Multiyear Volume, Liquid Freshwater, and Sea Ice Transports through Davis Strait, 2004–10*

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ABSTRACT

Davis Strait is a primary gateway for freshwater exchange between the Arctic and North Atlantic Oceans including freshwater contributions from west Greenland and Canadian Arctic Archipelago glacial melt. Data from six years (2004–10) of continuous measurements collected by a full-strait moored array and concurrent high-resolution Seaglider surveys are used to estimate volume and liquid freshwater transports through Davis Strait, with respective annual averages of -1.6 ± 0.5 Sverdrups (Sv; $1 \text{ Sv} \equiv 10^6 \text{ m}^3 \text{ s}^{-1}$) and $-93 \pm 6 \text{ mSv}$ (negative sign indicates southward transport). Sea ice export contributes an additional $-10 \pm 1 \text{ mSv}$ of freshwater transport, estimated using satellite ice area transport and moored upward-looking sonar ice thickness measurements. Interannual and annual variability of the net transports are large, with average annual volume and liquid freshwater transport standard deviations of 0.7 Sv and 17 mSv and with interannual standard deviations of 0.3 Sv and 15 mSv . Moreover, there are no clear trends in the net transports over the 6-yr period. However, salinity in the upper 250 m between Baffin Island and midstrait decreases starting in September 2009 and remains below average through August 2010, but appears to return to normal by the end of 2010. This freshening event, likely caused by changes in arctic freshwater storage, is not apparent in the liquid freshwater transport time series due to a reduction in southward volume transport in 2009–10. Reanalysis of Davis Strait mooring data from the period 1987–90, compared to the 2004–10 measurements, reveals less arctic outflow and warmer, more saline North Atlantic inflow during the most recent period.

1. Introduction

Quantifying and understanding how and in what form freshwater is delivered from the Arctic to the North Atlantic in response to oceanic and atmospheric variability

is necessary for understanding changes in North Atlantic thermohaline circulation (Holland et al. 2001; Jahn et al. 2010). Recent changes have been observed in the Arctic, including increased air temperatures (e.g., Overland et al. 2008; Stroeve et al. 2012), decreased sea ice extent and volume (e.g., Stroeve et al. 2012; Kwok and Untersteiner 2011), and increased variability in arctic wind-driven circulation and freshwater distribution (Timmermans et al. 2011). Changes have also been observed west of Greenland and in the Canadian subarctic, such as increased river discharge (Déry et al. 2009), accelerated mass loss of west Greenland and Canadian Arctic Archipelago (CAA) glaciers (Chen et al. 2011; Gardner

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FIG. 1. General circulation in Baffin Bay and Davis Strait (white arrows) and the location of the 2004–10 moored array (red line). AW, by way of the CAA, leaves Davis Strait as the broad, surface-intensified BIC. Northward flow on the eastern side of Davis Strait consists of the fresh WGC of Arctic origin on the shelf and warm, salty WGSC of North Atlantic origin on the slope.

et al. 2011), and ice-free channels in the CAA during summer (Canadian Ice Service; <http://ice-glaces.ec.gc.ca/>). These shifts suggest that volume, freshwater, and heat transports between the Arctic and North Atlantic Oceans may also be changing. Davis Strait is one of two major oceanic gateways for exchange between the Arctic and North Atlantic Oceans (Fig. 1). Since September 2004, a comprehensive observational program in Davis Strait, including a year-round full-strait moored array and concurrent Seaglider surveys, has been focused on quantifying volume and freshwater transport variability to aid in understanding how exchanges between the Arctic and North Atlantic are being modified due to recent changes observed in the Arctic.

Outflow from the Arctic Ocean enters the North Atlantic through two major pathways, east and west of Greenland. On the west side of Greenland, arctic outflow exits through the narrow, shallow channels of the CAA (Nares Strait, Jones Sound, and Lancaster Sound), joins the cyclonic circulation within Baffin Bay, and eventually exits in the broad, surface-intensified Baffin Island Current (BIC; Tang et al. 2004; Cuny et al. 2005) through Davis Strait to the Labrador Sea (Fig. 1). Outflow from Baffin Bay through Davis Strait carries freshwater inputs from integrated CAA outflows, the West Greenland Current (WGC), glacial runoff from west Greenland and the CAA, and sea ice, precipitation, and river contributions from Baffin Bay. A small component of CAA

FIG. 2. (a) Davis Strait 2004–10 (red line) and 1987–90 (black line) moored arrays with squares indicating mooring locations. (b) Summary of Davis Strait 2004–10 moored array instrumentation. Blue crosses indicate SBE37 MicroCAT conductivity, temperature, and pressure recorders; green dots represent RDI ADCPs; black dots denote Aanderaa RCM8 velocity, conductivity, and temperature recorders; red dots denote Aanderaa RCM8 velocity and temperature recorders; and orange dots denote ULSs. Inset image shows a close-up of the Baffin Island shelf instruments. Spatial coverage varies from year to year throughout the program.

outflow bypasses Baffin Bay to flow through Fury and Hecla Strait [-0.1 Sverdrup volume (Sv; $1 \text{ Sv} \equiv 10^6 \text{ m}^3 \text{ s}^{-1}$ or $31\,536 \text{ km}^3 \text{ yr}^{-1}$) and -38-mSv freshwater, where a negative sign indicates southward transport] before entering the Labrador Sea through Hudson Strait (Straneo and Saucier 2008). On the eastern side of Greenland, freshwater exits the Arctic Ocean through Fram Strait, flows along the eastern shelf of Greenland as the East Greenland Current (EGC), continues around the southern tip of Greenland, and travels northward through Davis Strait over the west Greenland shelf as the WGC (Cuny et al. 2005).

Davis Strait, along the moored array line (Fig. 2a), extends 330 km between Baffin Island, Canada, and Greenland, with a 5-km-wide western shelf (Baffin Island) and a 113-km-wide eastern shelf (Greenland). A sill (640-m maximum depth) south of the array limits deep exchanges between Baffin Bay and the Labrador Sea. Exchanges through the strait are predominately two way and topographically steered (Tang et al. 2004). Northward flow on the eastern side of Davis Strait consists of the low-salinity WGC on the shelf and the

warm, salty West Greenland Slope Current (WGSC) of North Atlantic origin. The WGC is a combination of the EGC flowing southward from the Arctic through Fram Strait (de Steur et al. 2009) and the East Greenland Coastal Current (EGCC) arising from the addition of east Greenland coastal inflow and glacial runoff (Bacon et al. 2002; Sutherland and Pickart 2008). The WGSC is a branch of the North Atlantic Current that enters and circulates cyclonically in the Irminger Sea and continues along the east Greenland slope seaward of the EGC around Greenland (Cuny et al. 2002; Myers et al. 2007). Both the WGC and WGSC flow around the southern tip of Greenland and then turn north toward Baffin Bay. Sea ice is generally present in western and central Davis Strait between November and July, reaching a maximum in March (Canadian Ice Service; <http://ice-glaces.ec.gc.ca/>), but warm WGSC water and strong offshore winds from west Greenland keep the west Greenland slope and shelf free of ice as far north as Disko Island for most of the year.

Warm North Atlantic water can enhance glacial erosion of Greenland outlet glaciers, causing meltwater to

be released into the ocean and therefore decreasing the salinity of the WGC and ECG (Straneo et al. 2010; Holland et al. 2008; Rignot et al. 2010). While the volume of meltwater might not be significant, the addition of very fresh water might generate shallow buoyancy-driven boundary currents around Greenland. Straneo et al. (2010) observed subsurface glacial melting off the coast of east Greenland, in Sermilik Fjord, associated with warm North Atlantic waters brought from the shelf into the fjord through wind-driven exchange. Subsurface glacial erosion off the coast of central-west Greenland has been linked to warm WGSC water moving northward through Davis Strait (Holland et al. 2008; Rignot et al. 2010). Glacial meltwater from eroding west Greenland glaciers within Baffin Bay contributes to the net freshwater outflow through Davis Strait.

Changes in arctic outflow through Davis Strait may affect deep convection in the Labrador Sea, which influences the strength of the Atlantic meridional overturning circulation (Holland et al. 2001; Jahn et al. 2010). However, views differ on the nature of these linkages. Våge et al. (2009) propose that the freshwater export through Davis Strait enhanced deep convection in 2007 by increasing the northwestern Labrador Sea ice cover, allowing the cold winds from the Canadian landmass to reach the deep convective regions largely unaltered, leading to enhanced heat transport from the ocean to the atmosphere. Model studies suggest two possible outcomes of enhanced freshwater flow from the CAA into the Labrador Sea. Goosse et al. (1997) conclude that an increase in freshwater discharge into the Labrador Sea leads to pronounced surface salinity and density decreases, increased stratification, and reduced deep water formation. In contrast, model results from Myers (2005) indicate that CAA outflow is confined to the western Labrador Sea shelf and slope and has little impact on the offshore deep convection area. Based on observations, Schmidt and Send (2007) suggest that freshwater from the WGC has a stronger impact on stratification in the central Labrador Sea than the BIC does. It has been suggested that increases in southward freshwater transport along the western North Atlantic continental shelf initiate regime shifts in the shelf ecosystems and alter the abundances and annual cycles of phytoplankton, zooplankton, and higher trophic-level consumers populations (Greene et al. 2008).

Current knowledge of transports and circulation in Davis Strait rests on results from the first year of this monitoring program, excluding Seaglider surveys (Curry et al. 2011), and year-round moored measurements collected between September 1987 and September 1990 (Ross 1992). The 1987–90 array (Fig. 2a) consisted of five moorings deployed every year at six positions in the

central strait but included measurements neither over the shelves nor in the upper 150 m. Two studies (Tang et al. 2004; Cuny et al. 2005), based on the 1987–90 measurements, adopted different approaches to extrapolate across the upper 150 m and either made highly uncertain estimates of shelf contributions or confined their analyses to the central strait. Cuny et al. (2005) estimate an average net volume transport of -2.6 ± 1.0 Sv including the west Greenland shelf and shear in the upper 150 m. Tang et al. (2004) also estimate -2.6 ± 1.2 Sv of volume transport but the estimate excludes the west Greenland shelf and shear in the upper 150 m. Liquid freshwater transports from the two studies range from -99 ± 34 mSv (Tang et al. 2004) to -92 ± 34 mSv (Cuny et al. 2005) with ice contributing an additional -21.3 mSv ($-873 \text{ km}^3 \text{ yr}^{-1}$) to -12.9 mSv ($-528 \text{ km}^3 \text{ yr}^{-1}$).

This paper presents objectively analyzed (OA) estimates of volume and liquid freshwater transports through Davis Strait over the period 2004–10 using year-round moored measurements and concurrent Seaglider surveys. The complementary combination of high temporal resolution provided by the moored array and fine spatial resolution achieved by Seaglider surveys allows for more accurate quantification of transports and uncertainties than previously possible. Seaglider surveys began in September 2005, with occupations during the ice-free months, and were extended to provide cross-strait sections during the ice-covered periods starting in 2008. Volume and freshwater transports by sea ice are estimated using sea ice velocity and concentration data derived from passive microwave measurements by the Advanced Microwave Scanning Radiometer for Earth Observing System (AMSR-E), National Centers for Environmental Prediction and National Center for Atmospheric Research (NCEP–NCAR) reanalysis products, and sea ice thickness estimated from upward-looking sonar (ULS) measurements. Section 2 introduces the data used in this study and section 3 briefly describes the methods employed to estimate transports and corresponding uncertainties. Additional details about the data, data processing, and methods are contained in the online supplemental material. Results are presented and discussed in section 4 in terms of along-strait velocity, temperature, and salinity variability; net and water mass component volume and liquid freshwater transports; and sea ice transport. Section 5 describes and investigates the causes of a freshening event observed starting in September 2009 in the upper 250 m between Baffin Island and midstrait. The 1987–90 moored measurements are reanalyzed using similar OA methodology and transports between the two time series are compared in section 6. The focus of this paper is to quantify and describe transport variability in Davis Strait. A complementary paper investigating the forces

TABLE 1. Davis Strait ULS sea ice thicknesses.

Instrument Year	Monthly avg thickness (m)									
	Nov	Dec	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug
C1										
2005–06	0.31	0.93	1.58	1.98	2.17	1.80	2.03	1.23	1.10	—
2007–08	0.53	1.42	2.88	2.39	2.46	2.94	2.69	0.92	—	—
C2										
2006–07	—	0.22	0.59	0.75	0.80	1.00	0.83	0.46	0.53	0.13
2007–08	0.43	0.94	1.91	2.32	2.13	2.07	2.33	1.50	—	—
C3										
2007–08	—	0.31	0.62	0.94	1.03	0.75	0.67	0.22	—	—
2008–09	0.47	0.53	0.77	0.88	1.14	0.89	1.15	1.07	1.29	—
C4										
2006–07	—	0.23	0.43	0.56	0.76	0.85	0.85	0.60	0.25	—
2007–08	—	0.26	0.63	0.90	0.86	0.89	0.61	0.68	—	—
C6										
2006–07	—	—	—	0.48	0.48	0.51	0.61	—	—	—
	Multiyear monthly avg thickness (m)									
C1	0.42	1.18	2.23	2.19	2.32	2.37	2.36	1.08	1.10	—
C2	0.43	0.58	1.25	1.54	1.47	1.54	1.58	0.98	0.53	0.13
C3	0.47	0.42	0.70	0.91	1.08	0.82	0.91	0.65	1.29	—
C4	—	0.25	0.53	0.73	0.81	0.87	0.73	0.64	0.25	—
C6	—	—	—	0.48	0.48	0.51	0.61	—	—	—

controlling the observed variability is forthcoming and only briefly addressed in this paper.

2. Data

The Davis Strait observing system began operating in 2004 with the goal of providing sustained, long-term quantification of Arctic–subarctic exchange west of Greenland. The system was designed to quantify transports and associated uncertainties, with the aim of isolating secular trends associated with environmental change and understanding the mechanisms driving observed changes. The observing system consists of year-round moorings (eight or nine on the shelf and six in the central strait; Fig. 2); continuous, year-round Seaglider-based sections in the central strait, roughly between the 500-m isobaths; and annual autumn hydrographic sections with chemical sampling. The array was positioned north of the sill, at a maximum depth of 1040 m, in an attempt to avoid bathymetrically induced flows that affected the interpretation of the results from 1987–90 array (Ross 1992) and to allow for concurrent Seaglider surveys. Seaglider surveys were conducted in both ice-covered and ice-free conditions, with at least two complete central strait surveys for each calendar month.

a. 2004–10 moored array

Teledyne RD Instruments (RDI) Acoustic Doppler Current Profilers (ADCP), Aanderraa Recording Current Meters (RCM8), and Sea-Bird Electronics (SBE)

37 MicroCATs measured velocity, temperature, and conductivity at 30 min (ADCP and MicroCAT) and hourly (RCM8) intervals (Fig. 2). On the shelves, SBE37 MicroCATs inside iceberg resistant floats (IceCATs) that are inductively coupled via a weak link to bottom-mounted dataloggers, measured conductivity and temperature in the upper 20–50 m at 5-min intervals. IceCATs mitigate the risk of data loss due to impacts of advected ice by preserving all measurements in a bottom-mounted datalogger that is isolated from the near-surface elements by a weak link. Salinity is computed using the 1978 Practical Salinity Scale equations and constants (Perkin and Lewis 1980). Spatial coverage varied during the program due to changes in instrumentation, losses from instrument failure, and unrecovered moorings. In the online supplemental material, appendix A describes how the mooring data were processed and includes a summary of the locations, depths, record lengths, and types of instruments deployed between September 2004 and September 2010 (Table 1).

b. Seagliders

Starting in September 2005, autonomous Seagliders (Eriksen et al. 2001) were added to the monitoring program to resolve temperature, salinity, and density variability at scales smaller than the mooring separation distances and to provide consistent measurements in the region near the ice–ocean interface. Seagliders are small, long-endurance autonomous underwater vehicles that propel themselves through the water by changing

TABLE 2. Davis Strait uncertainties and transports.

Years	Volume (Sv)				Liquid freshwater (mSv)			
	OA	Bias	Background	Total	OA	Bias	Background	Total
2004–05	0.2	0.5	0.1	-2.0 ± 0.5	12	10	2	-100 ± 16
2005–06	0.2	0.4	0.1	-1.7 ± 0.5	10	10	3	-97 ± 15
2006–07*	0.3	0.7	0.5	-1.6 ± 0.9	17	19	17	-90 ± 30
2007–08	0.1	0.3	0.1	-1.3 ± 0.4	9	9	7	-66 ± 15
2008–09	0.2	0.4	0.1	-1.8 ± 0.4	9	10	3	-110 ± 14
2009–10	0.2	0.5	0.2	-1.5 ± 0.5	13	14	6	-95 ± 20
2004–10	0.1	0.2	0.1	-1.6 ± 0.2	5	5	3	-93 ± 6

* Uncertainty is higher because mooring C3 was not recovered.

their buoyancy such that they alternately sink and rise. Seaglid­ers change their center of gravity to control their attitude, using wings and a rudder to project vertical motion into the horizontal. This allows Seaglid­ers to control their course and navigate from waypoint to waypoint as they sawtooth in depth. Seaglid­ers exploit this capability to occupy sections repeatedly across the central strait, roughly following the mooring line. Temperature and conductivity along dive paths are bin averaged onto a regular depth grid at 5-m intervals. Seaglid­ers profile from the sea surface (or approximately 5 m below the ice–ocean interface, when ice is present) to within 10 m of the bottom depth, as defined by the International Bathymetric Chart of the Arctic Ocean (IBCAO; Jakobsson et al. 2008) and by soundings from a short-range acoustic altimeter. They surface at 2–6-h intervals, with approximately 1–4-km horizontal distance between surfacings, depending on dive depth, and take roughly 10 days to complete one central strait section. The typical horizontal speed is 0.25 m s^{-1} (20 km day^{-1}). Missions completed between September 2005 and June 2011 yielded 46 full central strait sections with at least two sections within each calendar month. Details of the sections used in this analysis are summarized in appendix A of the online supplemental material (Table 2). Observations collected by Seaglid­ers are projected onto the mooring line to produce a series of standard sections. Further details about the data processing are presented in appendix A of the online supplemental material.

c. Sea ice data

1) UPWARD-LOOKING SONARS

Applied Physics Laboratory, University of Washington Mark 2 ULSs collected sea ice draft d measurements above the moorings in the central strait at preprogrammed intervals (Fig. 2b). The ULS emits a 1-ms pulse of high-frequency, narrow-beam, vertically oriented acoustic energy into the water column and then measures and records the round-trip travel time τ between emission of the pulse and detection of the signal

returning from the underside of the ice, or from the sea surface. Simultaneously, the ULS measures and records the in situ water pressure P_{uls} . The range R to the target is estimated from τ , and depth D below sea level is estimated from P_{uls} . The ice draft is estimated as depth minus range,

$$d = D - R,$$

and ice thickness h (m) is estimated as

$$h = 1.14d.$$

Thickness estimates (draft + freeboard) do not take into account snow cover on the ice and the multiplier (1.14) is less accurate for ridged ice than for smooth ice. The ULS data are averaged over the time series for each calendar month, excluding “open water” data points, defined here as $h < 5 \text{ cm}$ in winter and $h < 10 \text{ cm}$ in summer. The resulting sample statistic represents the monthly mean thickness of sea ice at points where there is ice, not open water. Further details about the ULS data processing and errors are presented in appendix A of the online supplemental material.

2) AMSR-E

Daily sea ice area transports across Davis Strait are computed using daily ice concentration estimates and gridded 89-GHz brightness temperature fields from the AMSR-E radiometer on the National Aeronautics and Space Administration (NASA) *Aqua* platform (National Snow and Ice Data Center; <http://nsidc.org/data/amsre/>; Cavalieri et al. 2004a,b).

3) CANADIAN ICE SERVICE

Daily ice coverage maps provided by the Canadian Ice Service (<http://ice-glaces.ec.gc.ca/>) are used to define surface ocean temperatures when ice is present over the array line. Ice is considered either present or not and no distinction is given between types of ice.

d. 1987–90 moored array

An array of five moorings deployed across central Davis Strait, south of the 2004–10 array (Fig. 2a), employed RCM5s to collect hourly measurements of velocity, temperature, and conductivity at depths of 150, 300, and 500 m from September 1987 to September 1990 (Ross 1992; Cuny et al. 2005). Quality-controlled data were obtained from the Bedford Institute of Oceanography (BIO) data archive (<http://www.mar.dfo-mpo.gc.ca/science/ocean/>). Data were processed in the same manner as the 2004–10 moored array data. Details are presented in appendix A of the online supplemental material.

3. Methods

a. Objective analysis

A modified objective analysis procedure (Bretherton et al. 1976) is used to construct daily, full-strait sections of along-strait velocity V , temperature T , and salinity S for the 2004–10 and 1987–90 datasets. The use of OA allows for the separation of the low-frequency background (average) field from the variable (anomaly) field to systematically deal with large spatial and temporal gaps. A Gaussian covariance function,

$$C(x) = \sigma^2 \exp \left[- \left(\frac{\Delta x}{L_x} \right)^2 \right],$$

based on horizontal separation Δx , decorrelation length scale L_x , and data anomaly variance σ^2 is used to describe the weight of a given observation on the surrounding grid points. Correlation functions calculated from the moored observations and compared with results from the high-resolution hydrographic and Seaglider sections yield horizontal decorrelation length scales of 20 km (V) and 40 km (T and S).

Mooring data and Seaglider T and S sections are used to calculate average OA large-scale background V , T , and S fields for each year; these fields are used with all six years of observations. When spatial gaps between the moorings exceed the decorrelation length scales, the OA relaxes to the background fields. Using low-frequency background fields to fill spatial gaps reduces the connection between total daily OA V , T , and S results and higher-frequency variability present in the daily V , T , and S data due to forcings such as winds. Monthly background field V , T , and S profile shapes at the mooring locations are used in combination with the daily moored observations to create daily full-depth V , T , and S profiles from moored observations collected at discrete depths. Formulation of the background fields and creation of the

daily full-depth V , T , and S profiles are detailed in appendix B of the online supplemental material.

b. Transport calculations

Daily volume and liquid freshwater transports are calculated for each grid cell, summed over the grid domain, and time averaged to compute monthly and annual transports. Annual transports start in October and end in September of the following year. September 2010 transports are an average of September 2005–09 transports due to the lack of September 2010 data. Volume transport is calculated by multiplying the mapped velocity field at each grid cell by its area. Positive values indicate northward transport into Baffin Bay; negative values indicate southward transport into the Labrador Sea. Transports are estimated between the surface and sill depth (640 m). Total area across the strait is 133 km² with the west Greenland shelf contributing 7 km². Liquid freshwater transport is calculated as

$$FW_{\text{transport}} = \sum V_i \frac{S_0 - S_i}{S_0} A_i,$$

where V_i , S_i , and A_i are the mapped grid cell velocity, salinity, and area, respectively. The reference salinity ($S_0 = 34.8$), considered the average Arctic Ocean salinity (Aagaard and Carmack 1989), is chosen to maintain consistency with previous studies (Tang et al. 2004; Cuny et al. 2005).

c. Sea ice transport calculations

Daily sea ice area transport through Davis Strait between November and May for each year is estimated using AMSR-E satellite data following the methods presented in Kwok (2007) using the latest data available. Sea ice area transport is estimated north of the moored array along a line at $\sim 68^\circ\text{N}$ spanning ~ 460 km across Baffin Bay (see Fig. 1b in Kwok 2007). The perpendicular component of the ice motion is integrated across the strait using the trapezoidal rule and the motion is constrained to zero at the coastal endpoints. Area estimates are weighted by the AMSR-E ice concentration to account for open water areas.

Sea ice volume transports are estimated using the monthly time-averaged ice thickness derived from the ULS data. Monthly ice thicknesses from the moored ULS (Table 1) are interpolated or extrapolated to obtain monthly ice thicknesses across the ice area transport line. Monthly ULS ice thicknesses are multiplied by the daily area transports across the strait and averaged to obtain monthly ice volume transports. Because of low ULS data returns, multiyear averages of the monthly ice thicknesses are used to fill data gaps when no data are

available (Table 1). Freshwater export via sea ice is calculated as

$$FW_{ice\ transport} = Vol_{ice} \frac{S_0 - S_{ice}}{S_0} \frac{\rho_{ice}}{\rho_{water}},$$

where Vol_{ice} is the annual sea ice volume transport through Davis Strait, $S_{ice} = 5$ is the sea ice salinity chosen to be consistent with Cuny et al. (2005), ρ_{ice} is its density (900 kg m^{-3}), and ρ_{water} is the density of freshwater (1000 kg m^{-3}). Monthly ice transports are averaged to obtain annual and the 6-yr average estimates.

d. Uncertainty analysis

Transport uncertainties for the 2004–10 and 1987–90 datasets are estimated by combining in quadrature three types of errors, those due to (i) noise (data – large-scale background fields) as estimated by the OA, (ii) assumptions regarding the creation of the large-scale background fields, estimated by altering the background fields, and (iii) biases due to spatial gaps in the moored array, estimated using data variance and the OA covariance function. Monthly errors for all three sources are averaged and divided by the square root of the number of monthly estimates minus one to obtain yearly error estimates. Similarly, yearly error estimates are averaged and divided by the square root of the number of years minus one to obtain the 6-yr (2004–10) and 3-yr (1987–90) average errors. The three types of error are summed in quadrature to determine total transport uncertainties. Details of the volume and liquid freshwater transport uncertainty analysis are discussed in appendix C of the online supplemental material.

Sea ice volume transport uncertainty is estimated following Kwok and Rothrock (1999) as

$$\sigma_{FW_{ice\ transport}} = \sqrt{(A_{ice}\sigma_h)^2 + (h\sigma_{A_{ice}})^2},$$

where A_{ice} is the annual-average ice area transport through Davis Strait, h is the annual-average ice thickness (1 m), σ_h is the average ice thickness standard error of the mean (0.1 m), and $\sigma_{A_{ice}}$ is the error in the ice area transport. Uncertainty in the ice area transport over a given interval is calculated as

$$\sigma_{A_{ice}} = \sigma_u L \sqrt{\frac{N_d}{N_s}},$$

where σ_u is standard error in the motion estimates (3 km), L is the length of coverage across the flux gate ($\sim 68^\circ\text{N}$), N_s ($=7$) is the number of independent samples across the gate, and N_d is the number of daily estimates in the record.

4. Results and discussion

a. General circulation and water mass variability

Exchanges through Davis Strait are predominantly two way and topographically steered (Tang et al. 2004). Only the along-strait component of the velocity is discussed here and, for brevity, described using north (into Baffin Bay) or south (toward the Labrador Sea). Southward flow extends over half the strait from Baffin Island until just off the west Greenland slope (175 km from Baffin Island) and northward flow extends over the west Greenland slope and shelf. Both maximum southward and maximum northward transport occur between October and December. Sea ice is generally present between November and July with peak coverage January to March.

Objectively analyzed results of V , T , and S are presented as 6-yr averages of mean monthly sections across the strait (Fig. 3). The accuracy and time–space resolution of the data presented for Davis Strait are unprecedented and allow for full-strait descriptions of V , T , and S variability and estimates of transports. These data comprise the basis for distinguishing variability on annual and interannual time scales. Data are also grouped into four different water mass classes identified in Davis Strait. Each water mass travels a unique path before crossing Davis Strait and suggests that the water mass properties and transports vary independently. Four primary water masses are identified by salinity and potential temperature θ following Tang et al. (2004):

- Arctic Water (AW; $\theta \leq 2^\circ\text{C}$; $S \leq 33.7$) is cold, low-salinity water flowing southward as the BIC at depths $<300\text{ m}$. Although of Arctic origin, this water has been modified by glacial runoff, air–sea fluxes, and local sea ice melt and formation in Baffin Bay and the CAA.
- West Greenland Irminger Water (WGIW; $\theta > 2^\circ\text{C}$; $S > 34.1$) is warm, saline water that originates in the North Atlantic and flows cyclonically around the Irminger Sea. It then rounds the southern tip of Greenland and travels northward along the west Greenland slope as part of the WGSC, passing through Davis Strait as a mostly barotropic current.
- West Greenland Shelf Water (WGSW; $\theta < 7^\circ\text{C}$; $S < 34.1$) is ultimately of Arctic origin, with contributions from Greenland glacial runoff. The WGSW travels from east Greenland, turns northward at Cape Farewell, and flows through Davis Strait as the WGC along the west Greenland shelf.
- Transitional Water (TrW; $\theta > 2^\circ\text{C}$; $S > 33.7$), usually found at depths $>300\text{ m}$. The TrW is the product of water masses that flow into Baffin Bay, mix and undergo

FIG. 3. OA monthly averages from daily estimates of along-strait (a) velocity, (b) salinity, and (c) temperature. Pink bars indicate areas along the moored array that are covered, on average, by sea ice 60% of the month. Gray lines indicate mooring locations in (a). Potential density lines (black lines) are noted in (b). The boundaries of the four dominant water masses (AW, WGIW, WGSW, and TrW) are shown for October in (c).

FIG. 4. OA water mass monthly averages for each year from daily along-strait velocity (negative = southward), salinity, and potential temperature estimates with the 2004–10 averages (annual cycles) noted by the black lines. Boundaries for the water masses (AW, WGIW, WGSW, and TrW) are defined in section 4a.

local modifications, and flow south through Davis Strait at depth.

The locations of the cores of the four water masses are indicated in Fig. 3c (top, left). The locations of the major current systems (BIC, WGC, and WGSC) are indicated in Fig. 1. The annual variation of the monthly average velocity, potential temperature, and salinity, averaged over each water mass class, is shown in Fig. 4 for each of the six years and for the 6-yr average (referred to in this paper as the annual cycle). Monthly averages are also presented as T – S diagrams (Fig. 5) where symbols identify the water masses and the color shading quantifies the velocity.

1) ARCTIC WATER

Changes in AW properties reflect changes in arctic outflow. Average AW velocities are southward the entire period, largest (-0.07 m s^{-1}) in September–October and smallest (-0.04 m s^{-1}) in December–January (Fig. 4). Velocity increases again between June and July, except during 2007 and 2008. Velocity variability is large, with an average interannual coefficient of

variability (ratio of the standard deviation to the absolute value of the mean) of 0.23 based on all of the monthly ($n = 71$) AW velocities and an annual coefficient of variability of 0.15 based on the six annual averages of AW velocity.

Although AW velocity is southward on average, small wintertime velocities are often accompanied by northward flow along the Baffin Island slope (Fig. 3). Velocities at mooring C1, both in the upper 100 m and at 250 m, and sometimes at C2 at 200 m, reverse and are directed northward between December and February (average $\sim 0.04 \text{ m s}^{-1}$). Smaller southward velocities on the Baffin shelf and in the upper 100 m at mooring C2 are typically observed during these reversals (Fig. 6). The timing and magnitude of the reversals vary, but the largest (peak velocities of 0.08 m s^{-1}) occurred in 2007–08 and the smallest (0.01 m s^{-1}) in 2008–09. Cuny et al. (2005) also observed flow reversals in the 1987–90 mooring data at M1 between November and February with peak velocities of 0.12 m s^{-1} and suggested local seasonal eddies as the cause rather than advection from the south. The wintertime reversals during 1987–90 are evident in the monthly along-strait currents at site A

FIG. 5. Monthly average potential temperature–salinity diagrams noting along-strait velocity and delineated into water mass classes. Boundaries for the water masses (AW, diamonds; WGIW, squares; WGSW, stars; TrW, circles) are defined in section 4a. Gray dots indicate unclassified water.

[station M1 here and in Cuny et al. (2005)] also shown in Tang et al. (2004) in their Fig. 23a. Shallower and more inshore measurements from 2004 to 2010 confirm that the reversals are not accompanied by shifts in T and S and are likely caused by eddies.

Salinity varies annually according to the timing of sea ice formation and melt within Baffin Bay as well as the timing of discharge of meltwater from Greenland and CAA glaciers. On average, ice starts to form within Baffin Bay in September, reaching a maximum extent in March, and begins to melt in April with ice-free conditions in August (Canadian Ice Service). Greenland surface melt generally occurs between June and August (Mote 2007). As sea ice forms in Baffin Bay, salt is rejected and stratification weakens as the dense water mixes downward. Arctic Water salinity increases during ice formation, reaching a maximum in April–May, with the annual cycle maximum (33.16) occurring in May (Fig. 4). The average monthly salinity was the highest of all years in April 2006, having a value of 33.28. When sea ice and glaciers begin to melt, the addition of freshwater increases stratification and decreases AW salinity. Salinity reaches a minimum in December–January, with the annual cycle minimum (32.92) occurring in January.

A simple model is used to determine if the observed increase in AW salinity is consistent with local ice growth using the conservation of mass and salt and annual cycle of AW salinity. The conservation of mass is expressed as

$$\rho_{w1}H_{w1}A = \rho_{ice}H_{ice}A + \rho_{w2}H_{w2}A,$$

and the conservation of salt is expressed as

$$\rho_{w1}H_{w1}S_{w1}A = \rho_{ice}H_{ice}S_{ice}A + \rho_{w2}H_{w2}S_{w2}A,$$

where ρ_{ice} is the ice density (900 kg m^{-3}), $\rho_{w1} = \rho_{w2} = 1027 \text{ kg m}^{-3}$ is the density of AW when ice begins to form and when ice begins to melt, S_{ice} is the salinity of ice (5), S_{w1} is the salinity of AW when ice begins to form, S_{w2} is the salinity of AW when ice begins to melt, $H_{w1} = 250 \text{ m}$ is the depth of AW as ice begins to form, H_{w2} is the unknown depth of AW as ice begins to melt, $H_{ice} = 2 \text{ m}$ is the ice thickness, and A is the area. The areas are equal when considering a column of water. Ice begins to form in Davis Strait in November ($S_{w1} = 32.95$) and the predicted increase in salinity ($S_{w2} = 33.15$), using the above formulas, agrees well with the annual maximum salinity in May (33.16). In addition to ice formation,

FIG. 6. Daily along-strait velocities on the Baffin Island slope (left) from October 1987 to October 1990 at mooring M1 for 150 m (black) and 300 m (gray) and (right) from October 2004 to October 2010 at mooring C1 for upper 100 m average (gray) and 250 m (black).

vertical mixing with TrW can also contribute to the increase in salinity observed. A vertical eddy diffusivity coefficient from 5×10^{-4} to $10 \times 10^{-4} \text{ m}^2 \text{ s}^{-1}$, a reasonable value for the upper 350 m, would be required for mixing to make a significant contribution to the AW layer. Vertical mixing should be considered in a more complete model looking at AW seasonal salinity variability.

Starting in September 2009 and extending through August 2010, AW salinity was much lower than all other years and is not consistent with local sea ice melting. Using the same equations above, instead to estimate the decrease in salinity due to local ice melting using the 2008–09 maximum salinity in May 2009 ($S_{w2} = 33.15$), salinity is estimated to decrease to 32.95 (S_{w1}) due to local melting, yet the observed 2009–10 minimum average salinity in November 2009 is 32.79. This freshening event is discussed in more detail in section 5.

The annual variation of AW temperature follows the annual variation of climatological air temperatures, with a lag of about 1 month. Average air temperatures are lowest in January–March with the minimum in February and warmest in June–August with the maximum in July (Environment Canada, National Climate Data and Information Archive, Canadian Climate Normals 1971–2000; <http://www.climate.weatheroffice.gc.ca/>). The annual cycle of AW temperature has a minimum (-1.5°C) in April and maximum (-0.6°C) in August. The average annual range of air temperature is $\sim 34^\circ\text{C}$ compared to $\sim 0.9^\circ\text{C}$ for AW. November 2006–January 2007 was anomalously warmer (average $\sim 0.2^\circ\text{C}$) than the annual

cycle. This warming also coincides with below average AW salinities in October–December 2006.

2) WEST GREENLAND IRMINGER WATER

Monthly average WGIW velocities are northward except between July and August 2010, when it is weakly negative (-0.01 m s^{-1}). Maximum velocity typically occurs between October and December. Velocity generally decreases after December reaching a minimum between June and August (Fig. 4). While the timing of the yearly maximum is fairly consistent from year to year, the timing of the minimum varies. The annual minimum velocity is in August (0.01 m s^{-1}) but the timing of the yearly minimum ranges from April to August and even December during 2006–07. Velocity variability is large, with an average interannual coefficient of variability of 0.71 based on all of the monthly WGIW velocities and an annual coefficient of variability of 0.12 based on the six annual averages of WGIW velocity.

Annual variations in WGIW velocity are connected with variability upstream in the Irminger and Labrador Seas. During the same time as the velocity yearly maximums, polar cyclones move eastward across the Labrador and Irminger Seas. Polar cyclones cause intense cyclonic wind stress on both sides of Greenland, which in models has been shown to enhance cyclonic circulation in the Irminger and Labrador Seas during autumn and winter (Spall and Pickart 2003). Observations from de Jong et al. (2012) support the modeling hypothesis and show that there is enhanced circulation in the Irminger Sea during autumn and winter. Baroclinic Rossby waves develop to maintain cyclonic circulation in the summer

until the wind increases again the following autumn (Spall and Pickart 2003).

Annual and interannual WGIW variations in velocity, temperature, and salinity prior to this monitoring program were not well resolved. The annual cycle of salinity exhibits two peaks, although these vary interannually in magnitude and timing. The first peak occurs between October and January and the second between March and May. In contrast, temperature has a clear annual signal with maximum values in November–December then decreasing until reaching a minimum in July–August. The annual maximum occurs in November (3.9°C) and the minimum in July (2.7°C). The mechanisms driving WGIW variability in and upstream of Davis Strait are poorly understood. Eddies shed off the west Greenland coast into the Labrador Sea during autumn likely drive lateral transports of heat and salt, and likely modulate transport of WGIW through Davis Strait (Lilly et al. 2003; Prater 2002).

3) WEST GREENLAND SHELF WATER

At the location of the mooring array, the west Greenland shelf extends 113 km from the coast to the shelf break, 34% of the total cross-strait distance. Monthly variations of WGSW V , T , and S are the most consistent from year to year compared to the other water masses (Figs. 4 and 5). Velocity is northward in all months, with maximum values in September–November and then decreasing until reaching a minimum in February–April. The annual maximum occurs in September (0.07 m s^{-1}) and the minimum in April (0.02 m s^{-1}). Velocity variability is large, with an average interannual coefficient of variability of 0.42 based on all of the monthly WGSW velocities and an annual coefficient of variability of 0.11 based on the six annual averages of WGSW velocity.

Salinity variability over the shelf is driven by upstream variations in arctic freshwater and sea ice outflow through Fram Strait and freshwater (both sea ice and glacier runoff) contributions from and around Greenland via the EGC and EGCC (Bacon et al. 2002; Sutherland and Pickart 2008; de Steur et al. 2009). Yearly maximum salinity occurs in April–June with the annual maximum (33.66) occurring in May. The maximum salinity (33.85) in 2005–06 was 0.18 higher than average. Salinity typically remains fairly constant between April and June then starts to decrease, reaching a yearly minimum in August–October with the annual minimum (32.75) occurring in October. Shelf salinities are freshest at station WG4, 34 km from the coast, between April and September. Using archived hydrographic surveys, Cuny et al. (2005) also noticed a salinity minimum ~ 20 km offshore of west Greenland during the summer. They suggest this could be

a continuation of the East Greenland Coastal Current, an inner branch of the EGC, driven by Greenland ice melt and glacial runoff as described by Bacon et al. (2002) and Sutherland and Pickart (2008).

The required ice thickness necessary to account for the annual cycle of WGSW salinity due to local sea ice formation and melt is 1.66 m using the annual cycle of WGSW salinity and conservation of mass and salt equations presented earlier, where ρ_{ice} is the ice density (900 kg m^{-3}), $\rho_{w1} = 1026$ is the density of WGSW when ice begins to form in November, $\rho_{w2} = 1027\text{ kg m}^{-3}$ is the density of WGSW when ice begins to melt in May, S_{ice} is the salinity of ice (5), $S_{w1} = 32.86$ is the salinity of WGSW when ice begins to form, $S_{w2} = 33.66$ is the salinity of WGSW when ice begins to melt, $H_{w1} = 50\text{ m}$ is the depth of WGSW as ice begins to form, and H_{w2} is the unknown depth of WGSW as ice begins to melt. Ice along the west Greenland shelf at the array line is seasonal and extrapolating ice thickness from the slope (Table 1) suggests that seasonal ice growth over the shelf is less than 1 m with seasonal ice growth accounting for only about 40% of the annual salinity cycle. To achieve the annual maximum salinity, other processes such as exchange with more saline waters over the slope must be important.

Annual atmospheric variations likely drive the strong annual temperature cycle on the shelf. Annual minimum WGSW temperatures occur in March–April and continue to rise as air temperatures increase. The shelf reaches a maximum temperature in August–September with the average maximum occurring in August (4.2°C). The annual minimum occurs in March (-1.0°C), one month earlier than for AW.

4) TRANSITIONAL WATER

Transitional Water is the largest water mass by area in Davis Strait and is composed of water that is modified during its passage through Baffin Bay and then exits through Davis Strait at depths below 250 m. Annual variations in TrW water properties are less than the other water masses in Davis Strait. Velocity is weakly southward for all months with minimum velocities in March–May and the annual minimum occurring in April (-0.01 m s^{-1}). Even though the maximum southward velocity in the annual cycle occurs in July (-0.02 m s^{-1}), yearly maximums occur in either October or June–July. Velocity variability is large, with an average interannual coefficient of variability of 0.28 based on all of the monthly TrW velocities and an annual coefficient of variability of 0.11 based on the six annual averages of TrW velocity.

Topographic steering in Baffin Bay creates stronger velocities along the edges and weaker velocities in the

interior at depth (Tang et al. 2004). In most months, just west of the northward flowing WGSC, a surface-intensified southward current is present off the west Greenland slope (Fig. 3). This southward current might be an extension of the BIC that has been directed eastward, following the isobaths, and then southward seaward of the west Greenland slope. The current is also present in the 1987–90 M5 mooring time series (Tang et al. 2004; Cuny et al. 2005). Tang et al. (2004) observed the current in model results along the 1000- and 1500-m isobaths off the west Greenland shelf between 72° and 68°N and then joining the southward flow off Baffin Island around 67°N. Forcing mechanisms driving the deep flow of TrW through Davis Strait are poorly understood. They likely arise from local topographic controls, winter convection, cyclonic circulation within Baffin Bay, and large-scale sea level pressure (SLP) and sea surface height (SSH) variations between the Arctic and North Atlantic (Tang et al. 2004; Rudels 2011).

Salinity variations in TrW are <0.1 with the annual minimum (34.27) occurring in December then building up to the annual maximum (34.29) in August. Interannual variability is also <0.1 with no statistically significant annual cycle. In contrast, TrW temperature annual variations are more consistent from year to year. Maximum temperatures occur in June–August with the annual maximum occurring in July (1.2°C). Temperature decreases after August and typically reaches a minimum in November–December with an annual minimum of 0.9°C in December. Pockets of warm TrW observed in some of the summer Seaglider sections suggest possible advection of warm WGIW into the central strait.

b. Transports

Daily V , T , and S OA results are combined with the cross-strait area and averaged to obtain annual and monthly volume and liquid freshwater transports (Figs. 7 and 8). Yearly transports and corresponding errors are presented in Table 2. Annual transports from the OA background fields are -1.6 Sv and -113 mSv for volume and liquid freshwater, respectively (Fig. 8).

Daily volume and liquid freshwater transports are also computed based on water mass classification and averaged to obtain monthly and annual water mass transports (Fig. 9). Net transport through Davis Strait is the summation of transport from the four water masses and unclassified waters. Water undefined by the water mass classes are omitted from Fig. 9 and represent 0.1 Sv and 4 mSv of the respective total average volume and liquid freshwater transports.

Interannual and annual variability of the net transports are large, with average annual volume and liquid

FIG. 7. OA yearly (October–September) (top) volume transport and (middle) liquid freshwater transport from 2004 to 2010 (negative = southward transport) including uncertainties and (bottom) the corresponding yearly sea ice freshwater transport. Annual-average transports and uncertainties are noted in the top right. Freshwater transports are referenced to 34.8. All transports are estimated from the surface to the sill depth, 640 m.

freshwater transport standard deviations of 0.7 Sv and 17 mSv and interannual standard deviations of 0.3 Sv and 15 mSv. For the annual volume and liquid freshwater transports (Fig. 7, Table 2), the respective interannual coefficients of variability are 0.15 and 0.16; monthly variability is larger with coefficients of variability of 0.43 and 0.23 (Fig. 8).

1) VOLUME TRANSPORTS

Yearly net volume transport through Davis Strait is southward for all years with an average transport of -1.6 ± 0.2 Sv (Fig. 7). Although the net southward transport decreases steadily between 2004 and 2008, transport increases again between 2008 and 2010 and no significant trend is observed (p values > 0.2) in the yearly or monthly transports (Figs. 7 and 8). Because volume transport depends on velocity and cross-strait area, monthly volume transport variability is similar to monthly velocity variability (Figs. 4 and 9). Maximum net transport occurs in 2004–05 (-2.0 ± 0.5 Sv) as a result of increased southward TrW transport in the deep central strait (Fig. 9). Minimum net transport occurs in

FIG. 8. OA monthly (top) volume transport and (bottom) liquid freshwater transport between October 2004 and September 2010 (negative = southward transport). Shaded areas denote the monthly uncertainty. For comparison, the OA background transports are shown with a dashed line. September 2010 transports are replaced with average September values due to lack of September 2010 data. Freshwater transports are referenced to 34.8. All transports are estimated from the surface to the sill depth, 640 m. Uncertainties are higher in 2006–07 due to the loss of the deepest central strait mooring, C3.

2007–08 (-1.3 ± 0.4 Sv), primarily from reduced southward transport of AW, due to longer periods of flow reversal along the Baffin slope and increased WGSW northward transport in November–December. Both AW and TrW have monthly average southward transports for all years with average contributions over the six years of -1.8 and -1.1 Sv. Northward contributions

come from WGIW and WGSW, with average transports of 0.4 and 0.7 Sv, plus an additional 0.1 Sv from undefined water. Maximum northward transport of WGIW and WGSW occurs at the same time as maximum southward transport of AW and TrW, resulting in a September–November minimum in net monthly transport, consistent with the strength of cyclonic circulation

FIG. 9. OA monthly average water mass volume and liquid freshwater transports for each year (negative = southward transport) with the 2004–10 averages (annual cycles) noted by the black lines. Boundaries for the water masses (AW, WGIW, WGSW, and TrW) are given in section 4a. Freshwater transport is referenced to 34.8.

in Baffin Bay and the Labrador Sea (Spall and Pickart 2003; Tang et al. 2004). In contrast, timing of minimum transport varies for each water mass.

The largest contribution to volume transport is made by AW. Variability in AW transport is driven primarily by variability in outflow through the CAA passages into Baffin Bay (Lancaster Sound, Jones Sound, and Nares Strait) in response to regional and SLP and SSH variations within Baffin Bay and between the Arctic and North Atlantic (Tang et al. 2004; Cuny et al. 2005; Rabe et al. 2012; Peterson et al. 2012). Transports through the CAA passages have been estimated using mooring data from various periods spanning the last 13 years. The longest time series, through Lancaster Sound, has been monitored year-round since 1998. Results from Lancaster Sound indicate an average transport of -0.46 ± 0.09 Sv over the 13-yr record with maximum transport in July–August and minimum transport in November–December (Peterson et al. 2012). Peterson et al. (2012) show that monthly alongshore wind anomalies in the Beaufort Sea account for 43% of the variance of the Lancaster Sound transport anomalies (p value < 0.01). In general, a cyclonic wind pattern centered in the area of the Arctic high and the Beaufort Gyre would favor a larger SLP gradient between both ends of Lancaster Sound and as a result, increased volume transport through the sound (Peterson et al. 2012). In 2007, the Beaufort Gyre experienced the largest anticyclonic wind-driven circulation in 60 years (Proshutinsky and Johnson 2010), which would reduce transport through Lancaster Sound, and thus Davis Strait. The AW transport minimum in 2007–08 (-1.3 Sv) through Davis Strait is consistent with reduced transport through Lancaster Sound (Peterson et al. 2012). Nares Strait has been monitored year-round over the periods 2003–06 and 2009–12. Results from 2003 to 2006 show maximum southward transport in January–June, minimum southward transport in July–December with an average full-depth transport of -0.72 ± 0.11 Sv (Münchow and Melling 2008). Moorings in Jones Sound during 1998–2002 indicate that an additional -0.3 ± 0.1 Sv flows through the smallest passage of the CAA (Melling et al. 2008). The sum of the CAA volume transports (-1.5 Sv) represents 82% of the average AW transport (-1.8 Sv). Additional AW transport contributions come from WGSW, precipitation less evaporation, sea ice melt, and river and glacial runoff. Some fraction of the fresh, low-density WGSW (0.4 Sv) that enters Baffin Bay through eastern Davis Strait likely remains as upper-layer water and exits Baffin Bay, through Davis Strait, as AW (Fig. 5). However, winter cooling and brine rejection in Baffin Bay may increase WGSW and AW density, causing it to sink and eventually leave Davis Strait as TrW.

Estimates of when outflows from the CAA passages should reach Davis Strait can be made assuming average speeds of 0.15 – 0.2 m s $^{-1}$, in agreement with speeds in Nares Strait, Lancaster Sound, and Baffin Bay (Tang et al. 2004; Münchow and Melling 2008; Peterson et al. 2012). The travel time between Nares and Davis Straits (1900 km) is ~ 4 – 5 months, while that between Lancaster Sound and Davis Strait (1450 km) is ~ 3 – 4 months. With these delays, peak transport through Lancaster Sound and Nares Strait would be expected to arrive in Davis Strait in October–November and May–November, respectively. Annual variability in AW on average tends to have two periods of increased southward transport, October–November and June–July, which correspond well to the estimates from this study.

Transitional Water, the second largest constituent of the net volume transport, is the only other component that exhibits net southward flow. Maximum transport of TrW typically occurs in June–September and minimum transport typically occurs in March–April. In 2004–05, the year of maximum TrW transport (-1.4 Sv), the annual cycle is unlike any of the others observed, with strong peaks in southward transport in January–February and June–July. The southward peak in January–February arises from increased transport in the southward current off west Greenland, but the peak between June and July corresponds to a general increase in southward TrW velocity. Dunlap and Tang (2006) modeled September circulation within Baffin Bay and note the presence of the southward current off west Greenland between $\sim 67.5^\circ$ and 71.1° N. In the model, the northern part of the current, just off the Greenland slope with a core 270 km from Baffin Island, transports -0.65 Sv with increased transport in the upper 400 m and weaker transport below. The current south of the array is located 150–210 km from Baffin Island, is more barotropic, and transports -2.4 Sv. Along the current mooring line, the 2004–09 average September southward transport 130–201 km from Baffin Island (the horizontal distance where the southward current is observed) is -0.8 Sv with stronger velocities in the upper 200 m compared to those at depth. Dunlap and Tang (2006) do not discuss the dynamics that govern the southward current off west Greenland but mention that it is part of a cyclonic gyre on the west Greenland shelf/slope. Average TrW transport roughly equals the sum of that from the inflows into Baffin Bay through eastern Davis Strait (WGIW, WGSW, and unclassified water). The average transport difference between TrW and incoming water is 0.1 Sv and varies between -0.1 and 0.2 Sv over the six years. However, some fraction of WGSW is likely to leave Davis Strait as AW. In addition, some fraction of AW likely leaves Davis Strait as TrW, particularly outflow from Nares Strait that transits through

regions of polynyas in northern Baffin Bay and is more likely to encounter warm, salty WGIW upwelled off the slope of west Greenland.

Annual variability of WGSW was unknown prior to this monitoring program. Quantifying the previously unknown annual range of WGSW transport is important for constraining Davis Strait transports and understanding the mechanisms controlling flow around Greenland. Since the 1950s, hydrographic data have been collected annually along the west Greenland shelf and slope, typically between June and July, by the Danish Meteorological Institute on behalf of the Greenland Institute of Natural Resources (Myers et al. 2007, 2009). Annual sections have been occupied between the southern tip of Greenland at Cape Farewell and near the Davis Strait mooring line at Sisimiut (Fig. 1). Transports of WGSW have been quantified by combining hydrographic sections, climatologies, and models (Myers et al. 2009), while WGIW transports have been estimated using a combination of hydrographic sections and climatologies (Myers et al. 2007). Caution must be used when attempting to identify multiple-year trends from those seasonal snapshots, but when analyzed together with time series from the moored array, the sections can be used to enhance understanding of transport variability along the west Greenland shelf and slope. Results from Myers et al. (2009) show the WGSW and WGIW system weakening as it moves north and water is diverted into the central Labrador Sea. The volume transport of WGSW decreases between Cape Farewell (3.0 ± 0.8 Sv) and Sisimiut (0.0 ± 0.2 Sv) with transport reducing to <0.5 Sv between Fylla Bank (0.8 ± 0.5 Sv) and Maniitsoq (0.2 ± 0.2 Sv). The average June–July WGSW transport during 2004–10 from this monitoring program is larger (0.4 Sv) than reported by Myers et al. (2009). Myers et al. (2007) use a slightly different definition of WGIW ($\theta > 3.5^\circ\text{C}$; $S > 34.88$) when estimating WGIW transport between Cape Farewell (4.9 ± 1.1 Sv) and south of Fylla Bank at Paamiut (0.8 ± 0.7 Sv) during 1995–2005. The average 2004–10 WGIW June–July volume transport from this monitoring program is lower (0.3 Sv).

2) LIQUID FRESHWATER TRANSPORTS

Yearly net liquid freshwater transport through Davis Strait is southward (average = -93 ± 6 mSv) for all years. There is no clear trend (p value > 0.2) and significant interannual variability is present in the monthly transports. Daily net liquid freshwater transport is significantly correlated with volume transport ($r = 0.72$, p value < 0.001) using a Student's t distribution at the 95% (α) confidence level and the integral time scale to determine the minimum degrees of freedom. The minimum net transport in 2007–08 (-66 ± 15 mSv) and

maximum net transport in 2008–09 (-110 ± 14 mSv) reflect variations in AW transport (Fig. 9). Arctic Water freshwater contributes 70% to the total liquid freshwater transport into and out of Davis Strait.

The variability in AW freshwater transport drives most of the variability in the total liquid freshwater transport. The minimum AW freshwater transport in 2007–08 (-73 mSv) is followed by maximum AW transport in 2008–09 (-115 mSv). The maximum annual AW freshwater transport typically occurs in September–November but the timing of minimum transport varies from year to year, ranging among December–January, April–May, and July. Annual liquid freshwater contributions from WGIW and undefined water are small, both ranging between 2 and 5 mSv. Liquid freshwater transport contributions from WGSW and TrW are in opposite directions and roughly equal with average contributions of 17 and -19 mSv. Even though WGSW freshwater transport is small, Rudels (2011) suggests that a decrease in WGSW salinity might lead to a reduction in baroclinic transport through the CAA by altering the density difference between Baffin Bay and the CAA passages.

The CAA passages provide most of the AW freshwater transport. Jones Sound transports the least amount of freshwater through the CAA with an average of -12 ± 3 mSv based on moorings from 1998 to 2002 (Melling et al. 2008). Lancaster Sound has been monitored year-round with moorings since 1998 and contributes approximately the same amount of freshwater as Nares Strait. An average of -32 ± 6 mSv of liquid freshwater transport (between 1998 and 2011) plus an additional 2.1 mSv from sea ice (Peterson et al. 2012) passes through Lancaster Sound. The average 2003–06 liquid freshwater transport through Nares Strait is -28 ± 3 mSv (Münchow and Melling 2008; Rabe et al. 2012). Secondary contributions from Baffin Island runoff, precipitation less evaporation, Greenland Ice Sheet runoff, and CAA sea ice are -46 mSv (Curry et al. 2011). Liquid freshwater transport variability in Davis Strait tracks well with Lancaster Sound freshwater transport, with both pathways experiencing maximums and minimums during the same years between 2004 and 2010 (Peterson et al. 2012). The sum of CAA liquid freshwater inflows plus secondary contributions into Baffin Bay, excluding Baffin Bay sea ice, is slightly more than (-118 mSv), but compares well with, the average liquid freshwater transport through Davis Strait (-93 mSv).

3) SEA ICE TRANSPORTS

Sea ice is generally present across the strait between November and June, with land fast ice sometimes present along the Baffin coast as late as July (Fig. 10a).

FIG. 10. (a) Daily satellite across-strait sea ice coverage estimated on a zonal line situated north of the moored array at $\sim 68^\circ\text{N}$ for 2004–10. Distance is eastward from Baffin Island and only coverage $>15\%$ is shown. (b) Monthly average sea ice area transports (November–May, plotted on the first of the month, negative = southward transport) between 2004 and 2010 using the ice coverage noted in (a). (c) Monthly average (November–May) sea ice volume transports between 2004 and 2010 estimated from ULS ice thickness and satellite-derived ice area transports.

Maximum ice cover typically occurs in March when ice extends across the full width of the strait even though the west Greenland shelf and slope is often ice-free due to warmer water near the surface and the prevailing winds coming off Greenland. The 6-yr average November–May sea ice area exported through Davis Strait between 2004 and 2010 is $-585\,000\text{ km}^2$ (Fig. 10b). Sea ice area transport ranges from $-372\,000\text{ km}^2$ in 2004–05 to $-833\,000\text{ km}^2$ in 2008–09 over the six years with a seasonal uncertainty of $\pm 6500\text{ km}^2$ ($\pm \sigma_{A_{\text{ice}}}$).

Observations of ice thickness were insufficient to establish yearly variations across the array in each month and year (Table 1). Therefore for each month and mooring site, all available data were averaged to estimate the portion of the mean annual cycle for the months of November–May. For years lacking ULS data, the mean annual ice thickness cycle was used to estimate annual ice volume transport. The 6-yr average November–May sea

ice volume transport through Davis Strait between 2004 and 2010 is -407 km^3 . Sea ice volume transport ranges from -262 km^3 in 2004–05 to -574 km^3 in 2008–09 over the six years, with a seasonal uncertainty of $\pm 59\text{ km}^3$ ($\pm \sigma_{\text{FW}_{\text{ice transport}}}$) and no clear trend (Figs. 7 and Figs. 10c). If the annual ice thickness cycle is used for all years, the difference in the average 2004–10 sea ice freshwater transport is 1 km^3 . Including an additional -17 km^3 of sea ice freshwater export between June and July from Jordan and Neu (1982), average 2004–10 sea ice freshwater export through Davis Strait is $-331 \pm 45\text{ km}^3\text{ yr}^{-1}$ or $-10 \pm 1\text{ mSv}$ (Fig. 10c).

5. 2009–10 Arctic Water freshening

Arctic Water began to freshen in September 2009, with an initial drop of salinity of about 0.1 and subsequently, remaining well below average (by about 0.2)

FIG. 11. Daily AW salinity between 2004 and 2010 at central strait C1–C3 100-m moorings.

through August 2010 (Fig. 4). Increased net liquid freshwater outflow is observed in September–November 2009 (Figs 8 and 9), but smaller-than-average southward velocities after November 2009 limit the increase in freshwater transport. Daily salinity records from the Baffin Island slope out to the midstrait, stations C1–C3 at 100 m (Figs. 2 and 11), indicate that freshening began during the annual salinity minimum in September 2009 and extended across to the central strait, station C3. Salinity records on the Baffin shelf also indicate freshening but instrument placement changes and data record lengths make it harder to compare temporal variability (appendix A; Table 1). Freshening is also observed at 200 m for C2 and C3 but not at 250 m at C1. The salinity minimum in 2009 is fresher, lasts longer, and spans a greater distance from Baffin Island than in previous years. The freshening continued through August 2010 at C1 but ended at C2 and C3 in May 2010. Preliminary results from 2010 to 2011 indicate the freshening did not continue and salinity returned to near-average values.

A year-round mooring deployed in the Switchyard region (between Ellesmere Island, Canada, north Greenland, and the North Pole) between April 2008 and May 2009 observed a similar freshening event starting in January 2009 (Jackson et al. 2014). Timmermans et al. (2011) also observed fresher conditions during 2009 in this region using CTD data collected between April and November. The freshening here is limited to 2009–10, with more saline conditions returning in April–May 2010 (Timmermans et al. 2011; Jackson et al. 2014). On the basis of numerical simulations and observations, Timmermans et al. (2011) suggest the observed freshening might be a result of the redistribution of freshwater in the Arctic Ocean forced by changes in the wind-driven circulation. In winter 2009, arctic wind-driven circulation was cyclonic instead of the more typical anticyclonic

regime causing the anticyclonic Beaufort Gyre to weaken and release stored freshwater (Timmermans et al. 2011). Timing of the upstream freshening event at the mooring region is consistent with an advective pathway between the two regions, allowing 11 months for transit through Nares Strait and into Baffin Bay before crossing Davis Strait. Negative salinity anomalies were observed in Lancaster Sound during 2009–10 (Peterson et al. 2012). Like Davis Strait, weaker than average volume transport in 2009–10 might limit the increase in freshwater transport observed in Barrow Strait.

In addition to freshwater changes in the Arctic propagating through Davis Strait, local changes in northwest Greenland and CAA glacier melting might also contribute to the 2009 freshening event as well as to an earlier, freshening signal observed at C1 starting in late 2008. Northwest Greenland and CAA glacier mass loss rates increased sharply between the early 2000s and 2007–09. The summer mass loss of glaciers in northwest Greenland increased from $30.9 \pm 8 \text{ Gt yr}^{-1}$ during 2002–05 to $128.2 \pm 33 \text{ Gt yr}^{-1}$ during 2007–09 (Chen et al. 2011). Similarly, mass loss from CAA glaciers increased sharply from $31 \pm 8 \text{ Gt yr}^{-1}$ between 2004 and 2006 to $92 \pm 12 \text{ Gt yr}^{-1}$ between 2007 and 2009 (Gardner et al. 2011). A rough calculation using mass loss rates (from Chen et al. 2011) and yearly mass loss rates (A. Gardner 2011, personal communication) illustrates an upper bound on the salinity change and the potential contribution of glacier runoff between 2004 and 2010. Assuming that all of the runoff stays and is mixed evenly in the upper 250 m of Baffin Bay (area = $607\,000 \text{ km}^2$), the salinity of glacial runoff is 0, and the same mass loss rates in 2008–09 continue through 2009–10, the estimated runoff from the glaciers is sufficient to reduce the salinity of the upper 250 m of Baffin Bay by 0.07 between 2004–05 and 2006–07 and by 0.11 between 2007–08 and

2009–10. The estimated decreases are greater than the observed decreases in annual-average AW salinity between 2004–05 and 2006–07 (0.04) and less than the observed changes between 2007–08 and 2009–10 (0.16). This suggests that glacial melt is likely a significant contributor to the freshening observed in Davis Strait. However, preliminary results suggest AW salinity returned to near-average values in 2011, which is inconsistent with the idea that increased glacial melt will continue to reduce AW salinity. A more rigorous and careful calculation is needed to better understand the implications of increased freshwater contributions from glaciers into Baffin Bay.

6. Comparison with 1987–90 transports

Data from the 1987 to 1990 moored array are objectively mapped in a similar manner as the current array to facilitate more accurate comparisons between the two periods. Transport comparisons are confined to the central strait due to the absence of shelf measurements in the earlier array. Average 2004–10 transport estimates from the shelves account for volume and liquid freshwater transports of -0.1 Sv and -6 mSv, respectively, from the Baffin shelf and 0.4 Sv and 18 mSv from the west Greenland shelf. Average central strait net volume and liquid freshwater transports for 1987–90 are -3.5 ± 0.6 Sv and -142 ± 23 mSv. Of the three previous studies, Cuny et al. (2005) is the one study that will be compared with the current reanalysis because it is also the only one that includes an estimate of the (unmeasured) velocity shear in the upper 150 m. Cuny et al. (2005) present only northward and southward estimates of the transports and corresponding uncertainties, excluding both the west Greenland and Baffin Island shelves. Net volume and liquid freshwater transports (northward plus southward, with uncertainties summed in quadrature) from Cuny et al. (2005) are -3.4 ± 1.3 Sv and -130 ± 60 mSv, respectively. Estimates from Cuny et al. (2005) are within the transport ranges presented here for 1987–90. The timing of annual variations in transports and water properties agrees well between 1987–90 [both reanalysis and Cuny et al. (2005)] and 2004–10.

While the timing of annual variations agrees well between the two periods, volume and liquid freshwater transports differ significantly. Average central strait volume and liquid freshwater transports for 2004–10, excluding the shelves, are -2.0 ± 0.2 Sv and -105 ± 7 mSv, respectively. Transport differences during 1987–90 and 2004–10 are due to changes in velocity and in the extent of the water mass areas, particularly for AW and WGIW. Average AW volume transport

to the Labrador Sea decreased from -2.2 to -1.7 Sv and average WGIW northward transport into Baffin Bay increased from 0.1 to 0.7 Sv. Similarly, the average areal extent of AW decreased from 33 to 30 km² and the average areal extent of WGIW increased from 24 to 26 km² from 1987–90 to 2004–10. Changes in temperature and salinity between the two periods are insignificant with nearly identical average values for all water masses except a slight decrease in WGIW salinity from an average of 34.67 to 34.59 . Changes between the two periods are driven by changes in volume transport and reflect changes in the mechanisms controlling transport northward and southward through Davis Strait.

It is unclear if more arctic outflow through the CAA occurred during 1987–90, because transport estimates are unavailable before the late 1990s, but variability in Arctic Ocean wind-driven circulation agrees with observed transport changes between the two periods. Annual anticyclonic circulation in the Arctic was much weaker between 1987 and 1990 relative to the current period (Proshutinsky et al. 2009). Stronger anticyclonic circulation concentrates freshwater in the Beaufort Gyre and reduces arctic outflow to the North Atlantic (Proshutinsky and Johnson 1997). The current increase in WGIW inflow is also consistent with changes observed in the strength position of the subpolar gyre, showing recent weakening as compared to the late 1980s (Häkkinen and Rhines 2004). When the subpolar and subtropical gyres weaken, as observed recently compared to the late 1980s, the subpolar gyre contracts and the subpolar front moves westward as the subtropical gyre expands, allowing more high-salinity waters to move northward and enter the Nordic and Labrador Seas (Häkkinen et al. 2011).

A more thorough discussion of the forcing mechanisms and corresponding variability controlling Davis Strait transport is beyond the scope of this paper, and will be treated in another study.

7. Conclusions

Six years of volume and liquid freshwater transports in Davis Strait (2004–10) show significant interannual variability, small annual cycles, and no clear trends with average net transports of -1.6 ± 0.2 Sv and -93 ± 6 mSv, respectively. Annual cycles of net volume and liquid freshwater transports are small because of phase cancellation in the annual cycles of water mass transports. Annual cycles of contributions to total transport by individual water masses are more easily discerned in the data, particularly over the west Greenland shelf and slope. Davis Strait outflow was significantly fresher in

2009–10, likely caused by increased freshwater export from the Beaufort Gyre into Baffin Bay, or increased glacial melt into Baffin Bay, or both. This event was not clearly evident in the net or water mass freshwater transports due to a reduction in southward volume transport and the choice of reference salinity.

A comparison of the 2004–10 results with reanalyzed transports for 1987–90 indicates a 43% decrease of net southward liquid volume transport (from -3.5 to -2.0 Sv, and significantly different) in the central, deep water area of Davis Strait. This is accompanied by a 26% decrease of freshwater transport (from -142 to -105 mSv, and significantly different); during both periods. This change is consistent with changes upstream in the Arctic Ocean, and downstream in the subpolar gyre, as reported elsewhere.

Data from the continental shelves and higher space and time resolution provided by the 2004–10 array result in much narrower confidence limits on the estimated transports, so that variability on annual, interannual, and longer time scales is better resolved. Seaglider sections across the central strait provide important high-resolution knowledge about annual variations in stratification that was unknown prior to this measurement program. Net volume and liquid freshwater transports through Davis Strait are similar in magnitude to those estimated for Fram Strait, the other major pathway connecting the Arctic and North Atlantic. Davis Strait net volume and freshwater transports are within the uncertainty of the Fram Strait estimates [-2.3 ± 4.3 Sv (Rudels et al. 2008; Schauer et al. 2008; Curry et al. 2011) and -120 ± 30 mSv (Kwok et al. 2004; de Steur et al. 2009)]. It is unclear how freshwater export will vary between these two gateways due to changes in the Arctic. Mechanisms driving Davis Strait transports are a complex and poorly understood combination of local Arctic and subarctic interactions. Continued measurements are needed in Davis Strait to isolate interannual variability from long-term trends and to attribute observed changes to driving mechanisms in the Arctic and North Atlantic Oceans.

8. Data distribution

Daily and monthly OA results are available online for download (at iop.apl.washington.edu/data.html), with the time series also available through ACADIS, the repository for U.S. Arctic Observing Network data. Please contact the author if these data are useful to you or used in publications so that we can inform you of changes to the data, document the usefulness of this program, and justify the need to continue monitoring Davis Strait variability.

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REFERENCES

- Aagaard, K., and E. Carmack, 1989: The role of sea ice and other fresh water in the Arctic circulation. *J. Geophys. Res.*, **94** (C10), 14 485–14 498, doi:10.1029/JC094IC10P14485.
- Bacon, S., G. Reverdin, I. Rigor, and H. Snaith, 2002: A freshwater jet on the east Greenland shelf. *J. Geophys. Res.*, **107** (C7), doi:10.1029/2001JC000935.
- Bretherton, F., R. Davis, and C. Fandry, 1976: A technique for objective analysis and design of oceanographic experiments applied to MODE-73. *Deep-Sea Res.*, **23** (7), 559–582, doi:10.1016/0011-7471(76)90001-2.
- Cavalieri, D., T. Markus, and J. Comiso, 2004a: AMSR-E/Aqua daily L3 12.5 km brightness temperature, sea ice concentration, and snow depth polar grids V002 [2004–2010]. National Snow and Ice Data Center. [Available online at http://nsidc.org/data/docs/daac/ae_si12_12km_tb_sea_ice_and_snow.gd.html.]
- , —, and —, 2004b: AMSR-E/Aqua daily L3 6.25 km 89 GHz brightness temperature polar grids V002 [2004–2010]. National Snow and Ice Data Center. [Available online at http://nsidc.org/data/docs/daac/ae_si6_6km_tbs.gd.html.]
- Chen, J., C. Wilson, and B. Tapley, 2011: Interannual variability of Greenland ice losses from satellite gravimetry. *J. Geophys. Res.*, **116**, B07406, doi:10.1029/2010JB007789.
- Cuny, J., P. Rhines, P. Niiler, and S. Bacon, 2002: Labrador Sea Boundary Currents and the Fate of the Irminger Sea Water. *J. Phys. Oceanogr.*, **32**, 627–647, doi:10.1175/1520-0485(2002)032<0627:LSBCAT>2.0.CO;2.
- , —, and R. Kwok, 2005: Davis Strait volume, freshwater and heat fluxes. *Deep-Sea Res. I*, **52** (3), 519–542, doi:10.1016/j.dsr.2004.10.006.
- Curry, B., C. Lee, and B. Petrie, 2011: Volume, freshwater and heat fluxes through Davis Strait, 2004–05. *J. Phys. Oceanogr.*, **41**, 429–436, doi:10.1175/2010JPO4536.1.
- de Jong, M., H. van Aken, K. Våge, and P. Pickart, 2012: Convective mixing in the central Irminger Sea: 2002–2010. *Deep-Sea Res. I*, **63**, 36–51, doi:10.1016/j.dsr.2012.01.003.
- Déry, S., M. Hernández-Hernández, J. Buford, and E. Wood, 2009: Observational evidence of an intensifying hydrological cycle in northern Canada. *Geophys. Res. Lett.*, **36**, L13402, doi:10.1029/2009GL038852.

- de Steur, L., E. Hansen, R. Gerdes, M. Karcher, E. Fahrbach, and J. Holfort, 2009: Freshwater fluxes in the East Greenland Current: A decade of observations. *Geophys. Res. Lett.*, **36**, L23611, doi:10.1029/2009GL041278.
- Dunlap, E., and C. Tang, 2006: Modelling the mean circulation of Baffin Bay. *Atmos.–Ocean*, **44** (1), 99–110, doi:10.3137/ao.440107.
- Eriksen, C., T. Osse, R. Light, T. Wen, T. Lehman, P. Sabin, J. Ballard, and A. Chiodi, 2001: Seaglider: A long-range autonomous underwater vehicle for oceanographic research. *IEEE J. Oceanic Eng.*, **26** (4), 424–436, doi:10.1109/48.972073.
- Gardner, A., and Coauthors, 2011: Sharply increased mass loss from glaciers and ice caps in the Canadian Arctic Archipelago. *Nature*, **473**, 357–360, doi:10.1038/nature10089.
- Goosse, H., T. Fichefet, and J.-M. Campin, 1997: The effects of the water flow through the Canadian Archipelago in a global ice-ocean model. *Geophys. Res. Lett.*, **24**, 1507–1510, doi:10.1029/97GL01352.
- Greene, C., A. Pershing, T. Cronin, and N. Ceci, 2008: Arctic climate change and its impacts on the ecology of the North Atlantic. *Ecology*, **89**, S24–S38, doi:10.1890/07-0550.1.
- Häkkinen, S., and P. Rhines, 2004: Decline of subpolar North Atlantic circulation during the 1990s. *Science*, **304**, 555–559, doi:10.1126/science.1094917.
- , —, and D. Worthen, 2011: Warm and saline events embedded in the meridional circulation of the northern North Atlantic. *J. Geophys. Res.*, **116**, C03006, doi:10.1029/2010JC006275.
- Holland, D., R. Thomas, B. De Young, M. Ribergaard, and B. Lyberth, 2008: Acceleration of Jakobshavn Isbræ triggered by warm subsurface ocean waters. *Nat. Geosci.*, **1**, 659–664, doi:10.1038/ngeo316.
- Holland, M., C. Bitz, M. Eby, and A. Weaver, 2001: The role of ice-ocean interactions in the variability of the North Atlantic thermohaline circulation. *J. Climate*, **14**, 656–675, doi:10.1175/1520-0442(2001)014<0656:TROI>2.0.CO;2.
- Jackson, J., C. Lique, M. Alkire, M. Steele, C. Lee, W. Smethie, and P. Schlosser, 2014: On the waters upstream of Nares Strait, Arctic Ocean, from 1991 to 2012. *Cont. Shelf Res.*, **73**, 83–96.
- Jahn, A., B. Tremblay, L. Mysak, and R. Newton, 2010: Effect of the large-scale atmospheric circulation on the variability of the Arctic Ocean freshwater export. *Climate Dyn.*, **34**, 201–222, doi:10.1007/s00382-009-0558-z.
- Jakobsson, M., R. Macnab, M. Mayer, R. Anderson, M. Edwards, J. Hatzky, H. Schenke, and P. Johnson, 2008: An improved bathymetric portrayal of the Arctic Ocean: Implications for ocean modeling and geological, geophysical and oceanographic analyses. *Geophys. Res. Lett.*, **35**, L07602, doi:10.1029/2008GL033520.
- Jordan, F., and H. Neu, 1982: Ice drift in southern Baffin Bay and Davis Strait. *Atmos.–Ocean*, **20**, 268–275, doi:10.1080/07055900.1982.9649144.
- Kwok, R., 2007: Baffin Bay ice drift and export: 2002–2007. *Geophys. Res. Lett.*, **34**, L19501, doi:10.1029/2007GL031204.
- , and D. Rothrock, 1999: Variability of Fram Strait ice flux and North Atlantic Oscillation. *J. Geophys. Res.*, **104** (C3), 5177–5189, doi:10.1029/1998JC900103.
- , and N. Untersteiner, 2011: The thinning of Arctic sea ice. *Phys. Today*, **64**, 36–41, doi:10.1063/1.3580491.
- , G. Cunningham, and S. Pang, 2004: Fram Strait sea ice outflow. *J. Geophys. Res.*, **109**, C01009, doi:10.1029/2003JC001785.
- Lilly, J., P. Rhines, F. Schott, K. Lavender, J. Lazier, U. Send, and E. D'Asaro, 2003: Observations of the Labrador Sea eddy field. *Prog. Oceanogr.*, **59**, 75–176, doi:10.1016/j.pocan.2003.08.013.
- Melling, H., and Coauthors, 2008: Fresh-water fluxes via Pacific and Arctic outflows across the Canadian polar shelf. *Arctic–Subarctic Ocean Fluxes: Defining the Role of the Northern Seas in Climate*, R. Dickson, J. Meincke, and P. Rhines, Eds., Springer Verlag, 193–247.
- Mote, T., 2007: Greenland surface melt trends 1973–2007: Evidence of a large increase in 2007. *Geophys. Res. Lett.*, **34**, L22507, doi:10.1029/2007GL031976.
- Münchow, A., and H. Melling, 2008: Ocean current observations from Nares Strait to the west of Greenland: Interannual to tidal variability and forcing. *J. Mar. Res.*, **66**, 801–833, doi:10.1357/002224008788064612.
- Myers, P., 2005: Impact of freshwater from the Canadian Arctic Archipelago on Labrador Sea Water formation. *Geophys. Res. Lett.*, **32**, L06605, doi:10.1029/2004GL022082.
- , N. Kulan, and M. Ribergaard, 2007: Irminger Water variability in the West Greenland Current. *Geophys. Res. Lett.*, **34**, L17601, doi:10.1029/2007GL030419.
- , C. Donnelly, and M. Ribergaard, 2009: Structure and variability of the West Greenland Current in summer derived from 6 repeat standard sections. *Prog. Oceanogr.*, **80**, 93–112, doi:10.1016/j.pocan.2008.12.003.
- Overland, J., M. Wang, and S. Salo, 2008: The recent Arctic warm period. *Tellus*, **60**, 589–597, doi:10.1111/j.1600-0870.2008.00327.x.
- Perkin, R., and E. Lewis, 1980: The Practical Salinity Scale 1978: Fitting the data. *IEEE J. Oceanic Eng.*, **5**, 9–16, doi:10.1109/JOE.1980.1145441.
- Peterson, I., J. Hamilton, S. Prinsenber, and R. Pettipas, 2012: Wind-forcing of volume transport through Lancaster Sound. *J. Geophys. Res.*, **117**, C11018, doi:10.1029/2012JC008140.
- Prater, M., 2002: Eddies in the Labrador Sea as observed by profiling RAFOS floats and remote sensing. *J. Phys. Oceanogr.*, **32**, 411–427, doi:10.1175/1520-0485(2002)032<0411:EITLSA>2.0.CO;2.
- Proshutinsky, A., and M. Johnson, 1997: Two circulation regimes of the wind-driven Arctic Ocean. *J. Geophys. Res.*, **102** (C6), 12 493–12 514, doi:10.1029/97JC00738.
- , and —, 2010: Decadal variability of Arctic climate: Cyclonic and anticyclonic circulation regimes. *2010 Fall Meeting*, San Francisco, CA, Amer. Geophys. Union, Abstract GC12B-09.
- , and Coauthors, 2009: Beaufort Gyre freshwater reservoir: State and variability from observations. *J. Geophys. Res.*, **114**, C00A10, doi:10.1029/2008JC005104.
- Rabe, B., H. Johnson, A. Münchow, and H. Melling, 2012: Geostrophic ocean currents and freshwater fluxes across the Canadian Polar Shelf via Nares Strait. *J. Mar. Res.*, **70**, 603–640.
- Rignot, E., M. Koppes, and I. Velicogna, 2010: Rapid submarine melting of the calving faces of west Greenland glaciers. *Nat. Geosci.*, **3**, 187–191, doi:10.1038/ngeo765.
- Ross, C., 1992: Moored current meter measurements across the Davis Strait. Northwest Atlantic Fisheries Organization, NAFO SCR Doc. 92/70, 8 pp.
- Rudels, B., 2011: Volume and freshwater transports through the Canadian Arctic Archipelago–Baffin Bay system. *J. Geophys. Res.*, **116**, C00D10, doi:10.1029/2011JC007019.
- , M. Marnela, and P. Eriksson, 2008: Constraints on estimating mass, heat and freshwater transports in the Arctic Ocean: An exercise. *Arctic–Subarctic Ocean Fluxes: Defining the Role of the Northern Seas in Climate*, R. Dickson, J. Meincke, and P. Rhines, Eds., Springer Verlag, 315–341.

- Schauer, U., A. Beszczynska-Möller, W. Walczowski, E. Fahrbach, J. Piechura, and E. Hansen, 2008: Variation of measured heat flow through the Fram Strait between 1997 and 2006. *Arctic–Subarctic Ocean Fluxes: Defining the Role of the Northern Seas in Climate*, R. Dickson, J. Meincke, and P. Rhines, Eds., Springer Verlag, 65–85.
- Schmidt, S., and U. Send, 2007: Origin and composition of seasonal Labrador Sea freshwater. *J. Phys. Oceanogr.*, **37**, 1445–1454, doi:10.1175/JPO3065.1.
- Spall, M., and R. Pickart, 2003: Wind-driven recirculations and exchange in the Labrador and Irminger Seas. *J. Phys. Oceanogr.*, **33**, 1829–1845, doi:10.1175/2384.1.
- Straneo, F., and F. Saucier, 2008: The Arctic–subarctic exchange through Hudson Strait. *Arctic–Subarctic Ocean Fluxes: Defining the Role of the Northern Seas in Climate*, R. Dickson, J. Meincke, and P. Rhines, Eds., Springer Verlag, 249–261.
- , G. Hamilton, D. Sutherland, L. Stearns, F. Davidson, M. Hammill, G. Stenson, and A. Rosing Asvid, 2010: Rapid circulation of warm subtropical waters in a major glacial fjord off east Greenland. *Nat. Geosci.*, **3**, 182–186, doi:10.1038/ngeo764.
- Stroeve, J., M. Serreze, M. Holland, J. Kay, J. Malanik, and A. Barrett, 2012: The Arctic’s rapidly shrinking sea ice cover: A research synthesis. *Climatic Change*, **110**, 1005–1027, doi:10.1007/s10584-011-0101-1.
- Sutherland, D., and R. Pickart, 2008: The East Greenland Coastal Current: Structure, variability, and forcing. *Prog. Oceanogr.*, **78**, 58–77, doi:10.1016/j.pocean.2007.09.006.
- Tang, C., C. Ross, T. Yao, B. Petrie, B. DeTracey, and E. Dunlap, 2004: The circulation, water masses and sea-ice of Baffin Bay. *Prog. Oceanogr.*, **63**, 183–228, doi:10.1016/j.pocean.2004.09.005.
- Timmermans, M., A. Proshutinsky, R. Krishfield, D. Perovich, J. Richter-Menge, T. Stanton, and J. Toole, 2011: Surface freshening in the Arctic Ocean’s Eurasian Basin: An apparent consequence of recent change in the wind-driven circulation. *J. Geophys. Res.*, **116**, C00D03, doi:10.1029/2011JC006975.
- Våge, K., and Coauthors, 2009: Surprising return of deep convection to the subpolar North Atlantic Ocean in winter 2007–2008. *Nat. Geosci.*, **2**, 67–72, doi:10.1038/ngeo382.