux contributions to consist of 0.3 Sv from Nares Strait, 0.3 Sv from Jones Sound, 1.1 Sv from Lancaster Sound, and 0.5 Sv from a recirculating West Greenland Current to give a total southward transport of 2.3 Sv. These values represent snapshots based on the generally untested assumptions that the flows at northern passages are both baroclinic and geostrophic. Rudels (2011) fully exploits these assumptions to derive volume and freshwater flux estimates for the entire region to the west of Greenland as well as sensitivities to additional freshwater inputs from Greenland's ice sheet.

These earlier measurements provide first descriptions of the larger basin-wide circulation features and ice drift climatology, however, they do not always resolve dynamically relevant vertical and horizontal scales of motions associated with both steeply sloping topography and baroclinic eddies. Hence it is unclear that geostrophically estimated volume fluxes associated with the cyclonic circulation are adequately resolved at both (small) spatial and (long) temporal scales. For example, hydrographic observations from which to estimate geostrophic shear do not resolve seasonal cycles. These cycles vary substantially across Davis Strait and Baffin Bay in both amplitude and phase (Zweng and Münchow, 2006) on account of different time histories of forcing of the West Greenland and Baffin Island Currents, respectively. Furthermore, the assumption of geostrophic balance is rarely tested and can break down near topography (Rabe et al., 2012).

We here discuss and analyze ocean data from the most recent expedition of the US Coast Guard to northern Baffin Bay in 2003. We made direct velocity measurements along several sections using a vessel-mounted acoustic Doppler current profiler (ADCP). These data allow evaluation of geostrophically estimated currents and, more importantly, they demonstrate mesoscale spatial variability. Enhanced delivery of fresher and colder waters from the Arctic along the shelves and slopes of Baffin Island and Labrador contributes to vertical stratification as far south as the Gulf of Maine and the Mid-Atlantic Bight where interannual ecosystem variability appears to correlate with upstream conditions (Greene et al., 2008). While our present study cannot address seasonal cycles of the salinity, temperature, and density for lack of sufficient data, we do test the assumption of geostrophy and investigate the spatial scales of velocity, salinity, and density fields in northern Baffin Bay. Our synoptic observations from the 2003 summer surveys reveal that the cyclonic circulation exhibits substantial spatial variability in the form of eddies generated via instabilities near sloping topography.

Study area and data

Baffin Bay is a semi-enclosed, seasonally ice-covered basin between northern Canada and Greenland. It is linked to the Atlantic Ocean across a 640 m deep sill in Davis Strait and to the Arctic Ocean via Lancaster Sound, Jones Sound, and Nares Strait with sill depths of about 125, 190, and 220 m, respectively (Melling et al., 2008). Fig. 1 shows locations. These channels and straits vary in minimal width from 320 km (Davis Strait), 65 km (Lancaster Sound), to 25 km (Nares Strait). They thus are generally wider than the local internal deformation radius that is about 10 km (Münchow et al., 2006). Hence even the narrowest channel can accommodate opposing baroclinic flows on each side (LeBlond, 1980). Baffin Bay contains wide and gently sloping shelf areas off Greenland in the east and narrower, more steeply sloping shelves off Baffin Island in the west. All shelves are disrupted by deep troughs and canyons that connect the continental slope and basin to the ice caps via fjords in mountainous terrain.

We primarily use data from the 2003 expedition of the USCGC Healy to northern Baffin Bay and Nares Strait. This ship contains a 75 kHz phased array ADCP that provides continuous profiles of instantaneous horizontal velocity along the ship track from about 20 m below the sea surface to about 300–600 m depth. For absolute positioning we use the ships's military grade p-code differential GPS as well as an AshTech GPS that also provides accurate heading, pitch, and roll information. For details on calibration, performance, and processing, we refer to Münchow et al. (2006) and Münchow et al. (2007) where the system, data processing, and results from Nares Strait north of 78 N are discussed. We note that data are obtained in 15 m vertical bins every 2 min. These data are further averaged in space along the track into roughly 3-km horizontal bins for display as sections. Tidal currents are initially removed using predictions from the barotropic model of Padman and Erofeeva (2004) at the location and at the time of our measurements, but prove insufficient off Baffin Island on account of large vertical variations of tidal currents that are not contained in the barotropic model. Instead, we determine tidal ellipse parameters at each vertical bin independently using the method of least

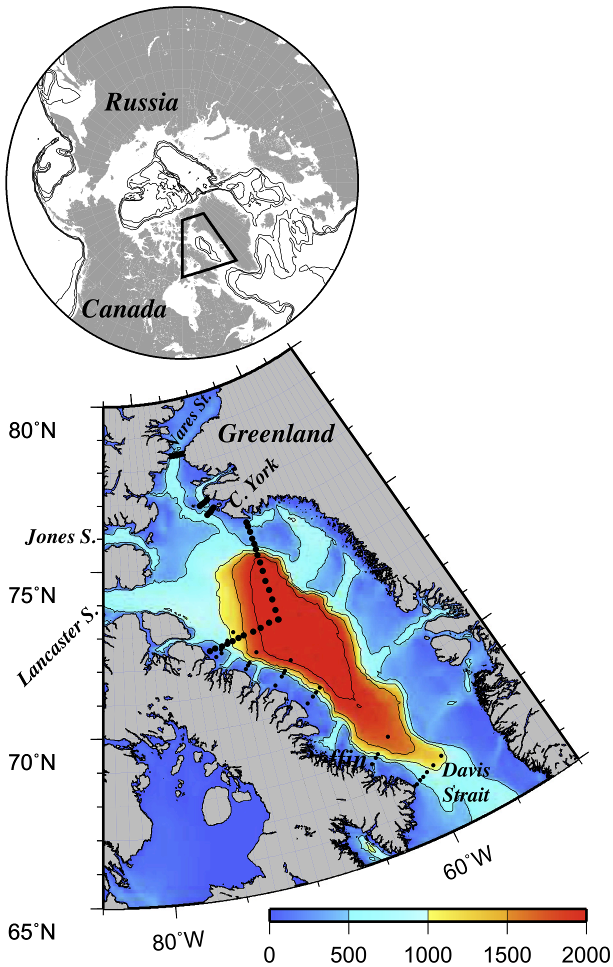

Fig. 1. Map of the study area over topography along with CTD station locations in northern Baffin Bay for 2003 (large circles) and along Baffin Island for 1979 (small circles).

squares assuming negligible horizontal variations at sections (Münchow, 2000). Detided currents are extrapolated to the surface by fitting detided subsurface velocity profiles to an Ekman layer profile with an eddy viscosity of 0.18 m^2/s giving a 50 m thick frictional layer. Münchow et al. (2007) describe, discuss, and evaluate the method and parameter choices. Assuming a random standard error of 1 cm/s for vertically averaged currents due to uncertainties in the reference velocity (from bottom-tracking or GPS), detiding, and surface extrapolation, we find 95% confidence limits for volume transport across sections of about ±0.25 Sv.

All hydrographic data were taken with a SeaBird 911Plus sensor package mounted on a 24 bottle rosette system with dual temperature and conductivity sensors that were factory calibrated 3 months prior to their use in Baffin Bay. An Autosal by Guildline was used throughout the expedition to compare bottle salinities with those derived from the SeaBird 911Plus package to ensure integrity of the CTD data collection. The 2003 data are processed identically to those described in Münchow et al. (2007) and salinities are accurate within ±0.001 (PSS78).

We collected 30 CTD casts in northern Baffin Bay between July 26 and August 3 of 2003. Thirteen stations are on a line emanating southward from Cape York, Greenland at 76 N along longitude 67 W to the center of Baffin Bay. Six stations are distributed across Smith Sound near 78 N latitude. A third section emanating from northern Baffin Island consists of 11 stations and connects almost perpendicularly to the Cape York section at the center of Baffin Bay near 73 N. Our focus is on properties above 600-m, which roughly coincides with the sill depth of Davis Strait. All temperatures are presented as potential temperatures.

In order to place our 2003 data into a larger spatial and temporal context, we also use temperature, salinity, and density data collected from 1916 through 2003 in Baffin Bay as it has been assembled by the U.S. National Oceanographic Data Center and the Canadian Bedford Institute of Oceanography (NODC/BIO data). Zweng and Münchow (2006) describe these data, their distribution in space and time, and report on a distinct warming trend in central Baffin Bay below Davis Strait sill depths and a small, but significant freshening trend of surface shelf waters from Nares Strait to Labrador.

We use the NAO index derived from normalized winter sea level pressure differences (December through March) between Lisbon, Portugal and Reykjavik, Iceland as a proxy for atmospheric variability over the northern hemisphere (Hurrell and Deser, 2009).

The West Greenland Current regime

The data from the zonal Cape York and southern Nares Strait sections portray the principal water masses with southern and northern signatures, respectively. All CTD casts exhibit pronounced subsurface temperature maxima at salinities larger than 33.9 psu

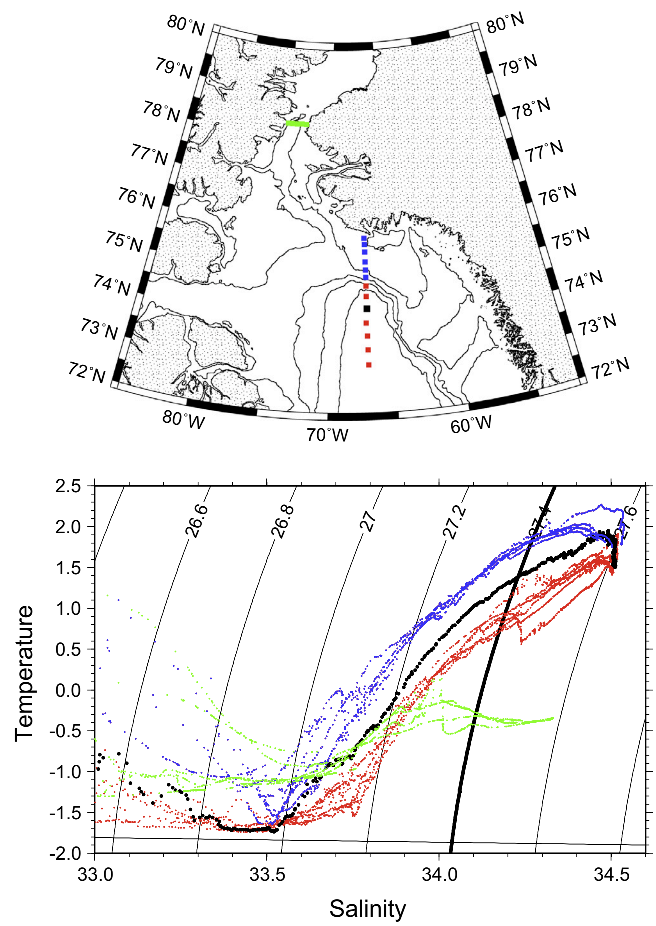

Fig. 2. CTD station locations over 500-m, 1000-m, 1500-m, and 2000-m contours of bottom depth (top) and potential temperature (°C) salinity (in psu) correlations above 600 m over contours of density (bottom). The 27.4 kg m^{-3} contour is highlighted as temperatures on this isopycnal demonstrate the influence of Arctic and Atlantic waters. Colors represent locations and properties of different physical domains (see text for details). (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

(Fig. 2) that are indicative of waters from the North Atlantic Ocean. Following Bacle et al. (2002), we distinguish between such water entering our study area from the north via Nares Strait, which has subsurface (salinity <33.5 psu) temperatures no higher than −0.4 °C and water entering our study area from the south, wherein the subsurface maximum temperature is +2.0 °C (Fig. 2). The distinction becomes particularly clear for temperatures along isopycnals in the $\sigma_t = (27.2, 27.6)$ kg m^{-3} range.

The southern waters with 2.0 °C near 34.5 psu are often associated with the West Greenland Current. However, we find these waters in at least 2 flavors with a slightly fresher (and warmer) branch located on the continental shelf inshore of the 500 m isobath and a saltier (and cooler) branch seaward of this isobath. The exception is a single cast of intermediate temperature and salinity that represents an anomaly seaward of the 2000 m isobath. Both the spatial distribution of salinity and temperature as well as underway ADCP velocity along this section suggest that this is a anti-cyclonic eddy of West Greenland shelf waters in deep Baffin Bay. Such eddies have not previously been reported in Baffin Bay.

Fig. 3 shows the density, salinity, and temperature along a north–south line that is oriented perpendicular to bathymetric contours. The shelf off Cape York slopes steeply from 50 m to 400 m within 30 km off the coast, flattens for about 40 km to plunge below 2000 m about 100 km from the coast. The salinity of the surface water is lowered by ice melt-water, warmed by insolation, and well-mixed to 20 m. Underlying waters are cooler than −1.5 °C or within about 0.3 °C of the freezing point. Bourke et al. (1989) refer to this water as the Baffin Bay Arctic Water consisting of a mixture of waters impacted by the annual summer melting and winter freezing cycle, as well as local runoff from Greenland. Below this layer which extends to about 200 m depth, we find water of about 1.2 °C at salinities of about 34.4 psu. Bourke et al. (1989) called this Atlantic Intermediate Water. Within these

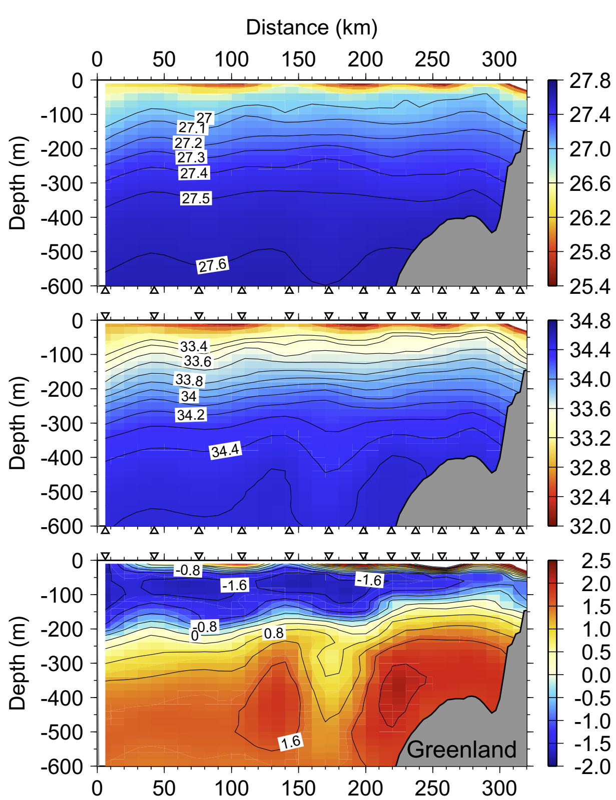

Fig. 3. Section off Cape York, Greenland, July 30/31, 2003 for density anomaly σ_t (top panel), salinity (middle panel), and potential temperature (bottom panel). Station locations are indicated by triangles.

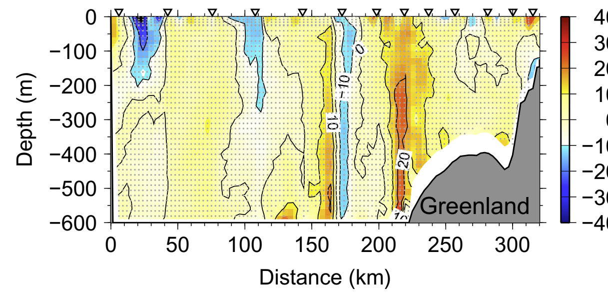

Fig. 4. Velocity section off Cape York, Greenland, July 30/31, 2003 from ship-based ADCP surveys. Large inverted triangles indicate CTD station locations to ease comparison with Fig. 3. Small symbols indicate locations of velocity measurements.

waters, however, we find two distinct cores with temperatures exceeding 2 °C and 1.6 °C shoreward and seaward of the 600 m isobath, respectively. Between these cores we find cooler and fresher waters with properties between those seaward and landward of the 600 m isobath (see Fig. 2). Velocity observations discussed below will reveal this to be an anti-cyclonic eddy. Note that the isopycnals are largely flat near the 350 m depth where this feature is most pronounced, i.e., the large lateral temperature and salinity gradients compensate such that the lateral density gradient is small.

If lateral density gradients are small in a geostrophic flow, then we expect vertical gradients of horizontal velocity to be small also. Figs. 4 and 5 display snapshots of the West Greenland Current system, derived via ship-based ADCP survey, both as a section and a vertical average. This current system consists of (1) a surface intensified westward coastal current, (2) a sluggish flow on the shelf, (3) an intense, narrow westward jet over the continental slope that spills onto the shelf near the shelf break, (4) an anti-cyclonic eddy, and (5) a sluggish circulation over the deep Baffin Bay (details to follow). The net transport of this current system above 600 m depth combines to about 3.8 ± 0.27 Sv with more than 2 Sv carried by a less than 40 km wide slope current that we will refer to as the West Greenland Slope Current.

Coastal current

Although our survey of northern Baffin Bay was not designed to resolve baroclinic flows within 10 km of the coast, both the alongshore velocity (Fig. 4) and three casts within 25 km off Cape York, Greenland (Fig. 3) reveal a wedge of warm, buoyant surface waters with salinities less than 33.4 psu and density anomalies less than 27.0 kg m^{-3}. Adjacent to the coast, this buoyant wedge extends to 100 m depth but shoals within 30 km to less than 20 m depth. Relatively large westward flows (reaching 0.2 m s^{-1}) are estimated by extrapolating measured flows below 25 m depth with an Ekman layer profile (Münchow et al., 2007). Similarly large flows (0.17 m s^{-1}) are estimated from Margule's equation that assume geostrophic flow relative to negligible flow below a sloping frontal boundary, e.g., $v = i \times g/f \times \Delta\rho/\rho$ where f is the local Coriolis parameter (1.41×10^{-4} s^{-1}), g is the constant of gravity (9.81 m s^{-2}), ρ is the density of the dynamically active upper layer (1026 kg m^{-3}), $\Delta\rho$ is the density difference across the density interface (1 kg m^{-3}) which has a slope of i (80 m over 30 km). Geostrophic coastal currents driven by local buoyancy fluxes are common at both mid-latitudes (Münchow and Garvine, 1993; Pimenta et al., 2008) and off Greenland (Bacon et al., 2002; Sutherland and Pickart, 2008). While the impact of such coastal currents on basin scale volume flux may be small, about 0.5 Sv here, their potential contribution to freshwater flux is larger as

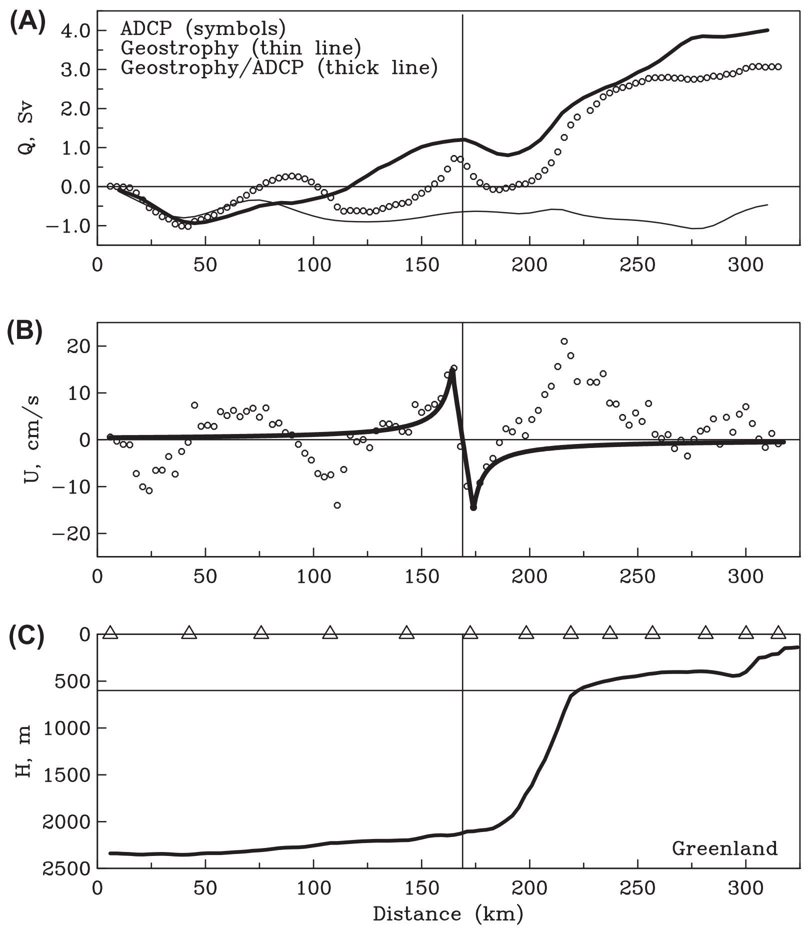

Fig. 5. Cape York section, Greenland, July 30/31, 2003: (a) cummulative volume flux over top 600 m from ADCP (symbols) and thermal wind relative to zero flow at bottom or 600-m (thin line) or relative to bottom ADCP (thick line); (b) vertically averaged along-shore ADCP velocity component (symbols) with Rankine vortex profile (line), and (c) bottom topography. The vertical line in each panel indicate the center of the Rankine vortex, symbol indicate CTD station locations, e.g., Fig. 3.

the swift surface flow carries low salinity waters far from their origins as coastally trapped flows (Sutherland et al., 2009).

Shelf flow

The flow seaward of the coastal current varies little with depth and is always less than 0.1 m s^{-1}. The depth-averaged flow is always westward and reaches a local minimum about midway across the shelf at km-275 (Fig. 5). The integrated volume flux from the shelf to this location carries about 0.5 Sv. The waters are somewhat warmer than waters on the same isopycnal over the deep basin offshore. The shelf break jet that we discuss next, spills onto the shelf near the bottom. Largest subtidal velocities exceeding 0.1 m s^{-1} occur near the 400 m deep bottom close to the shelf break.

West Greenland Slope Current

We find a pronounced westward flow over the continental slope where the water depth plunges from 600 m at the shelf break to 2000 m within 30 km. The largest vertically-averaged velocity occurs at the 600 m isobath reaching 0.2 m s^{-1} (Fig. 5). This flow is about 40 km wide at the surface, but it becomes more intense below 200 m depth where it exceeds 0.2 m s^{-1} (Fig. 4). The shoreward edge of this velocity core coincides with the subsurface temperature maximum at 350 m depth near the 500 m isobath (Fig. 3). In the seaward direction the current extends to the 1500 m isobath. The current is thus contained entirely over the slope and does not extend to the foot of the continental slope where the bottom changes its slope from 0.05 to 0.002. We will refer to this current as the West Greenland Slope Current to distinguish it from the weaker westward flows on the shelf. The slope current carries a volume of about 2.0 Sv westward over the top 600 m (Fig. 5).

Horizontal density gradients associated with the West Greenland Slope Current are small, because higher temperature and higher salinity relative to ambient waters compensate each other with regard to density. Thus while conventional hydrographic measurements may trace the origin of waters off western Greenland, they cannot reveal the geostrophic circulation, because the West Greenland Slope Current contains a large barotropic component.

The relative vorticity ξ of a geostrophic flow is much smaller than the planetary vorticity f (Gill, 1982). We estimate ξ for the vertically averaged flow (Fig. 5) as $\xi \approx \Delta u/\Delta y \approx 0.1f$ where $f = 1.4 \times 10^{-4}$ s^{-1} and $\Delta u = 0.15$ m s^{-1} is the along-slope velocity difference over an across-slope distance $\Delta y = 12$ km. Since nonlinear inertial effects are scaled by $\xi/f \approx 0.1$, we discern that they are small relative to Coriolis effects and that the barotropic flow is in geostrophic balance to first order during our expedition.

Anti-cyclonic eddy

Seaward of the West Greenland Slope Current near km-170 both a single CTD cast and the velocity measurements approaching and leaving this location from south to north reveal anomalous water properties and ocean currents. Ocean currents change from 0.15 m s^{-1} westward to 0.15 m s^{-1} eastward over a distance of less than 20 km just seaward of the continental slope (Figs. 3–5). Water between 200 m and 500 m within this feature is cooler and fresher than adjacent waters. This signature extends to about 800 m depth (not shown). The locally depressed isopycnals suggest a clock-wise geostrophic circulation relative to no flow at greater depths which is consistent with the observed flow shown in Fig. 4. We thus interpret our observations to represent an anti-cyclonic eddy.

The almost axisymmetric velocity distribution with a linear shear of 0.3 m s^{-1} over 10 km suggests an eddy core with radius $r_m \approx 5$ km in solid-body rotation that can be modeled as a Rankine vortex (Timmermans et al., 2008). The Rankine vortex emerges as a particular simple solution in steady fluids where nonlinear advective and pressure gradient forces contribute to the dynamics. For a Rankine vortex the azimuthal velocity increases from zero at the center of the vortex to a maximum V_g at r_m (0.15 m s^{-1}) and then decreases with the inverse distance from the eddy center, e.g., $v(r) = V_g r/r_m$ for $r \leqslant r_m$ and $v(r) = V_g r_m/r$ for $r < r_m$. Fig. 5 shows the analytical solution demonstrating that it fits the observed velocity distribution well both for the 10-km wide eddy core and at least another 10 km to either side. The Rankine vortex has a uniform potential vorticity distribution $\Pi = 2V_g/r_m$ for $r \leqslant r_m$ and zero potential vorticity for $r > r_m$. An estimate of the Rossby number $Ro = \Pi/f \approx 0.4$ indicates a nonlinear flow. It recirculates a volume flux of at least 0.5 Sv within its core of uniform potentential vorticity.

The Baffin Island Current regime

Water masses

Fig. 6 shows potential temperature salinity relationships above 600 m as well as the measurement locations over bottom topography. Off Baffin Island, the temperature of water between the surface mixed layer and the 33.7 isohaline is almost constant at −1.6 °C. The low temperature of this part of the halocline reflects the impact of wintertime freezing within the polynyas of northern Baffin Bay – in Smith, Jones and Lancaster Sounds. As salinities increase towards 34.5 psu, temperature increases towards a maximum of +1.0 °C near the 27.6 σ_θ density surface. Again, these waters are distinct from Nares Strait waters which are almost 2 °C cooler. Nevertheless, the warm subsurface waters off Baffin Island are always cooler than those found off western Greenland at similar salinities. We thus identify the West Greenland Current System as the main source of the subsurface waters off Baffin Island which is consistent with the cyclonic circulation in northern Baffin Bay. Waters from Nares Strait are a minor source that modify fresher waters near the surface towards warmer temperatures while saltier waters at depth are modified towards cooler temperatures on density surfaces.

Fig. 7 presents the same data along a section that extends from the coast of Baffin Island near 72 N latitude towards the center of Baffin Bay (Fig. 1). A cold and relatively fresh layer above 300 m depth separates a seasonally warmed 20 m thin surface mixed layer from the warm and salty West Greenland Current waters. At salinities below 33.8 psu, the coldest waters of −1.6 °C are remnants of winter waters. Comparing the properties of these waters along isopycnals of Figs. 2 and 6, we find the northern waters along isopycnal surfaces such as the 27.0 σ_θ warmer (and thus saltier) by almost 0.5 °C at a salinity near 33.6. The waters at these salinities in Smith Sound likely contain a larger fraction of Pacific waters that enter the Arctic Ocean via Bering Strait (Woodgate and Aagaard, 2005; Münchow et al., 2007).

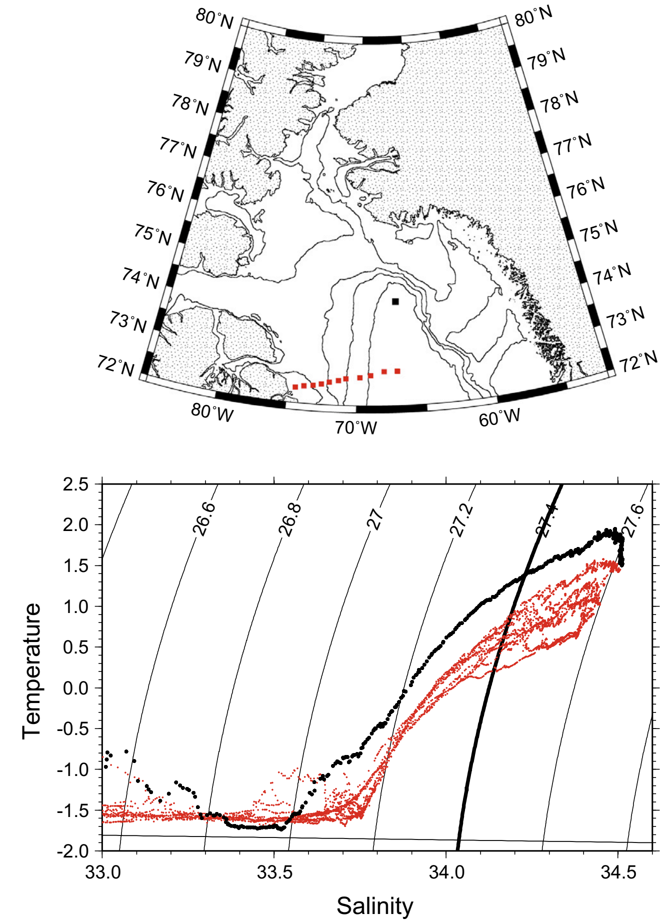

Fig. 6. CTD station locations over contours of bottom depth from 500 m to 2000 m (top) and potential temperature salinity correlations above 600 m over contours of density (bottom). The 27.4 kg m^{-3} contour is highlighted as temperatures on this isopycnal demonstrate the influence of Arctic and Atlantic waters (see Fig. 12). Bold black symbols indicate a cast that is also shown in Fig. 2 for comparison.

Velocity

The most dramatic feature in Fig. 7, however, are undulating isopycnal excursions that exceed 50 m over 40 km. If the associated baroclinic pressure gradients are balanced by the Coriolis force, then we can estimate the geostrophic velocity field that these isopycnals imply. In Fig. 8 we show these geostrophic (thermal wind) velocities that we reference at 600 m depth to observed ADCP velocities. The reference velocities are always smaller than 0.1 m s^{-1} while the geostrophic surface velocities exceed 0.3 m s^{-1} in both northward and southward directions as isopycnals slope upward and downwards towards the east, respectively. Opposing flows are particular strong about 220 km from the coast where a southward jet exceeds 0.4 m s^{-1} adjacent to a northward flow of about 0.2ms^{-1}. We find weak geostrophic flows over both the narrow shelf and steeply sloping continental shelfbreak within 50 km off Baffin Island.

The flow calculated via geostrophy compares favorably to concurrent direct observations of velocity derived via ADCP, shown in Figs. 9 and 10. These direct observations consist of both geostrophic and ageostrophic velocity components. The directly observed flow clearly reveals the strong vertical component of shear above 300 m depth. The largest lateral velocity gradient occurs near the surface between 200 km and 220 km from the coast, where the value changes from about −45 cm/s to +25 cm/s in both realizations. Clearly the lateral shear is closely linked to

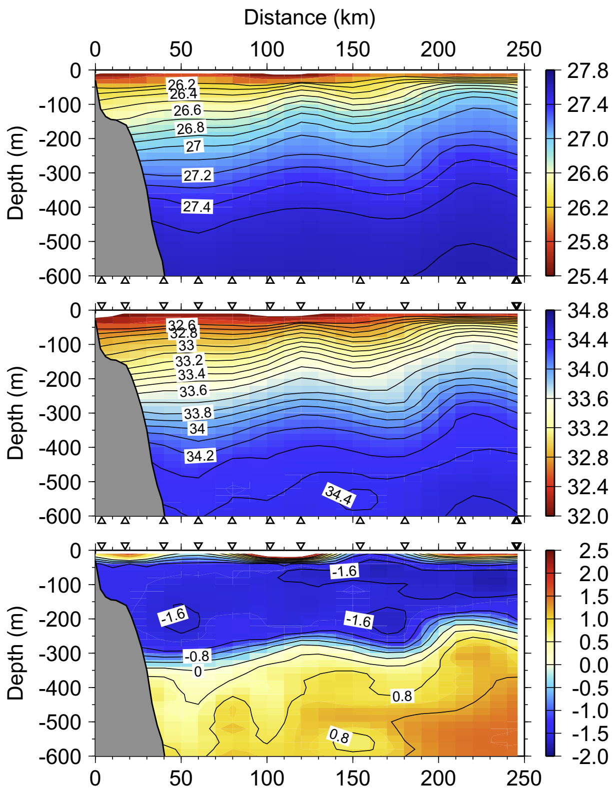

Fig. 7. Section off Baffin Island, Canada, July 26/27, 2003 for density anomaly σ_t (top panel), salinity (middle panel), and potential temperature (bottom panel). Station locations are indicated by triangles.

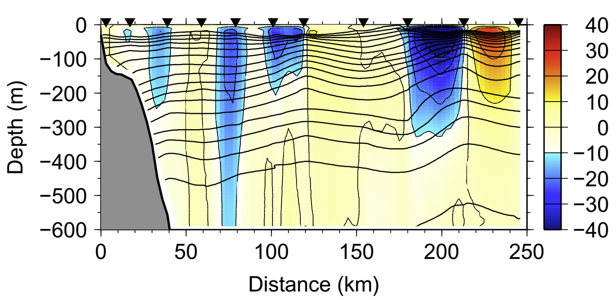

Fig. 8. Velocity section off Baffin Island as predicted from the thermal wind relation, July 26/27, 2003 relative to measured ADCP flow at the bottom or 600 m depth. Contours are those of density (Fig. 7) from CTD casts whose location is shown as triangles.

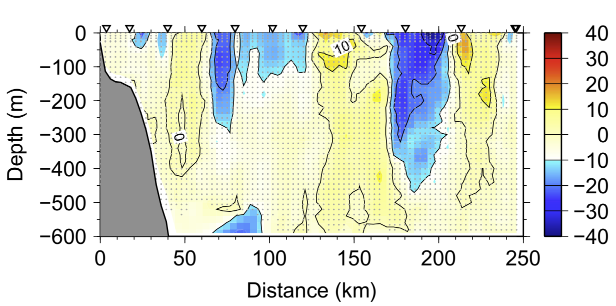

Fig. 9. Velocity section off Baffin Island, Canada, July 26/27, 2003 as measured by vessel-mounted ADCP. Triangles indicate CTD locations, small symbols indicate ADCP measurement locations averaged into 3-km wide and 15 m deep bins.

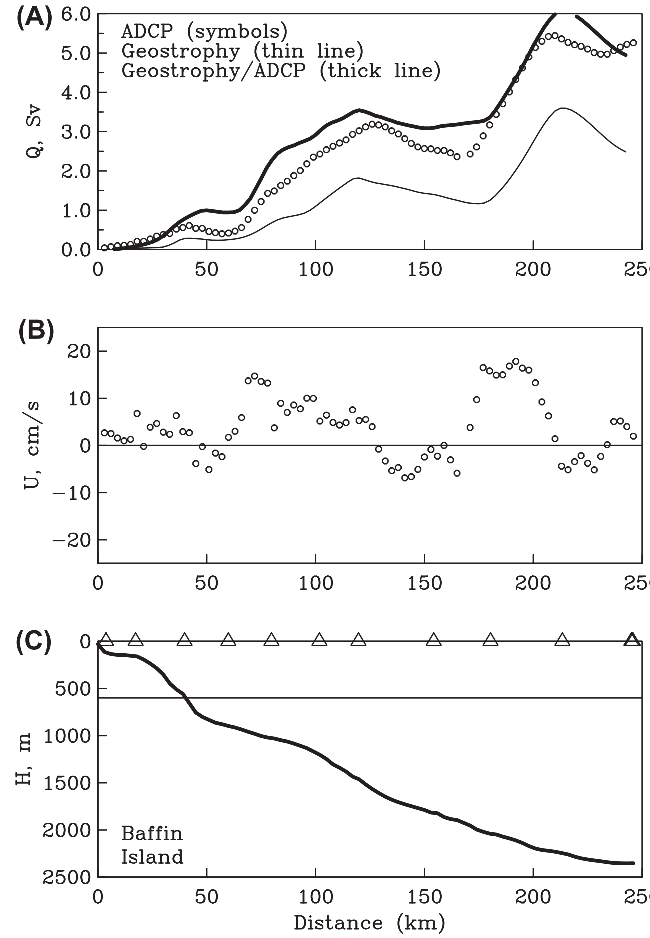

Fig. 10. Northern Baffin Island section, July 26/27, 2003: (a) cummulative volume flux over top 600 m from ADCP (symbols) and thermal wind relative to zero flow at bottom or 600-m (thin line) or relative to bottom ADCP (thick line); (b) vertically averaged along-shore ADCP velocity component (symbols), and (c) bottom topography. Symbol indicate CTD station locations, e.g., Fig. 7.

the undulations of density surfaces at this location (Fig. 7). Scaling this velocity difference of $\delta U \approx 0.7\ \mathrm{m\,s^{-1}}$ over $L = 20$ km by the Coriolis parameter $f = 1.38\ \mathrm{s^{-1}}$, we find a first rough estimate of the Rossby number $R = \delta U/(fL) \approx 0.25$ which indicates that non-linear inertial effects may contribute to the dynamics and stability characteristics of the observed currents.

Integrating the vertically averaged alongshore velocity along the section, we show with Fig. 10 how the volume transport perpendicular to our section reaches 5.1 ± 0.24 Sv. Over the shelf and shelfbreak within 50 km of Baffin Island, volume transports are below 0.2 Sv. Seaward of the 800 m isobath, two much larger current structures emerge. The first represents a broad and sluggish flow less that 10 cm/s from about 50 km to 130 km offshore. This flow carries about 1.6 Sv, but a 30 km wide counter-current or eddy structure reduces the net transport to less than 0.8 Sv over the top 500 m of the water column. Most of the volume transport across the Baffin Island section is contained within a southward jet about 160 km from shore near the 2000 m isobath. It alone carries almost 3.5 Sv over its 60 km width from 160 km to 220 km from the coast. The same flow emerges via geostrophy from the hydrographic observations (Figs. 8 and 7). The along-shore velocity within this 60 km wide jet is vertically sheared, so that lateral shear vanishes at 600 m depth, that is, the velocity observed by ADCP at 600 m contributes little to the geostrophic currents that contain both vertical and lateral shear above 300 m. This gives confidence that directly observed ADCP surface currents are consistent with independently estimated geostrophic currents.

Baffin Island Current hydrography 1979

We here exploit the finding that the thermal wind relation holds well off Baffin Island by applying it to hydrographic data from the most comprehensive survey conducted along coastal Baffin Island. Fissel et al. (1982) introduce these 1979 data, but focus on the surface circulation off Bylot Island at the entrance to Lancaster Sound just to the north-west of our study area. In Fig. 11 we show 6 across-shore sections from 72°N latitude to Davis Strait in the south near 67°N latitude, see Fig. 1 for locations. Excluding the northernmost section, geostrophic volume and freshwater flux is towards Davis Strait in the south. Values vary between 1.8 and 4.0 Sv for volume and 78 and 154 mSv for freshwater. Mean values are 3.0 ± 0.9 Sv and 118 ± 35 mSv where the uncertainty is a standard deviation of the along-shore variability. The along-shore continuity is not perfect, but the generally upward sloping isopycnals from about 300 m near the shelf-break to 100 m about 100 km offshore are ubiquitous. This is the cold and fresh outflow from the Canadian Archipelago. The northernmost section with the negative (northward) flux probably does not resolve the offshore extent of the outflow and implies a weak circulation in the lee of Bylot Island. Fissel et al. (1982) report on vigorous eddy activity near the surface at this location which also coincides with our 2003 section revealing less than 0.2 Sv volume flux within 60 km of Baffin Island (Fig. 10). Note also, that 1979 was an NAO-negative year (Fig. 13) which implies weaker than average baroclinic circulation. We indeed find the geostrophic thermal wind circulation weaker in 1979 as compared to 2003.

Volume and freshwater flux

Our two 2003 sections across the shelf, slope, and basin off Greenland and Baffin Island intersect near 73 N latitude at the thalweg over the deep Baffin Bay basin (Fig. 1). These sections define a volume that is open to Nares Strait and Jones Sound in the north-east and Lancaster Sound in the north-west. Table 1 summarizes volume and freshwater flux estimates in the form of a closed budget. The net flow into our study area from Nares Strait in the summer of 2003 is about 1.0 ± 0.2 Sv in the absence of winds (Münchow et al., 2007). The West Greenland current systems adds 3.8 ± 0.27 Sv inflow from the south-east. Mooring observations suggest that Jones and Lancaster Sounds can provide an additional 0.3 Sv and 1.0 Sv, respectively. Our control volume thus conserves volume within an uncertainty of 1.0 Sv or about 20% of the outflow. Furthermore, seasonal cycles are large in Lancaster Sound with volume flux reaching 1.0 Sv in summer and dropping below 0.2 Sv in winter (Peterson et al., 2012).

The situation is similar for freshwater flux q_f which we estimate relative to a salinity $S_0 = 34.8$ psu by integrating observations of salinity S (x,z) and velocity $u(x,z)$ over the sectional area A in the (x,z) plane, e.g., $q_f = \int_A (1 - S/S_0)\, u\, dA$. Figs. 3 and 4 show S and u for the West Greenland Current System which provides $q_f = 72 \pm 20$ mSv into our volume. The Baffin Island Current System (Figs. 7 and 9) exports $q_f = 187 \pm 30$ mSv towards Davis Strait. Table 1 summarizes these results and adds estimates for Nares Strait, Lancaster Sound, and Jones Sound. Observations from the summer of 2003 give 34 ± 6 mSv for Nares Strait (Münchow et al., 2007) and 75 ± 10 mSv for Lancaster Sound (Peterson et al., 2012) which balances the freshwater flux within 3% of the outflow, but this is perhaps fortuitous as the uncertainties are an order of magnitude larger (Table 1).

These estimates compare to the net annual mean freshwater flux through Davis Strait of 116 ± 41 mSv for 2004–2005 and a mean volume flux of 2.3 ± 0.7 Sv that exit Baffin Bay (Curry et al., 2011). About 1.9 Sv of volume and 27 mSv of freshwater enter Baffin Bay from the south (Curry et al., 2011). Assuming that these Davis Strait inflows to Baffin Bay are contained within our West Greenland section, we conclude that about 1.9 Sv of volume and 27 mSv of freshwater recirculate within northern Baffin Bay. Furthermore, Curry et al. (2011) find that the shelf off West Greenland carries 0.4 Sv in volume and 15 mSv in freshwater flux into Baffin Bay from the south that relate to glacial meltwater (Azetsu-Scott et al., 2012). We thus speculate that most of the observed 2003 circulation over the shelf and slope off north-west Greenland in Baffin Bay is a recirculation of both salty Atlantic and fresher Arctic waters.

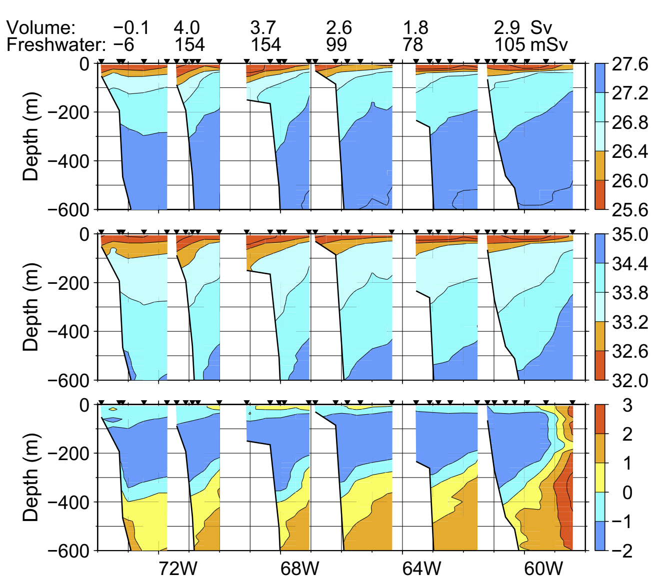

Fig. 11. Potential temperature (bottom), salinity (middle), and potential density anomaly (top) along coastal Baffin Island with sections from 72.1°N (left) to 67.1°N latitude (right). The geostrophic volume and freshwater flux estimates are relative to no flow at 600 m depth are shown also. Sections are shown against longitude with 1° representing 38 km at 70°N. Station locations are indicated by small symbols while numbers at the top are geostrophic volume and freshwater flux estimates relative to no flow at 600 m.

Table 1
Flux estimates for 2003 surveys of northern Baffin Bay and published mooring data, e.g., Peterson et al. (2012) for Lancaster and Melling (2000) for Jones Sounds. Positive (negative) sign indicates flux into (out of) a closed volume.

Section	Volume (Sv)	Freshwater (mSv)	Source
West Greenland	3.8 ± 0.3	72 ± 20	Survey
Nares Strait	1.0 ± 0.2	34 ± 6	Survey
Baffin Island	−5.1 ± 0.2	−187 ± 30	Survey
Lancaster Sound	1.0 ± 0.2	75 ± 10	Moorings
Jones Sound	0.3 ± 0.1	Unknown	Moorings
Sum	1.0 ± 1.0	−6 ± 66	

Climatological context 1916–2003

1 We start our discussion of the hydrographic climatology of northern Baffin Bay with Fig. 12 which shows the potential temperature on a constant density surface ($\sigma_t = 27.4$ kg m^{-3}) derived from the NODC/BIO data as well as our own 2003 data. Zweng and Münchow (2006) discusses the seasonal bias of these data that are generally collected in the summer in ice-free waters. The vertical location of this density surface varies from about 50-m

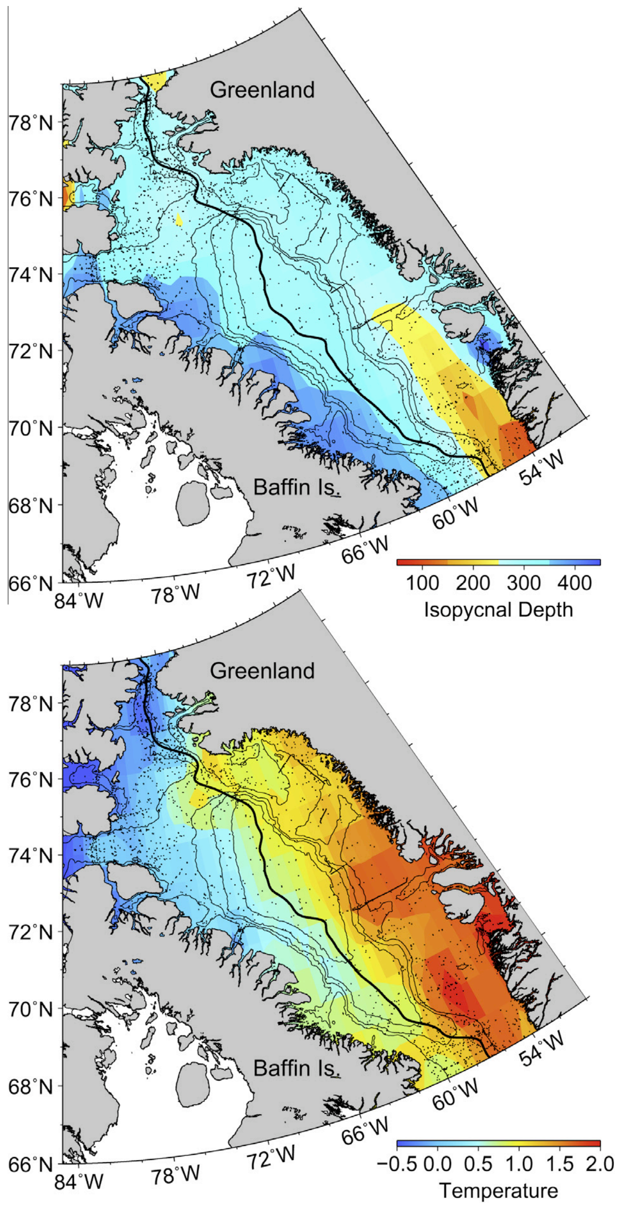

Fig. 12. Depth in meters (top) and potential temperature in °C (bottom) on the 1027.4 kg m^{-3} isopycnal shown in color from climatological data over contours of bottom depth from 500 m to 2000 m in 500 m increments. Thick line is the thalweg. Symbols indication station locations; note the absence of data from the Baffin Island shelf between 69 N and 72 N latitude. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

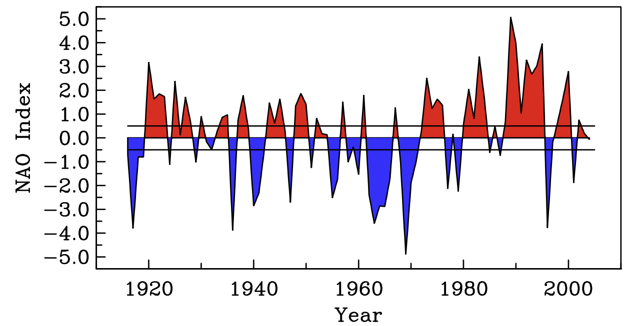

Fig. 13. Time series of the North-Atlantic Oscillation index from 1910 through 2004. Horizontal lines at ±0.5 separate larger positive and larger negative NAO years that we use in the NAO conditional averaging in Fig. 14.

in eastern Davis Strait to more than 350 m depth along Baffin Island. The distribution of potential temperature in Fig. 12 reveals the warm (and thus salty) waters off Greenland exceeding 1 °C as compared with the cold (and thus less salty) waters off Ellesmere and Devon Islands in the north-west with temperatures below 0 °C. These are the signatures of the West Greenland and Baffin Island Currents that together comprise the cyclonic circulation. The shallow occurrence of the 27.4 σ_t feature in the center of Baffin Bay near 66 W longitude and 72 N latitude is consistent with this cyclonic circulation. Note also the patchy temperature distribution north of 72 N where the colder, fresher northern waters meet the warmer, saltier southern waters. This is an area of water mass transition and transformation.

The temporal context of our detailed 2003 observations is demonstrated via the North-Atlantic Oscillation (NAO) which represents the dominant mode of atmospheric variability in the northern hemisphere (Hurrell and Deser, 2009). A more deeply depressed Icelandic Low than normal constitutes the positive NAO phase. Zweng and Münchow (2006) demonstrate that subsurface temperature fluctuations in Baffin Bay correlate significantly with the NAO reflecting Baffin Bay's connection to the climate regime of the North-Atlantic. Fig. 13 shows the NAO index from 1910 through 2004 with non-dimensional amplitudes with a range of ±5. We form conditional averages of hydrographic properties for years with NAO amplitudes larger than +0.5 and smaller than −0.5. Positive values are associated with more intense storms at higher latitudes causing drier and colder winter conditions over Greenland and northern Canada while negative values are associated with both weaker and more southerly storm tracks (Hurrell and Deser, 2009). The NAO value for 2003 is +0.2, close to the average or normal conditions.

It is instructive to examine how the implied geostrophic circulation in Baffin Bay differs between positive and negative NAO states. We average the climatological data into a section across Baffin Bay created by binning observations according to water depth. By doing so, we assume that hydrographic properties are uniform along isobaths from 67 N to 76 N latitude. Depths on the Greenland side have been separately binned from those on the Canadian side separated by the thalweg – the locus of points at having the greatest depth for each latitude. The number of casts entering this average varies from a low of 15 in the deep center of Baffin Bay to a high of 236 over the slope off West Greenland.

Fig. 14 shows the results of this conditional averaging in time (separately for high and low NAO years) and space (separately east and west of the thalweg by bathymetry) for a vertical bin that extends from 200 m to 400 m below the surface. This layer likely contains elements of both the Baffin Island and the West Greenland Currents in Baffin Bay. During the positive NAO phase this layer is both fresher and cooler off Canada in the west and saltier and warmer off Greenland in the east relative to the NAO negative phase. This indicates a stronger (weaker) than normal cyclonic circulation during the NAO positive (negative) phases consistent with the earlier findings of Smith (1931) that more (less) icebergs occur off Newfoundland during years with a positive (negative) NAO anomaly, because it implies enhanced southward flux of cold, fresh Arctic waters along Canada and enhanced northward flux of warmer, saltier Atlantic waters along West Greenland.

The depth of isopycnals across Baffin Bay increase from west to east in all years. This indicates a net baroclinic outflow from the Arctic into the North-Atlantic Ocean relative to zero flow below. Nevertheless, there are smaller baroclinic features confined to the continental slope (1500–2000 m isobaths) that demarcate the shelf from the deep basin off Baffin Island. Specifically during negative NAO years, the density anomaly σ_t varies by only

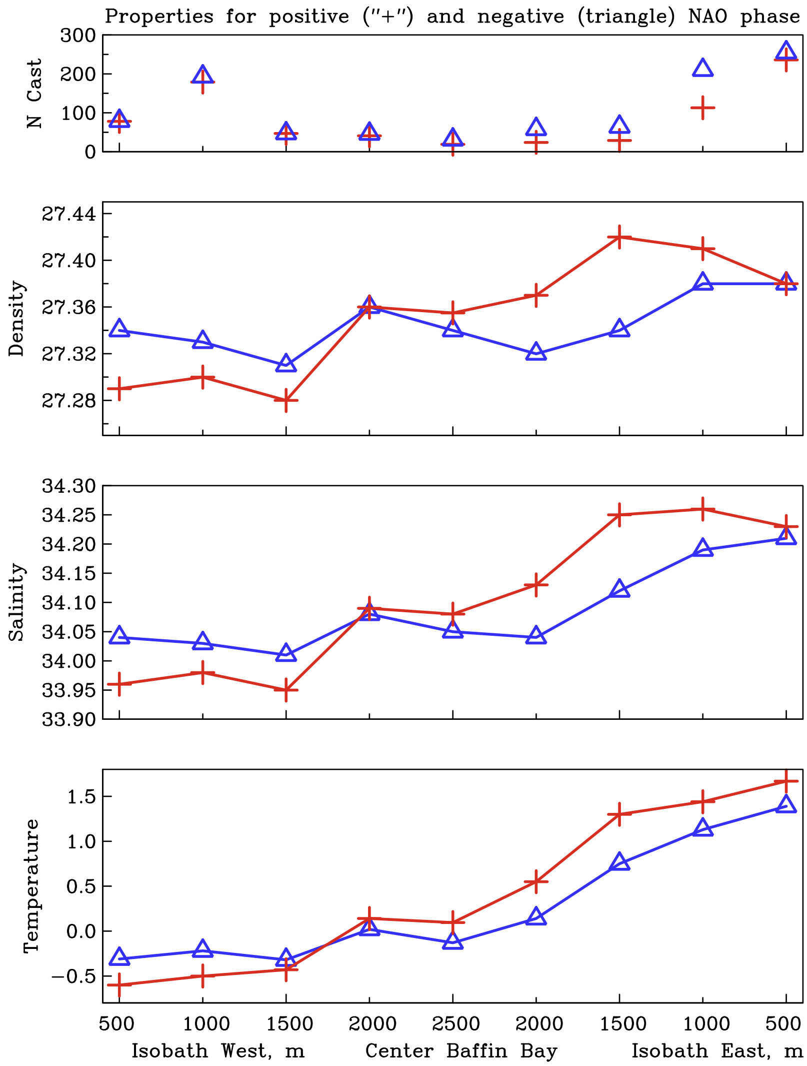

Fig. 14. Water Properties 200–400 m level across Baffin Bay averaged conditionally for positive and negative NAO phase by bottom depths west (off Canada) and east (off Greenland) of the thalweg. A positive NAO is associated with fresher and cooler waters off Canada and saltier and warmer waters off Greenland relative to a negative NAO. Top panel shows the number of vertical casts entering the average.

0.03 kg m^{-3} (27.32–27.35 kg m^{-3}), but during positive NAO years σ_t varies across the shelf break by more than 0.09 kg m^{-3} (27.28–27.35 kg m^{-3}). We speculate that the waters above the 1500–2000 m isobath at 200–400 m depth are part of the climatological Baffin Island Current.

Density at fixed depth increases between the 1500-m isobath and the coast of Baffin Island on the Canadian side. The size of the increase is greater within the domain centered at 200-m depth than that centered at 300 m (Fig. 14 and 15). If the lateral density gradients are geostrophically balanced relative to a deeper level without flow, then this gradient implies a possible northward counter-current in shallow water that has also been noted by Tang et al. (2004) and Curry et al. (2011). The same geostrophic shear could also be facilitated by a subsurface flow that has a southward maximum at depth. A velocity section across Nares Strait shows such enhanced subsurface jet in geostrophic thermal balance (Münchow et al., 2007).

Identical results emerge for a smaller vertical interval closer to the surface: Fig. 15 shows conditionally averages properties between 150 m and 250 m below the surface. This layer emphasizes the shallower Baffin Island Current over the deeper West Greenland Slope Current. The northward countercurrent off Baffin Island appears stronger in the 150–250 m as compared to the 200–400 m averages for both positive and negative NAO states. Despite these details, the conditional averaging by NAO along isobath reveals robust features of the mass and heat distribution within Baffin Bay that do not depend on the details of the vertical averaging or NAO cut-off.

Discussion

Analyses of hydrographic data in Baffin Bay during the 1916–2003 period indicate that the NAO index modulates the baroclinic pressure distribution inside Baffin Bay (Figs. 14 and 15), creating a stronger geostrophic cyclonic circulation during the NAO-positive years such as 1919, 1973, 1984, and 1990 than it is during the NAO-negative years such as 1916, 1936, 1969, and 1996 (Fig. 13). These finding also demonstrates why differencing hydrographic properties from the early 1960s from those of the 1990s reveals large signals, e.g., Dickson et al. (2003) and Dickson et al. (2002). During this period the NAO goes from an extreme negative to an extreme positive state with attendant large variations in circulation. Only long-term records covering a full cycle of such oscillations will provide the data to distinguish such climate oscillations from the more steady man-made globally warming signals observed in the atmosphere (Ring et al., 2012). Zweng

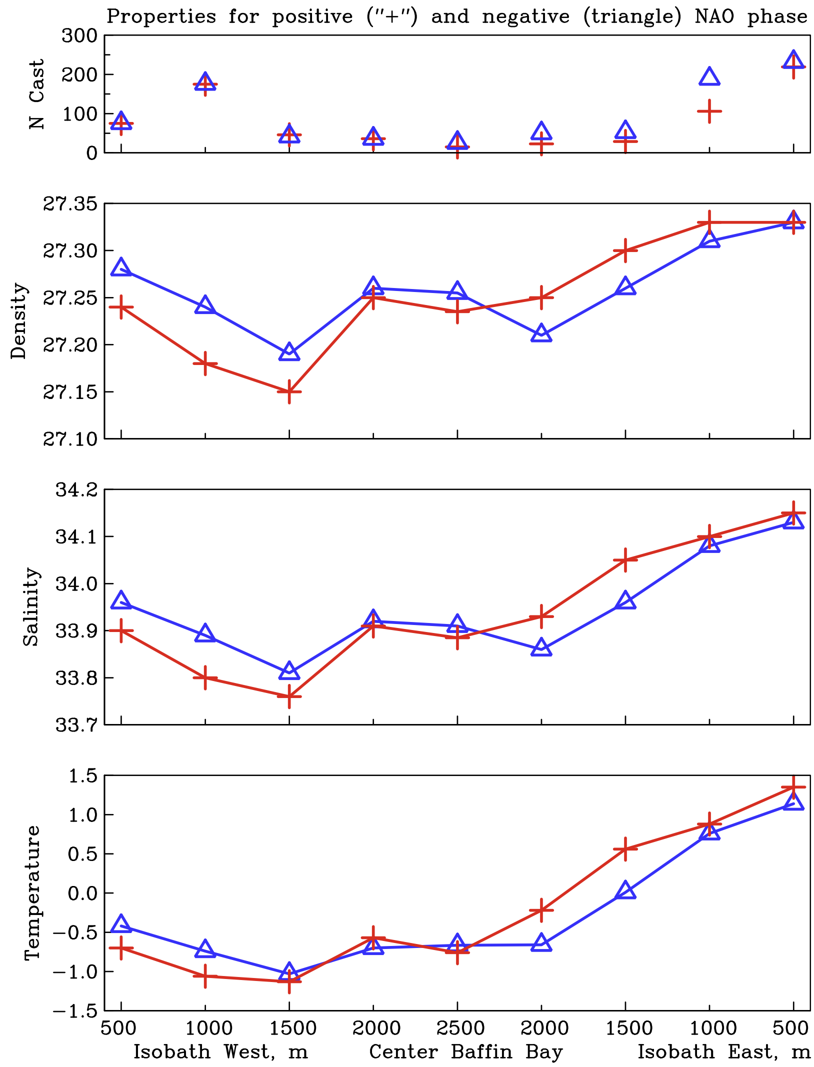

Fig. 15. As Fig. 14, but for water Properties 150–250 m level.

and Münchow (2006) demonstrate that these warming signals reach into Baffin Bay at 600 m depth. Inside Baffin Bay they promote subsurface melting of tidewater glaciers along West Greenland (Holland et al., 2008).

Surface waters off Baffin Island are fresher and colder during NAO-positive years while those off Greenland are saltier and warmer. This finding is consistent with a more energetic circulation in Baffin Bay. Zweng and Münchow (2006) demonstrate that the warming of subsurface waters in Baffin Bay correlates significantly with the NAO which emphasizes the connection of the regional oceanography with remote atmospheric forcing over the North-Atlantic at interannual time scales. We consider our 2003 observations to represent a climatological mean rather than an extreme state, because the NAO index was close to zero in both 2002 and 2003.

During our summer 2003 survey of northern Baffin Bay we find a delicate spatial arrangements of water masses and ocean currents within about 600 m of the surface. The waters off western Greenland are strongly impacted by relatively warm and salty waters originating from the North-Atlantic entering via Davis Strait (Cuny et al., 2005; Tang et al., 2004) while those off Baffin Island are strongly impacted by relatively cold and fresh waters originating from the Arctic Ocean (Münchow and Melling, 2008; Prinsenberg and Hamilton, 2005; Peterson et al., 2012). Waters are strongly stratified in the vertical both off West Greenland in the east and off Baffin Island in the west. In contrast, lateral density gradients over the slope off Greenland are small relative to those found off Baffin Island and imply a weak baroclinic circulation in geostrophic (thermal wind) balance.

Weak baroclinic circulation does not imply weak flows, however, because the total flow also contains a barotropic component. Specifically, we find a 20 km wide, largely barotropic flow centered over the 600 m isobath west of Greenland. This flow, which we call the West Greenland Slope Current, carries about 2 Sv ($10\ m^6\ s^{-1}$) towards the north-west during our survey. It contains the warmest waters found in northern Baffin Bay with potential temperatures exceeding 2 °C at 400 m below the surface in 600 m deep water. This slope current is distinct from the flows both seaward over the deep basin and landward over the continental shelf.

Flows off Baffin Island are largely in baroclinic geostrophic balance as evidenced by directly measured ocean currents (Figs. 8 and 9). Both velocity observations and geostrophic diagnostics reveal the main circulation features over the sloping topography off Baffin Island, namely (1) a slow broad southward flow within about 150 km of the coast and (2) an intense, surface intensified cyclonic feature with southward velocities exceeding $0.4\ m\ s^{-1}$ within 15 km of a weaker, but northward surface velocity core reaching $0.2\ m\ s^{-1}$. The broad sluggish inshore flow carries about 1 Sv while the 50 km wide offshore feature carries another 4.4 Sv of volume southward. The latter is centered near the 2000 m isobath about 180 km from the coast of Baffin Island. It coincides with isopycnals that slope by about 100 m over 20 km almost uniformly from 450 m to 50 m depth. Comparing the direct velocity observations with those estimated from hydrography via the geostrophic thermal wind relation, we conclude that the vertical shear measured by the vessel-mounted ADCP off Baffin Island is largely geostrophic and baroclinic.

Seaward of the West Greenland Slope Current we identify anomalous waters that extend from the surface to about 800 m depth within an anti-cyclonic circulation feature that is well modeled as a Rankine vortex with a diameter of about 10 km (Fig. 5). The small 10-km scale of this eddy corresponds to the internal (baroclinic) Rossby radius of deformation which is the dominant spatial scale for a stratified fluid in geostrophic balance (Gill, 1982), however, it also corresponds to the width of the continental slope and the width of the barotropic West Greenland Slope Current. The anti-cyclonic eddy extends across the entire halocline with a thickness exceeding 400 m near continental slope where most of the kinetic energy is contained within the largely barotropic West Greenland Slope Current. It recirculates a volume flux of about 0.8 ± 0.2 Sv.

A single CTD cast from the core of the vortex distinctly separates temperature-salinity correlation curves from a grouping representative of West Greenland shelf waters off Cape York and a grouping representative of Baffin Bay basin waters with salinities above 33.5 psu (Figs. 2 and 6). Specifically, above salinities of 33.5 psu, the entire salinity-temperature correlation falls between the shelf and basin cluster of CTD profiles. Water temperatures within the vortex are about 0.2 °C warmer than slope and basin waters and 0.4 °C cooler than shelf waters on the same isopycnals. The northward flow of Atlantic waters via Davis Strait is the source of the waters over the slope (Tang et al., 2004). This inflow is seasonally modulated in both its velocity magnitude (Cuny et al., 2005), its subsurface temperature maximum (Zweng and Münchow, 2006), and its heat transport into Baffin Bay. Davis Strait is about 800 km to the south. Assuming a swift flow over the slope of 0.2 m s^{-1}, a water parcel would arrive at our study region about 45 days later. We thus speculate that the eddy represents hydrographic conditions of the West Greenland Slope Current at least 2 months prior.

A definite explanation for the origin of this anticyclonic circulation feature requires more comprehensive observations and numerical modeling. Katsman et al. (2004) and Spall et al. (2008) discuss eddy dynamics related to slope and boundary currents in numerical models to explain observations off south-west Greenland and north-west Alaska, respectively. Both these studies identify baroclinic instability as the main eddy formation process, however, our limited observations indicate that most of the kinetic energy over the slope off West Greenland is barotropic. The scale of the vortex is of the same order of magnitude as both the width of the slope and the internal deformation radius. Hence we are presently unsure which instability process generated the anti-cyclonic eddy seaward of the West Greenland Slope Current. We do note, however, that the barotropic circulation over the continental slope off West Greenland resembles the West Spitsbergen Current in the Greenland Sea (Walczowski et al., 2005) which is postulated to become barotropically unstable (Teigen et al., 2010).

It is unclear, however, how a barotropic circulation over the steeply sloping shelf break off West Greenland transforms into a baroclinic circulation largely detached from the bottom over much deeper water off Baffin Island. Within about 150 km off Baffin Island the circulation is sluggish during our observations in July 2003, however, we find largest currents as a single surface intensified, 50-km wide baroclinic jet about 180 km from the coast. Vertical currents shears predicted from geostrophy agree well with vertical shears measured from vessel-mounted ADCP surveys. Counter-currents or eddies appear in the velocity section (Fig. 9) that correspond to sloping and undulating isopycnals in apparent geostrophic balance. Lateral current shears suggest Rossby numbers of up to 0.25 or a weakly nonlinear flow off Baffin Island.

Conclusions

Direct velocity observations from vessel-mounted ADCP reveal that the circulation off both West Greenland and Baffin Island contains multiple velocity cores, eddies, and counter-currents at scales that correspond to both the internal Rossby radius of deformation and topographic slopes. Most of these flow features correlate well with distinct water mass properties that suggest geostrophic dynamics. More intense flows over the slope off West Greenland implicate nonlinear inertial forces as Rossby numbers reach 0.4 within an anti-cyclonic eddy that is well represented as a Rankine vortex.

The outflow of cold and fresh Arctic waters from Nares Strait and Lancaster Sound transforms into the Baffin Island Current with substantial contributions from the warmer and saltier Atlantic waters of the West Greenland Current system. Comparing directly measured current shears to those estimated from hydrographic observations in 2003, we find the Baffin Island Current, but not the West Greenland Current system in geostrophic thermal wind balance. Hydrographic observations alone thus will not provide accurate velocity or flux estimates over the slope off West Greenland on account of a strong barotropic flow. We thus conclude that direct velocity measurements are needed to describe current off western Greenland.

Analysis of almost 100 years of historical hydrographic (summer) data indicates that our 2003 observations are close to a climatological mean state as defined by the NAO. Conditional averaging along isobaths reveals a more intense baroclinic counter-clockwise circulation in Baffin Bay during positive NAO years as compared to years with negative NAO. This is consistent with earlier findings by Smith (1931) of larger iceberg counts off Labrador and Newfoundland during positive NAO years. It is also consistent with recent modeling work of the Arctic Ocean that the freshwater accumulated within the Beaufort gyre (Proshutinsky et al., 2009) is released into the Atlantic preferentially during years with positive NAO (Haine, 2013, pers. comm.).

Acknowledgments

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