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# A Synthesis of Exchanges Through the Main Oceanic Gateways to the Arctic Ocean

BY AGNIESZKA BESZCZYNSKA-MÖLLER,  
REBECCA A. WOODGATE,  
CRAIG LEE, HUMFREY MELLING,  
AND MICHAEL KARCHER

Deployment of a deep mooring  
with an Aanderaa current  
meter in Fram Strait during  
R/V *Polarstern* cruise ARKXIII  
in 2008. Photo courtesy of  
A. Beszczynska-Möller

**ABSTRACT.** In recent decades, the Arctic Ocean has changed dramatically. Exchanges through the main oceanic gateways indicate two main processes of global climatic importance—poleward oceanic heat flux into the Arctic Ocean and export of freshwater toward the North Atlantic. Since the 1990s, in particular during the International Polar Year (2007–2009), extensive observational efforts were undertaken to monitor volume, heat, and freshwater fluxes between the Arctic Ocean and the subpolar seas on scales from daily to multiyear. This paper reviews present-day estimates of oceanic fluxes and reports on technological advances and existing challenges in measuring exchanges through the main oceanic gateways to the Arctic.

## INTRODUCTION

The Arctic Ocean is tightly connected to the global ocean system via water mass exchanges with the Pacific and Atlantic Oceans through several main oceanic gateways: Bering Strait, the Canadian Arctic Archipelago, Davis Strait, Fram Strait, and the entrance to the Barents Sea (Figure 1). Changes to subarctic seas thus can affect the Arctic Ocean. For example, changing poleward oceanic heat flux can potentially impact perennial sea ice. In recent decades, a decline in sea ice coverage, thickness, and volume has been observed (e.g., Kwok et al., 2009), with the largest sea ice retreat in summer 2007 (e.g., Stroeve et al., 2008). A large-scale shift to warmer conditions in the Arctic has been observed since the early 1990s, including warmer Atlantic inflow (Holliday et al., 2008; Schauer et al., 2008); sporadic warming of the Pacific Water inflow, especially in 2007 (Woodgate et al., 2010); intrusions of anomalously warm water over Arctic shelves (Dmitrenko et al., 2010), along the Arctic continental margin (Dmitrenko et al., 2008), into the central Arctic (Polyakov et al., 2005), and in the Canada Basin (McLaughlin et al., 2009); and a large increase of ocean-atmosphere heat fluxes due to sea ice losses (Kurtz et al., 2011). In addition, heat carried

by Pacific inflow through Bering Strait appears to play an important, triggering role in recent sea ice decline in the western Arctic (Woodgate et al., 2010).

Climate simulations indicate that increased freshwater outflow from the Arctic may reduce convection in the deepwater formation regions of the North Atlantic, potentially influencing the Atlantic meridional overturning circulation (MOC; Wadley and Bigg, 2002; Vellinga et al., 2008). Arctic freshwater, stored in the upper few hundred meters, is delivered to the North Atlantic in the form of liquid freshwater or sea ice through Fram Strait or the Canadian Arctic Archipelago. Details of how this flux influences MOC are still poorly known, although it is believed the impacts on convection and Atlantic deepwater formation depend on the exit route of the water from the Arctic (Koenigk et al., 2007). In recent years, substantial changes have been observed in the storage and distribution of freshwater in the Arctic Ocean, such as a shift in the position of the transpolar drift (Rigor et al., 2002), changes in Arctic-wide freshwater content (e.g., Polyakov et al., 2008; Rabe et al., 2011), and trajectory shifts in the Arctic-subarctic freshwater cycle (Peterson et al., 2006).

In this paper, we review present-day observational efforts to quantify fluxes of

mass, liquid freshwater, and heat through the main Arctic gateways and the estimates derived from these observations. Year-round moorings, almost uninterrupted since 1990, measure Pacific inflow through Bering Strait. Since 2004, the program has been part of the US National Oceanic and Atmospheric Administration (NOAA)-led Russian-American Long-Term Census of the Arctic (RUSALCA) program. Since 2007, a high-resolution mooring array has been deployed in the strait as part of a US National Science Foundation (NSF) International Polar Year (IPY) effort, and since 2010, as part of the NSF Arctic Observing Network program.

In the Atlantic inflow sector, the first concerted but regional effort to monitor oceanic fluxes through the Nordic Seas started in the VEINS project (Variability of Exchanges in Northern Seas, 1997–2000) and was followed by the international pan-Arctic ASOF study (Arctic-Subarctic Ocean Fluxes, 2003–2006). The subsequent DAMOCLES project (Developing Arctic Modelling and Observing Capabilities for Long-term Environment Studies, 2006–2010) aimed to understand the effects of climate change in the Arctic (in particular, its potential to reduce perennial sea ice). DAMOCLES was the European contribution to IPY and provided the largest ever effort to assemble simultaneous observations of the full Arctic atmosphere-ice-ocean system, including oceanic exchanges through the Arctic-Atlantic gateways.

To the west of Greenland, measurements in Davis Strait carried out as part of the NSF Freshwater Initiative (2004–2007) were followed by intensive observational efforts undertaken



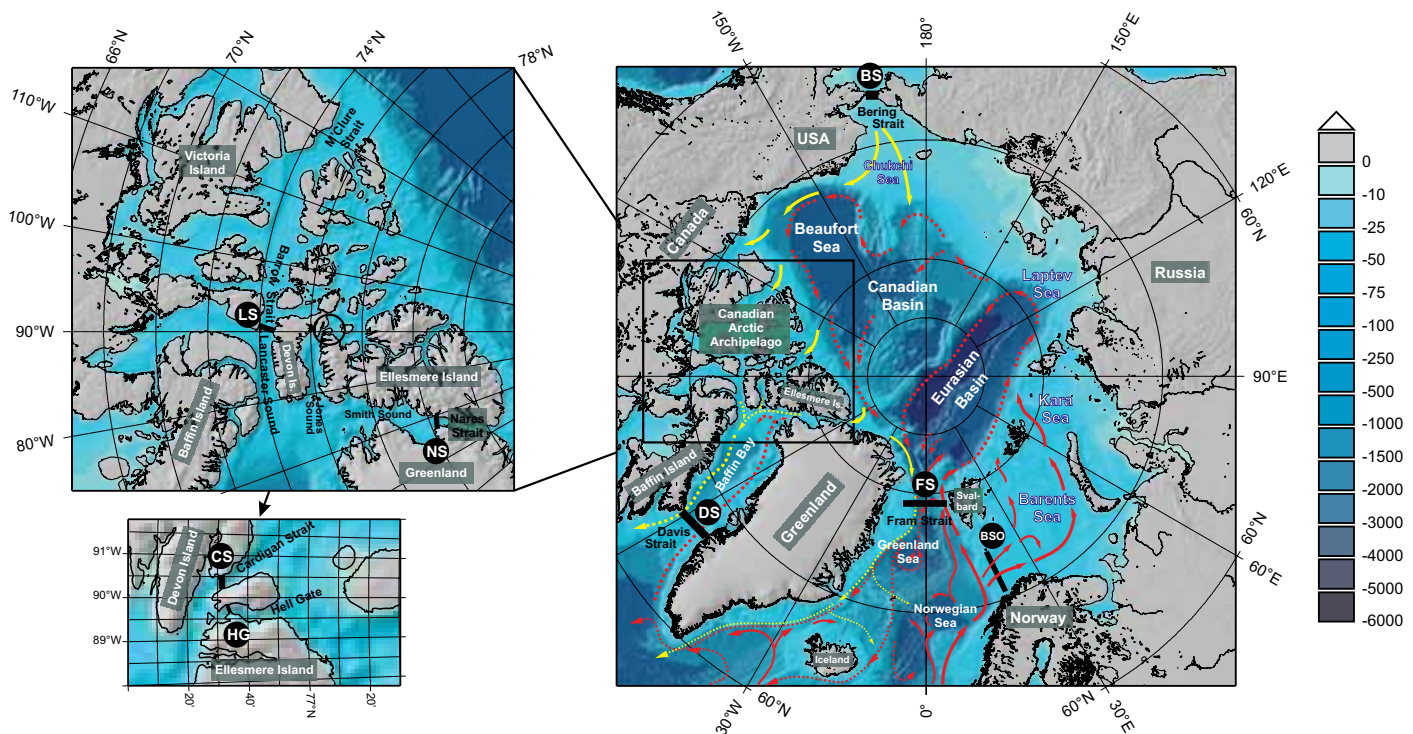


Figure 1. Locations of moored arrays in the main Arctic oceanic gateways, overlaid on a bathymetry map (in color; bathymetry data from the International Bathymetric Chart of the Arctic Ocean [http://www.ngdc.noaa.gov/mgg/bathymetry/arctic; Jakobsson et al., 2003]). A schematic of the main pathways of the Atlantic (red) and Pacific (yellow) derived waters is shown (after Jones et al., 2001; Rudels et al., 2004; Aksenov et al., 2010). The solid red line depicts the Atlantic Water (AW) inflow, and the dashed red line represents circulation of the modified AW in the Arctic Ocean. The solid yellow line shows the Pacific inflow, and the dashed yellow line indicates the freshwater outflow (a mixture of Pacific-, river-, and ice-melt-origin waters) from the Arctic Ocean. Black bars represent the moored arrays: FS = Fram Strait. DS = Davis Strait. BSO = Barents Sea Opening. BS = Bering Strait. NS = Nares Strait. LS = Lancaster Sound. CS = Cardigan Strait. HG = Hell Gate. A magnified map of the Canadian Arctic Archipelago is shown on the upper left panel, and details of Cardigan Strait and Hell Gate are zoomed on the lower left panel.

from 2007 to present as part of the NSF IPY and Arctic Observing Network programs. Monitoring of oceanic fluxes through three main gateways of the Canadian Archipelago from 1998–2006 took place as part of the international ASOF program and the NSF Freshwater Initiative, and was continued during IPY under the Canadian Archipelago Throughflow Study (CATS).

### BARENTS SEA OPENING AND BARENTS SEA—STRONGEST MODIFICATIONS OF THE ATLANTIC INFLOW

The Atlantic inflow across the Greenland-Scotland Ridge brings 8.5 Sv ( $1 \text{ Sv} = 10^6 \text{ m}^3 \text{ s}^{-1}$ ) of warm, saline waters into the Nordic Seas (Østerhus et al., 2005). More than half of those waters

continue northward as two separate branches of the Norwegian Atlantic Current (NwAC). The western baroclinic (i.e., varying with depth) flow follows the Arctic Front through the Norwegian and Greenland Seas and ultimately converges with the West Spitsbergen Current (WSC) in Fram Strait (Orvik and Niiler, 2002; Figure 1). The eastern barotropic (i.e., similar at all depths) branch, the Norwegian Atlantic Slope Current (NwASC), is topographically trapped over the Norwegian continental slope and provides the main source of Atlantic Water (AW) and heat to the Barents Sea (Orvik and Skagseth, 2005). On the western slope of the Barents Sea, part of AW enters a 400 km wide passage between Bear Island and the North Cape (known as the Barents Sea Opening,

BSO) as the North Cape Current. Only the upper part of the AW layer can travel through this 450 m deep passage into the Barents Sea. The remaining AW flow continues along the shelf slope into Fram Strait.

From long-term measurements at the BSO moored array, the North Cape Current was found to depend on local wind forcing (Ingvaldsen et al., 2004a), and the AW inflow exhibited strong seasonal variability with a maximum in winter (Ingvaldsen et al., 2004b). The mean inflow of warm, salty AW through BSO was estimated by Skagseth et al. (2008) as 1.8 Sv, with large interannual variations (between 0.8 and 2.9 Sv) and updated to 2 Sv for 1997–2007 by Smedsrud et al. (2010). However, a fraction of the Atlantic inflow recirculates

along a short pathway after passing through BSO and returns as a 2–2.5°C colder outflow (leaving at 2°C) through the northern part of the opening (Rudels, 1987; Skagseth et al., 2008). An additional inflow enters the Barents Sea along the Norwegian coast as the warm, fresh Norwegian Coastal Current (NCC), carrying a mixture of Atlantic and coastal waters. In a recent study, Skagseth et al. (2011) estimated the NCC volume transport as 2.6 Sv (half of it confined to the slope) with 86 coastal-water components being 1.8 Sv of that (including 0.6 Sv in the slope branch).

While no significant trend was found in NwASC volume flux, Mork and Skagseth (2010) reported a positive trend in the temperature, corresponding to a linear increase of 0.5°C in 1992–2009. In BSO, a temperature increase of 1°C over the period 1997–2006 (to values above 6°C), reported by Skagseth et al. (2008), was accompanied by a positive trend in volume flux of 0.1 Sv yr<sup>-1</sup>.

AW circulation in the Barents Sea consists of multiple narrow currents following bottom topography. During its transit through the shallow (mean depth of 230 m) sea, AW density increases significantly due to atmospheric cooling and ice formation, despite dilution by precipitation, sea ice melting, and an admixture of fresh surface and coastal waters. Warm AW inflow keeps large areas of the Barents Sea ice-free year-round, and, as a result, nearly all oceanic heat entering through BSO is lost to the atmosphere. Smedsrud et al. (2010) showed that as much as 67 TW (92% of the heat input of 73 TW through BSO in the total inflow of 3.2 Sv; 1 TW = 10<sup>12</sup> W) is lost to the atmosphere already in the southern

Barents Sea and ~ 4 TW in its northern part. Most of the AW leaving the Barents Sea into the Arctic Ocean via St. Anna Trough is already cooled to temperatures below 0°C (Schauer et al., 2002).

In contrast to extensive observations in the western Barents Sea, only a few long-term measurements and sporadically repeated hydrographic sections are available in the eastern and northern Barents Sea. Using the only data available (i.e., from a one-year moored array between Novaya Zemlaya and Franz Josef Land in 1991–1992), Schauer et al. (2002) estimated a net volume flux into the Arctic Ocean of 1.5 Sv, half of it consisting of dense, cold (temperature of about –0.5°C) bottom waters. Recently, Gammelsrød et al. (2009) combined the same observations with two numerical models and updated the earlier estimate to 2 ± 0.6 Sv. A modeling data assimilation study (using a variational inverse) of monthly hydrographic and atmospheric climatologies (Panteleev et al., 2006) provided estimates of BSO inflow of 3.2 Sv, recirculation in the northern BSO of 1.5 Sv, and the outflows between Novaya Zemlya and Franz Jozef Land and through Kara Gate (the passage between Novaya Zemla and the Russian coast) of 1.1 and 0.7 Sv, respectively, similar to available observations. Recent observational activities in the Barents

Sea exit took place under the Norwegian IPY Project BIAC (Bipolar Atlantic Thermohaline Circulation), when five moorings were deployed for one year (2007–2008) between Novaya Zemlya and Franz Jozef Land.

## FRAM STRAIT— A DEEP AND WIDE PASSAGE FOR EXCHANGE OF ATLANTIC AND ARCTIC ORIGIN WATERS

Fram Strait provides a wide, deep (sill depth of 2600 m) conduit both for Atlantic Water inflow into the Arctic Ocean and export of polar waters and sea ice out of the Arctic and into the Nordic Seas (Figure 1). While AW inflow is concentrated in the West Spitsbergen Current (WSC) above the eastern shelf slope, the outflow (both liquid and sea ice) from the Arctic Ocean is carried in the western part of the strait, both by the East Greenland Current (EGC) and by waters above the Greenland shelf. In comparison to the strong atmospheric modification of the AW branch crossing the shallow Barents Sea, AW in Fram Strait is spread over deeper layers, and thus preserves its warm core, losing less heat to the atmosphere.

After the detour of part of the Atlantic inflow into BSO, AW continues northward along the continental shelf slope as the West Spitsbergen Current

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**Agnieszka Beszczynska-Möller** ([agnieszka.beszczynska-moeller@awi.de](mailto:agnieszka.beszczynska-moeller@awi.de)) is Physical Oceanographer, Alfred Wegener Institute for Polar and Marine Research, Bremerhaven, Germany. **Rebecca A. Woodgate** is Associate Professor, Applied Physics Laboratory, University of Washington, Seattle, WA, USA. **Craig Lee** is Associate Professor, Applied Physics Laboratory, University of Washington, Seattle, WA, USA. **Humfrey Melling** is Senior Researcher, Fisheries and Oceans Canada, Institute of Ocean Sciences, Sidney, BC, Canada. **Michael Karcher** is Physical Oceanographer, Alfred Wegener Institute for Polar and Marine Research, Bremerhaven, Germany, and also a researcher at Ocean Atmosphere Systems, Hamburg, Germany.

(WSC). Farther to the north, WSC splits into three separate branches (see Schauer et al., 2004, for discussion). The easternmost branch follows the upper shelf slope north of Svalbard (Svalbard branch) and constitutes the main source of AW to the Arctic Ocean Boundary Current, the topographically steered pan-Arctic circulation of AW (Woodgate et al., 2001). The second branch circumvents the northwestern rim of the Yermak Plateau (Yermak Plateau branch), continues eastward, and ultimately rejoins the Svalbard Branch, thus also entering the Arctic. The westernmost branch of WSC mostly carries waters from the western NwAC extension and recirculates directly in the northern Fram Strait, and exits again to the south.

Transport and properties of AW carried by WSC west and north of Svalbard have been extensively studied since the beginning of the previous century. Before the mid 1990s, most of the data was obtained by summer hydrographic surveys or sparse moorings that did not extend across the entire strait (Aagaard et al., 1973; Foldvik et al., 1988). Since 1997, long-term year-round observations have been carried out through an extensive moored array in northern Fram Strait (Schauer et al., 2004, 2008). The joint German-Norwegian array of 16 deep moorings extends from the shelf edge west of Svalbard to the east Greenland shelf slope with a few shallow moorings on the western shelf dedicated to monitoring freshwater fluxes in EGC (Figure 2). The moored array covers the 300 km wide section with a spatial resolution from ~ 10 km at the upper shelf slope to ~ 30 km in the deep area. The array is instrumented with current meters and conductivity-temperature-depth

(CTD) sensors distributed at five levels: in the subsurface layer (~ 50 m), in the AW layer (~ 250 m), at the AW lower boundary (~ 750 m, since 2002), in the deep layer (~ 1500 m), and above the bottom.

The 13 years of measurements provide estimates of the magnitude and variability of oceanic fluxes through Fram Strait over a wide range of time scales (from hourly to multiyear), and suggest a mean net volume flux to the south of  $2 \pm 2.7$  Sv (Schauer et al., 2008). There are large uncertainties in estimates of volume flux through the entire strait because the spatial variability of the flow with a Rossby radius of ~ 10 km is only poorly resolved by the mooring spacing in the deep part of the strait. (A Rossby radius is a theoretically derived scale typical of oceanic horizontal variability; it is determined from the ocean depth and the density structure.) AW inflow in WSC, where mooring coverage is denser, is better constrained. On interannual time scales between 1997 and 2010, a nearly constant volume flux in the WSC Svalbard branch (long-term mean of  $1.8 \pm 0.1$  Sv northward, including  $1.3 \pm 0.1$  Sv of AW warmer than 2°C and showing no seasonal variability) was accompanied by a highly variable volume flux of between 2 and 6 Sv in the Yermak Plateau branch (long-term mean of  $5 \pm 0.4$  Sv, strong seasonal variability, 1–2 Sv of warm AW).

Significant AW warming has been observed at the moored array in Fram Strait since year-round measurements began. Schauer et al. (2008) report distinctive temperature increases in two periods (1998–2000 and 2003–2006), with a net increase over the last decade of 1°C. Using a stream-tube approach to estimate the oceanic heat flux through

Fram Strait, Schauer and Beszczynska-Möller (2009) obtained a mean value of  $36 \pm 6$  TW for the 1997–2009 period. In recent years, substantial warming was also observed in the AW recirculating directly in Fram Strait. This warm water, joining the southward flow on the western side of the strait, may potentially modify water masses leaving the Arctic Ocean toward the North Atlantic.

The East Greenland Current in western Fram Strait carries fresh, cold Arctic waters and sea ice into the Nordic seas and toward the North Atlantic. In the Arctic freshwater budget studied by Serreze et al. (2006), roughly equal contributions from Fram Strait sea ice and liquid freshwater fluxes deliver about half of the freshwater exported from the Arctic Ocean. While the contributions of Pacific, meteoric (precipitation and riverine), and ice meltwaters to the EGC freshwater outflow vary significantly between years (e.g., Dodd et al., 2009; Rabe et al., 2009) with an alternative exit route for these waters being via the Canadian Arctic Archipelago, nearly the entire sea ice flux from the Arctic passes through Fram Strait (Kwok, 2009). Long-term, year-round measurements of EGC freshwater flux (freshwater being a relative measure, referenced to an arbitrary chosen salinity) have been carried out since 1997 at the Fram Strait moored array. The preliminary estimates of Holfort et al. (2008) of 32 mSv of freshwater transport in EGC and an equal amount over the East Greenland shelf for the period 1997–2005 were updated by de Steur et al. (2009), using data from 1997–2008. Both studies yield similar liquid freshwater flux through Fram Strait (64 and 66 mSv), but the de Steur et al. (2009) study yields a higher estimate of freshwater transport

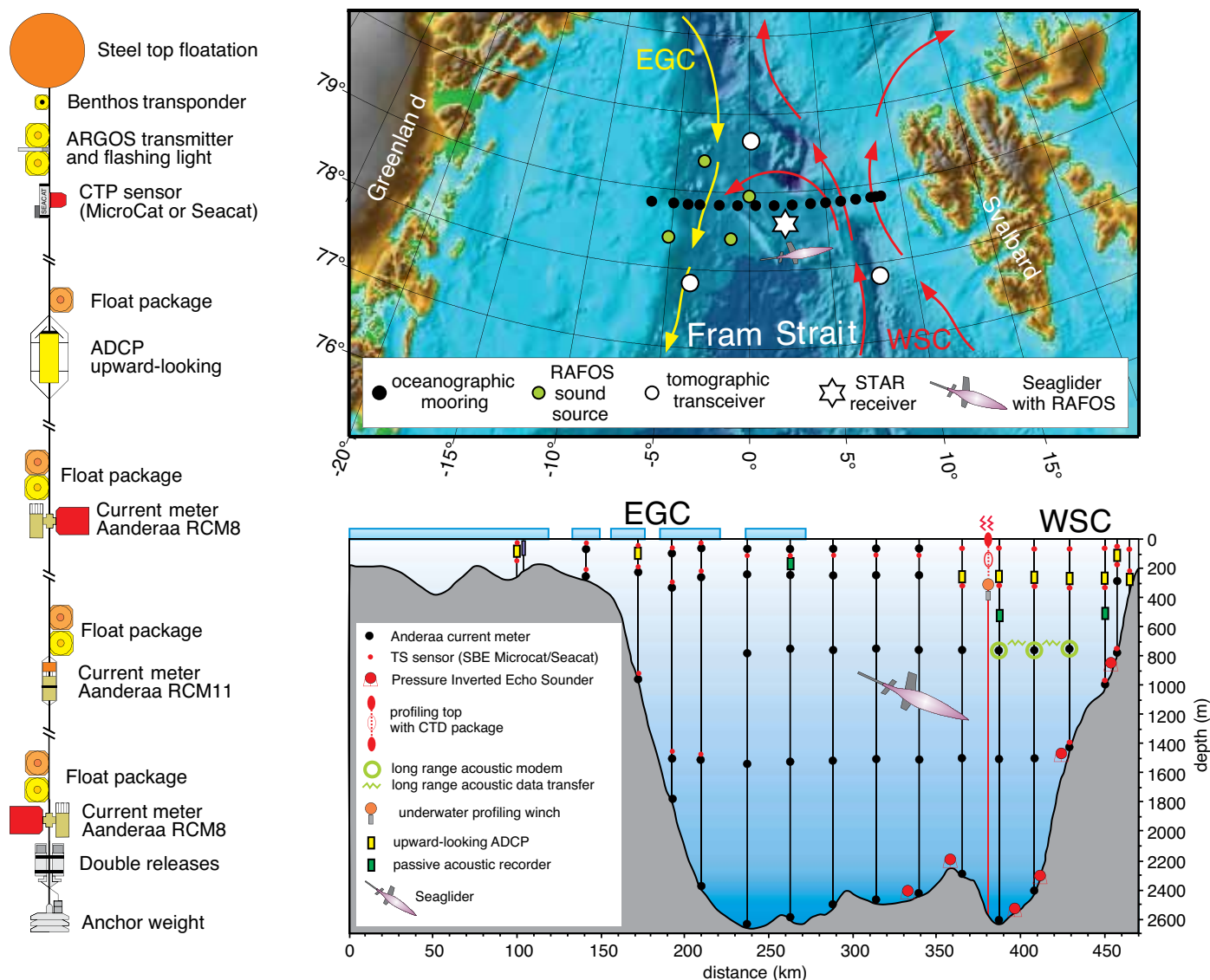


Figure 2. Schematic of the Fram Strait Observatory. An example of a deep mooring is shown in the left panel: CTP = conductivity, temperature, pressure sensor. ADCP = acoustic Doppler current profiler. RCM = recording current meter. The upper right panel shows positions of the oceanographic moorings, tomography moorings (STAR = Simple Tomographic Acoustic Receiver), and RAFOS (SOund Fixing And Ranging, spelled backward) sound sources in Fram Strait. The lower right panel presents a distribution of instruments in the array (the view looking north). EGC = East Greenland Current. WSC = West Spitsbergen Current.

in EGC (with a long-term annual mean of 40.4 mSv in 1997–2008) and a lower contribution from the shelf of 25.6 mSv. The same study also shows that the total EGC volume flux more than doubled since 2001 (from a minimum of 4 Sv in 2002 to a maximum of nearly 10 Sv in 2007), mostly due to stronger transport in deep layers (below 700 m). In the upper layer above 700 m, the volume transport increased only slightly, and

the freshwater flux remained nearly constant. Rabe et al. (2009) estimated the average liquid freshwater transport in summer (in the EGC and over the shelf) to be a net 80 mSv, based on hydrographic data combined with current meter measurements. They also derived the composition of freshwater flux from salinity and  $\delta^{18}\text{O}$  (oxygen isotope) measurements. Meteoric freshwater contributed 130–160 mSv to the

total flux, while an increase of salinity in the upper layer due to sea ice formation was equivalent to a decrease of the total freshwater flux by 60–80 mSv.

### BERING STRAIT— THE ONLY ENTRANCE FOR PACIFIC WATERS

In contrast to the broad entrance points for Atlantic Water into the Arctic, Pacific Water has only one oceanic entrance, via



the shallow (~ 50 m), narrow (~ 85 km wide) Bering Strait, which is divided into two channels by the Diomed Islands and the US-Russian border (Figure 1). Sporadic observations in the strait have been made for centuries (see Coachman and Aagaard, 1966, for a review). Year-round moorings have been deployed in Bering Strait almost continuously since 1990 (Roach et al., 1995; Woodgate et al., 2006), with a one-year gap (summer 1996–1997).

Typically, moorings have been deployed in the center of each of the channels (other than in several years in the 1990s when access was limited to US waters) and at a site ~ 55 km north of the strait proper, hypothesized to give a useful average of the mean flow properties (Woodgate et al., 2007). Due to the dangers of trawling by ice keels, these moorings typically only measured up to ~ 10 m above bottom (i.e., ~ 40 m below the surface). Since 2001, an extra mooring has been added to measure a seasonal, warm, fresh coastal current on the US side of the strait, the Alaskan Coastal Current. Since 2007, an array consisting of eight moorings, including upper-layer sensors (Iscats, also called ICECATs, described below in the new technology section), has been deployed in the strait to quantify cross-strait structure and elucidate the physical driving mechanisms of the flow (Woodgate, 2008).

Although only ~ 0.8 Sv in the long-term annual mean (Roach et al., 1995), the Bering Strait throughflow is a large oceanic source of nutrients for the Arctic (Walsh et al., 1989), bringing in one-third of the freshwater entering the Arctic Ocean (Serreze et al., 2006). Being generally fresher, Pacific waters lie higher in the water column than Atlantic

waters, and provide part of the strong upper-ocean stratification that typifies the Arctic and (in winter) isolates Arctic sea ice from subsurface oceanic heat. In summer, however, Pacific waters carry significant amounts of heat into the Arctic—although highly variable from year to year, the heat flux (relative to the freezing temperature of seawater) is enough to melt 1–2 million km<sup>2</sup> of 1 m thick ice (Woodgate et al., 2010). Certainly, the pathways of Pacific Water into the Arctic are clearly reflected in the structure of the sea ice edge, implying that Pacific Water heat acts as a trigger for the onset of western Arctic sea ice melt, especially in 2007 when the Bering Strait oceanic heat flux was over twice that in 2001 (Woodgate et al., 2010).

The above estimates are based on near-bottom data, using a simple approach (based on satellite data and summer hydrographic stations) to include the substantial contributions from the Alaskan Coastal Current and stratification in the upper water column. While only ~ 0.1 Sv in volume, the Alaskan Coastal Current is believed to contribute one-third of the heat and one-quarter of the freshwater fluxes through the strait (Woodgate and Aagaard, 2005; Woodgate et al., 2006). The ~ 0.1 Sv contribution of another coastal current, the Siberian Coastal Current (present only in some summers on the Russian side of the strait and entering Bering Strait from the north) is still mostly unquantified (Woodgate et al., 2005b).

The denser array of moorings (seven across the strait proper, with velocity profiles and upper layer sensors at five of the moorings) also allows the first year-round quantitative description of the flow structure in the strait. These new data show the flow to be strongly

coherent across the section, confirming older ideas of a dominantly barotropic and homogeneous flow throughout the strait except in the boundary currents, a hypothesis developed from analysis of prior mooring and hydrographic data and assumptions about the large-scale forcings of the flow (see Woodgate et al., 2005b, for discussion). This coherence in velocity is in stark contrast to the lack of coherence across the much wider Fram Strait, reflecting the comparative smallness of the strait (~ 85 km versus ~ 350 km for Fram Strait) and the fact that the main Bering Strait forcings have length scales larger than the strait. Specifically, the flow through Bering Strait is generally attributed to a northward pressure head between the Pacific and the Arctic, modulated by local winds (see Woodgate et al., 2005b, for discussion). While it has been noted that the estimated size of the pressure-head forcing is comparable to the difference in steric heights between the basins (i.e., differences arising due to different densities of water in the basins; e.g., Stigebrandt, 1984), other causes of the pressure head are also being considered. Recent results indicate significant temporal variability in this term, suggesting that both far-field and local forcings determine the net flow through the strait (Woodgate et al., 2005b, 2010; Aagaard et al., 2006).

## CANADIAN POLAR SHELF— A NETWORK OF CHANNELS

The other half of the freshwater outflow from the Arctic Ocean takes routes west of Greenland across the Canadian polar shelf, a patchwork of islands and straits, basins, and sills (Figure 1). All the Canadian Arctic Throughflow (CAT) passes through four gateways, with Nares



Strait, Lancaster Sound, and Cardigan Strait (plus Hell Gate) being most important (Melling et al., 2008).

Pathways across the Canadian shelf are deeper than those for Pacific Arctic inflow across the ~ 50 m deep Bering and Chukchi shelves; the depths of obstructing sills vary, with the shallowest being 80 m (Penny Strait) and the deepest 220 m (Nares Strait). The gateways are hydrodynamically wide (two to 10 times greater than the Rossby radius), despite being geometrically narrow (15–60 km). This characteristic permits currents in both directions simultaneously, with Arctic outflow favoring the surface on the right (looking southeast) and the counterflow at greater depth on the left. Each one of these gateways resembles Fram Strait in miniature, with similarly complicated recirculation and mixing.

The Canadian shelf is the place where a tidal signal arriving from the west, weakened during travel around the Arctic, meets a stronger tidal wave arriving directly from the Atlantic via Baffin Bay (Kowalik and Proshutinsky, 1994). Disparity in the amplitudes and phases of the waves creates strong tidal currents within the gateways, strong turbulence, and enhanced frictional damping of CAT. Also, the high terrain of the Arctic islands creates preferred pathways for airflow, which accelerates within the gateways, resulting in very strong along-strait winds at times (Samelson et al., 2006).

The measurement of CAT faces challenges: remoteness, heavy sea ice that becomes land-fast in winter, strong currents, and an unreliable geomagnetic reference direction because it occurs close to the north magnetic pole. Much of the volume flux is carried by baroclinic flow in narrow currents concentrated

near the surface, and freshwater flux is even more strongly surface intensified (Melling, 2000). Small-scale structure necessitates high resolution in measurement (5 km), whereas surface intensification requires ingenuity in measuring above 30 m where ice risk is high. Year-round capability to measure CAT has emerged only recently through technological innovation (i.e., I-CYCLERS,

Ice thickness is measured by sonar. The spatial structure of the flow was revealed at high resolution by Doppler sonar from the icebreaker USCGC *Healy* in 2003 (Münchow et al., 2006).

Cardigan Strait and Hell Gate are 8 and 5 km wide and 200 and 120 m deep, respectively. Only Cardigan Strait is instrumented, with three Doppler sonars providing ice drift and current

“ALTHOUGH MUCH OF THE RECENT IPY DATA IS STILL BEING ANALYZED, RESULTS ALREADY SHOW THE BENEFITS OF NEW TECHNOLOGIES, ESPECIALLY WITH RESPECT TO SAMPLING NEAR THE ICE-OCEAN INTERFACE.”

discussed in the technology section below; e.g., Fowler et al., 2004). Synchronous observations in all gateways were first acquired in 2003, measuring current, temperature, salinity, ice drift and thickness, sea level, and wind (with different spatial and temporal resolutions in the different channels).

The section with a moored array in Nares Strait is 38 km wide and 300 m deep (Figure 1b). Mooring observations (deployed since 2003) at 5–7 km spacing resolve the Rossby radius, except where moorings have been lost to icebergs. The moored current measuring device (a Doppler sonar) provides data above 40 m depth only if ice is present and then only at the surface—fortunately, ice is prevalent here. Salinity is measured at five levels on moorings that are drawn down greatly by the tide. Salinity above 35 m is not known, except for the near-zero salinity layer (viz. ice) at the surface.

below 30 m. Salinity is measured at the seabed, but  $2 \text{ m s}^{-1}$  tides preclude shallower measurements. The section in Lancaster Sound is over 60 km wide and 200 m deep. Four sites were instrumented during 2004–2005 (13 km spacing), but otherwise only one or two sites are maintained. Here again, Doppler sonar cannot measure above 30 m, but the moored ICYCLER profiler (discussed below) deploys a salinity sensor toward the ice daily at one location (Fowler et al., 2004). The detailed structure of flow through Lancaster Sound is not known.

Volume and freshwater fluxes through Nares Strait (and through other straits with several moorings, for example, Fram Strait, BSO, Davis Strait) are routinely calculated from interpolated current and salinity fields. In Lancaster Sound, where the cross sections of current and salinity are not routinely measured, fluxes are

estimated from a single mooring using empirical scaling factors (Prinsenber and Hamilton, 2005). Because there are few data available to constrain flux estimates for the upper 30–40 m of the ocean, values are subject to unknown biases. For this reason, published flux values (Table 1) for Nares and Cardigan Straits do not include transport within the topmost 40 m.

There are three data sources for Nares Strait fluxes. The first is ship-based acoustic Doppler current profiler (ADCP)/CTD snapshots from 2003, yielding  $-0.9$  Sv for volume and  $-31$  mSv (relative to 34.8) for freshwater flux, with lower-bound estimates for surface-layer transports included (Münchow et al., 2007). Two other studies based on long-term moored observations do not include fluxes above 35 m depth. In the first of these studies, current measurements by Doppler sonar during 2003–2006 yield a three-year average volume flux of  $-0.57 \pm 0.1$  Sv for fluxes below 35 m depth (Münchow and Melling, 2008); seasonal variation is  $\pm 50\%$ . A second estimate for the same period is based upon measured salinity and temperature profiles via the geostrophic method (which assumes a balance between the Coriolis force and oceanic pressure gradients; Rabe et al., 2010). The calculated fluxes,  $-0.47 \pm 0.05$  Sv for volume and  $-20 \pm 3$  mSv for liquid freshwater, represent the instrumented part of the strait only (61% of the width and depth below 35 m). An unpublished study (Humfrey Melling, Department of Fisheries and Oceans, Canada, *pers. comm.*, 2011) of ocean-surface velocity via ice tracking suggests that an additional 0.25 Sv of seawater flux may be transported within the neglected upper layer.

Volume fluxes through Cardigan Strait and Hell Gate are  $-0.2$  and  $-0.1$  Sv, including surface current via ice tracking (2000–2002; Melling et al., 2008). Given estimates are simplistic, derived from a single sonar per strait without knowledge of spatial structure of the flow. Moreover, because of very strong tidal currents, it is not technologically feasible to measure salinity over depth. Therefore, freshwater flux via these passages is not known.

The program in Lancaster Sound dates from 1998. Data from four sites during 2004–2005 have been used to calibrate fluxes derived from fewer sites in other years. Surface-layer salinity has been measured at one site in some years, but the salinity and current in the top 30 m have generally been estimated. The average volume flux during 1998–2006, including the surface, was  $-0.7$  Sv ( $\pm 0.3$  Sv interannual variation) and the freshwater flux  $-48 \pm 15$  mSv (Prinsenber and Hamilton, 2005; Melling et al., 2008). The seasonal cycle is strong and variations appear to be driven by conditions in the Beaufort Gyre (Prinsenber et al., 2009).

At the time of writing, June 2011, instruments continue to record data in all three gateways, but these data will only become available for scientific interpretation when the instruments are retrieved.

### DAVIS STRAIT— A WIDE GATEWAY TO THE NORTH ATLANTIC

The narrow channels of the Canadian Arctic Archipelago (CAA) empty into Baffin Bay, ultimately funneling the Arctic outflow from the western side of Greenland through the wide (360 km), deep (650 m) Davis Strait (Figure 1). Fram and Davis Straits account for 84% of the total freshwater

export from the Arctic, with over half the freshwater export, and slightly more of the liquid freshwater export, exiting through Davis Strait. Arctic waters from the CAA flow southward along Baffin Island as the broad, surface-intensified Baffin Island Current (Tang et al., 2004; Cuny et al., 2005). On the eastern side of Davis Strait, the northward flow consists of the fresh West Greenland Current of Arctic origin above the shelf and the warm, salty West Greenland Slope Current of North Atlantic origin occupying the slope. These inflowing waters circulate cyclonically (anticlockwise) in Baffin Bay and join the southward outflow in the western Davis Strait. Arctic waters undergo transformations before passing through Davis Strait as a result of terrestrial inputs from Baffin Island and Greenland, an annual average heat loss of roughly  $70 \text{ W m}^{-2}$ , brine and freshwater inputs from sea ice formation and melting, and mixing with underlying warmer, more saline Atlantic waters (Rudels, 1986). Davis Strait thus provides a strategic location for characterizing the integrated CAA outflow after transformation during its transit through Baffin Bay and just prior to its entry into the Labrador Sea.

Although scientifically convenient, Davis Strait presents severe observational challenges. As in Fram Strait and in CAA, Davis Strait is dynamically wide (i.e., much larger than the Rossby radius), and thus the flow through the strait admits small-scale eddies and filaments that must be resolved to produce accurate flux estimates. Sea ice cover and icebergs with deep keels make it risky to measure in the upper 100 m (where the majority of the freshwater transport occurs) as well as over the broad (100 km), shallow ( $< 100$  m)

Table 1. Summary of up-to-date estimates of fluxes through the critical oceanic gateways and their variability on different time scales, compiled from available sources. Values are positive for Arctic inflow.

Gateway	Width (max. depth)	Net volume flux (Sv) (mean for years)	Variability (Sv)	Liquid freshwater flux (mSv)	Heat flux (TW=10 <sup>12</sup> W)
Barents Sea Opening	~ 350 km (500 m)	2.0 <sup>1,2</sup> (1997–2007)	interannual 0.8 to 2.9 <sup>2</sup>	55 to 60 <sup>1</sup> (1997–2007)	50 <sup>1,2,*</sup> to 70 <sup>1,**</sup> (1997–2007)
Barents Sea Exit	NZ-FJL <sup>#</sup> ~ 400 km (400 m) SAT <sup>#</sup> ~ 200 km (600 m)	1.5 <sup>3</sup> to 2.0 <sup>4</sup> (1991–1992)	seasonal 0.6 to 2.6 <sup>3</sup> seasonal ± 0.6 <sup>4</sup>		4 <sup>3,*</sup>
Fram Strait	~ 300 km, with shelf ~ 500 km (2,600 m)	−2.0 ± 2.7 <sup>5</sup> (1997–2007) −1.7 <sup>17</sup> (snapshots 1980–2005)	monthly −8.4 to −0.2 <sup>6</sup> interannual −4.7 to −0.3 <sup>6</sup> (2002–2008)	−80 to −66 <sup>7,8</sup> (1997–2008) −65 <sup>17</sup> (snapshots 1980–2005)	36 ± 6 <sup>9,**</sup> (1997–2009) 29 <sup>17</sup> (snapshots 1980–2005)
Bering Strait	~ 85 km, (50 m)	0.8 ± 0.2 (1990–2007) <sup>10,11,14</sup>	daily −2 to 3 <sup>12</sup> monthly 0.4 to 1.3 <sup>12</sup> interannual 0.6 to 1 <sup>12</sup> (uncertainties of 25%)	80 ± 20 <sup>14</sup> (1999–2005)	10 to 20 <sup>11,***</sup> (1998–2007)
Nares Strait	~ 40 km (220 m)	−0.9 <sup>15</sup> (2003 snapshot, 0–30 m estimated)	Not applicable	−31 ± 10 <sup>15</sup> (2003 snapshot, 0–30 m estimated)	Not calculated
		−0.57 ± 0.1 <sup>16</sup> (2003–2006, neglects 0–35 m, which prob- ably adds ~ 0.25 Sv)	± 50% (monthly to interannual periods)	Not measured	Not calculated
		−0.47 ± 0.05 <sup>21</sup> (2003–2006, neglects 0–35 m which probably adds ~0.25 Sv)	± 50% (monthly to interannual periods)	−20 ± 3 <sup>21</sup> (2003–2006, neglects 0–35 m and ice flux, only 61% of width)	Not calculated
Cardigan Strait/Hell Gate	8 km (180 m)/ 4 km (125 m)	−0.3 <sup>13</sup> (2000–2002, includes 0–30 m)	± 50% (monthly to interannual periods)	Not measured	Not measured
Lancaster Sound	~ 40 km (125 m)	−0.7 <sup>18</sup> (1998–2006)	seasonal −1.3 to 0 <sup>18</sup> interannual −1.0 to −0.4 <sup>18</sup>	−48 ± 15 <sup>18</sup>	4 × 10 <sup>−7</sup> 18,*
Davis Strait	~ 320 km (1,040 m)	−2.6 ± 1.0 <sup>19</sup> (1987–1990) −2.3 ± 0.7 <sup>20</sup> (2004–2005)	Not available	−92 ± 34 <sup>19</sup> (1987–1990) −116 ± 41 <sup>20</sup> (2004–2005)	18 ± 17 <sup>19,*</sup> (1987–1990) 20 ± 9 <sup>20,*</sup> (2004–2005)

<sup>1</sup>Smedsrud et al. (2010); <sup>2</sup>Skagseth et al. (2008); <sup>3</sup>Schauer et al. (2002); <sup>4</sup>Gammelsrød et al. (2009); <sup>5</sup>Schauer et al. (2008);

<sup>6</sup>Unpublished data of author Beszczynska-Möller; <sup>7</sup>de Steur et al. (2009); <sup>8</sup>Rabe et al. (2009); <sup>9</sup>Schauer and Beszczynska-Möller (2009); <sup>10</sup>Roach et al. (1995);

<sup>11</sup>Woodgate et al. (2010); <sup>12</sup>Woodgate et al. (2005a) <sup>13</sup>Melling et al. (2008); <sup>14</sup>Woodgate et al. (2006); <sup>15</sup>Münchow et al. (2007); <sup>16</sup>Münchow and Melling (2008);

<sup>17</sup>Rudels et al. (2008); <sup>18</sup>Prinsen et al. (2009); <sup>19</sup>Cuny et al. (2005); <sup>20</sup>Curry et al. (2011); <sup>21</sup>Rabe et al. (2010)

\* heat flux calculated with reference temperature −0.1°C

\*\* heat flux for closed volume budget (reference temperature arbitrary)

\*\*\* heat flux referenced to freezing temperature

<sup>#</sup>NZ-FJL = a passage between Novaya Zemlya and Franz Jozef Land; SAT = Saint Anna Trough

West Greenland shelf. The Davis Strait observing system addresses these challenges by combining a 14-element moored array, which provides spatially coarse measurements of velocity, temperature, salinity, and ice draft, with an extensive marine chemistry program, which provides measurements that integrate broadly over time and space, and new technologies developed as part of the recent Davis Strait efforts (described in the technology section below; Figure 3).

Davis Strait has been the site of the least scientific activity of all the major Arctic gateways. The current effort, started in 2004, builds on measurements taken during 1987–1990 by a six-element moored array (Ross, 1992) that neglected the upper 150 m and the shallow shelves. Three different studies have employed different extrapolation schemes to estimate fluxes from the 1987–1990 data (Loder et al., 1998; Tang et al.,

2004; Cuny et al., 2005), yielding annual mean volume flux estimates of  $-3.3$  to  $-2.6 (\pm 1.2)$  Sv and liquid freshwater fluxes of  $-120$  to  $-92 (\pm 34)$  mSv. In contrast, Curry et al. (2011) use the improved coverage and resolution provided by the current Davis Strait observing system to estimate the annual mean volume ( $-2.3 \pm 0.7$  Sv) and freshwater ( $-116 \pm 41$  mSv) fluxes from 2004–2005. After normalizing for differences in coverage, the 1987–1990 and 2004–2005 flux estimates show no changes that exceed the uncertainties. Curry et al. (2011) also present volume and freshwater budgets for Baffin Bay, finding that they close to within the somewhat large uncertainties in the constituent estimates, with imbalances of 26% (volume) and 4% (freshwater) of the net 2004–2005 Davis Strait fluxes. Ongoing efforts focus on calculating robust flux estimates for subsequent years (2005–present)

and investigating processes to explain the observed variability.

Recent studies (e.g., Holland et al., 2008) show that northward penetration of warm Atlantic waters along the Greenland coast may accelerate the motion of outlet glaciers by melting and breakup of the floating tongues that resist seaward motion. Increased northward heat transport through Davis Strait could thus accelerate Greenland ice cap melting, driving sea level rise and injecting additional freshwater into sensitive regions of the subpolar North Atlantic.

### HIGHLIGHTS OF NEW TECHNOLOGIES IN THE MAIN GATEWAYS

The primary technological innovations for measuring the fluxes of volume and freshwater through Arctic oceanic gateways have been (a) establishing ways of measuring salinity near the

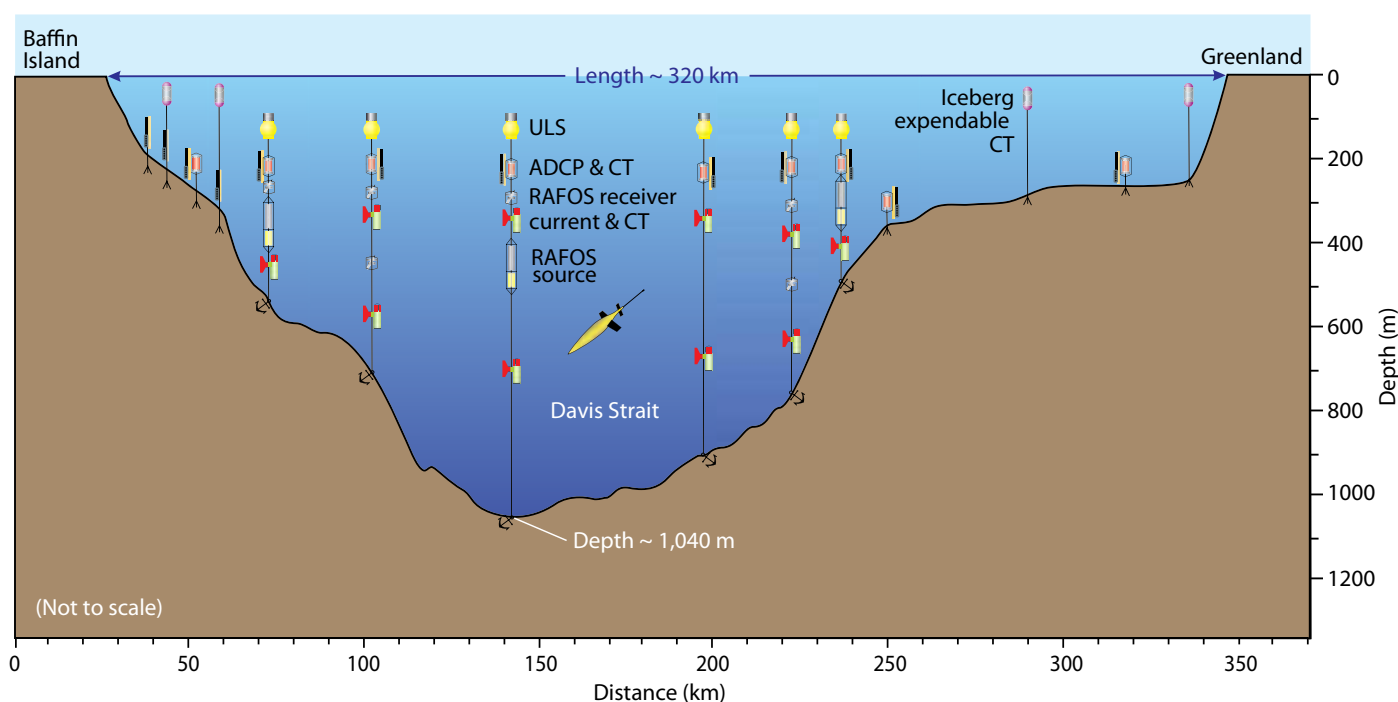


Figure 3. Schematic of the moored array in Davis Strait, looking north. ULS = upward-looking sonar. CT = conductivity and temperature sensor. ADCP = acoustic Doppler current profiler.



ocean surface in the ice-hazard zone and (b) deducing how to practically measure current direction so near to the north magnetic pole.

Several methods have been developed to acquire oceanographic data near the ocean surface, close to the sea ice. In Nares Strait, temperature and salinity sensors have been placed on highly compliant submerged moorings that are passively dragged deeper, away from the ice, when the current is strong and the risk of ice impact is high. Loss to moving ice is thereby minimized. In Lancaster Sound, a novel ICYCLER has been used. This instrument incorporates a small float that carries a CTD and hazard-avoidance sonar close to the surface while a larger mid-water float (containing a controller, a power-minimizing winch system, and data storage) moves away from the surface (Fowler et al., 2004). A third under-ice measurement package, the ICECAT, uses a (possibly) sacrificial temperature-salinity sensor package that transmits its data via inductive coupling to a recording module at a deeper (safer) depth. The connecting line is sufficiently weak that it breaks if the near-surface float is snagged and dragged by ice. This system has been used successfully in both Bering Strait and Davis Strait, with some (anticipated) losses.

An additional new technology—autonomous Seagliders—has been used to gather hydrographic cross sections repeatedly over many months in Davis Strait (Figure 4). Seagliders are underwater vehicles that glide without propellers from the surface to as deep as 1,000 m while collecting information on temperature, salinity, dissolved oxygen, and chlorophyll. When they return to the ocean surface, they transmit their data via satellite and can be remotely

commanded. Gliders provide the ability to resolve hydrographic structures (such as fronts, eddies, and filaments) at the roughly 10 km scale, and to measure salinity near the ice-ocean interface. Sustained support from the NSF Office of Polar Programs and the US Office of Naval Research led to the successful development and implementation of Seagliders capable of extended operation in ice-covered waters in Davis Strait since 2006. Seagliders navigate on a regional scale (hundreds of kilometers), determining their location by measuring sound travel time (viz. distance) from an array of underwater sound sources. Because under-ice gliders operate without human intervention for extended periods (weeks to months), they incorporate enhanced autonomy, including algorithms for navigation, emergency response (if lost or experiencing system failure), and identification and exploitation of ice leads for surfacing to telemeter data back to shore. During a 2009 mission that included 51 days under the ice, a glider made repeated sections across ice-covered Davis Strait while finding and exploiting eight leads (in 33 attempts) for communication through the ice.

Seagliders are also used in Fram Strait to complement mooring observations by providing high-resolution measurements



Figure 4. Applied Physics Laboratory-University of Washington team members launch a Seaglider from R/V Knorr into Davis Strait in 2008. Photo courtesy of Craig Lee

in the upper 1,000 m layer. The acoustic system for under-ice navigation, successfully implemented in Davis Strait, is currently being tested in Fram Strait to permit gliders' access to the ice-covered western part.

The challenge of reference current direction arises from the weakness and variability of the horizontal geomagnetic field component at high latitudes. This problem was addressed by two approaches: (1) the addition of a sensitive compass to the Doppler sonar package and continuous monitoring of local magnetic declination or (2) with installation of the Doppler sonar on a torsionally

rigid mooring at the seabed (data are recorded in the instrument's coordinate reference, and the true heading is established a posteriori from analysis of the tidal oscillations of current).

In recent years, acoustic tomography has been used for instantaneous measurements over long distances in the wide and deep Fram Strait. After a single track acoustic thermometry experiment with acoustic transmissions between a single source and a receiver array, carried out in 2008–2009, an acoustic tomography array consisting of three tomographic sound sources, one receiver array, and passive listening systems has operated in Fram Strait under the ACOBAR (Acoustic Technology for Observing the Interior of the Arctic Ocean) project since 2010. Acoustic travel times along measured tracks provide integrated measures of ocean heat content and currents that will be used for assimilation into numerical models, in combination with data from oceanographic moorings, gliders, and satellites (Sagen et al., 2010).

### UP-TO-DATE ESTIMATES OF OCEANIC FLUXES TO AND FROM THE ARCTIC OCEAN

Despite extensive efforts to estimate oceanic fluxes through the critical gateways, the goal of balanced seawater and freshwater budgets is still far from being achieved. Using available estimates of long-term means of the fluxes (Table 1), the budget for seawater volume can be balanced to within uncertainties; however, the uncertainties in the Fram and Davis Strait fluxes are larger than the total fluxes through other gateways. This problem also contaminates attempts to quantify freshwater fluxes because they are strongly dependent on volume fluxes.

Atmospheric fluxes must also be included in order to understand the heat and freshwater budgets for the Arctic Ocean. Serreze et al. (2006) achieved a balanced freshwater budget for the Arctic Ocean by combining available observations with model results, but they recognized that this balance was likely fortuitous, that uncertainties were still large, and that the variability of freshwater fluxes remains highly undetermined.

With awareness of these challenges, it is likely that better insight may be gained from examining changes in oceanic fluxes through each gateway and assessing their impacts on the Arctic Ocean than from attempting to establish a property flux balance. Because instrumenting all straits with moorings spaced at the Rossby radius is impractical, efforts must focus on (1) understanding the physics of the system, (2) establishing the uncertainties and biases linked to under-resolving measurement arrays, and (3) developing alternative approaches for characterizing the fluxes.

Hydrographic sections can resolve the spatial scales of motion and provide estimates of geostrophic fluxes, but they are scarce and available mainly in summer. In an attempt to balance geostrophic volume fluxes to and from the Arctic Ocean, Rudels et al. (2008) used summer hydrography and estimated net volume transports in Fram and Davis Straits to be smaller than mooring observations, although agreeing within observational uncertainties (Rudels et al.: Fram 1.7 Sv, Davis 1.4 Sv; mooring-based transports: Fram 2 Sv and Davis 2.3 Sv). However, summer sections are likely not representative of the whole year. A good example is Bering Strait, where the large seasonal cycle in water properties in the

strait, from freezing in winter to 10°C at the surface in summer (Woodgate et al., 2005a), and the high day-to-day variability in flow direction and water properties (Woodgate et al., 2005b) necessitate year-round measurements.

In all critical gateways, with the possible exception of Bering Strait where there is also a strong ~ 40 m deep coastal current, the bulk of freshwater is carried in a thin (order 10 m) surface layer. Measurements in this layer (made possible, in part, by the new technologies described above) are critical—including increases in freshwater estimates in this layer in some straits by 20–60% (respectively, Lancaster Sound [Melling et al., 2008] and Davis Strait [Curry et al., 2011]). Near-surface access is thus essential for meaningful assessment of freshwater fluxes, and earlier estimates from the literature need to be revised in this context.

Another difficulty in estimating heat fluxes into the Arctic Ocean arises from the physical concept of oceanic heat transport (see Schauer and Beszczynska-Möller, 2009, for discussion). In general, each gateway carries a nonzero volume flux. Thus, to determine how much heat is delivered to the Arctic through a gateway, it is necessary to ascertain the temperatures of all those waters exiting the Arctic (possibly through different gateways). Useful progress can be made in specific cases. For example, Pacific waters are believed to exit the Arctic with temperatures close to freezing, so this value can be used as a reference for the heat flux of Pacific inflow (Woodgate, et al., 2010). However, the general conclusion is that oceanic heat transport to the Arctic Ocean can be estimated only when all of the inflows and outflows are taken into account.

## NUMERICAL SIMULATIONS— COMPLEMENTING OBSERVATIONAL EFFORTS

Given that many of the observational efforts are regional in nature, basin-scale models are needed to provide connectivity to the rest of the Arctic/subarctic system. Ideally, the combination of modeling and observations allows us to validate model performance in some regions while interpolating between data points in other (data-sparse) regions, while dedicated sensitivity experiments allow us to test hypotheses and suggest causal relationships (Proshutinsky et al., 2011, in this issue).

One major challenge for modeling efforts is to simulate and understand interannual variability in the Arctic and how it relates to inflows and outflows through the gateways. In the early 1990s, intense warming was observed in the Eurasian and Makarov basins (Quadfasel et al., 1991). Model studies (e.g., Häkkinen and Geiger, 2000; Karcher et al., 2003) attributed the observed warming to an increased AW volume inflow and reduced heat loss to the warmer atmosphere. Since the second half of the 1990s, more recent inflows of anomalously warm water into the Arctic Ocean have been documented (Schauer et al., 2008; Skagseth et al., 2008). Karcher et al. (2008, 2011) point out that, both in observations and model results, the anomalously warm inflows were also characterized by negative density anomalies. Karcher et al. (2011) suggest that these anomalies, though reduced in magnitude, survive the long (decadal) passage through the Arctic Ocean, exit via Fram Strait, and go on to alter the properties of the Denmark Strait overflow.

In contrast to Fram Strait, the water

properties of inflows into the Arctic Ocean via the Barents Sea Opening are greatly modified on their roughly two-year-long transit across the Barents Sea. Model simulations may assist in understanding the details of these conversion processes, which are a consequence of variable atmospheric forcing and strong surface heat and salt fluxes (e.g., Harms et al., 2005; Aksenov et al., 2010).

On the Pacific side, there are few recent dedicated Bering Strait through-flow modeling efforts, perhaps because of the complexity of open boundary conditions required for a regional model, or because the strait's small size means it is poorly represented in global models. A notable exception is a data assimilation study of the Chukchi Sea (Panteleev et al., 2010), which combines a year of mooring data into a dynamically consistent model solution, and considers the force balances within the strait.

Model experiments also have contributed to a better insight into the dynamics of freshwater export through the gateways of the Arctic, linking large-scale Arctic Ocean changes with the freshwater flux to subpolar waters. Haak et al. (2003) found that large freshwater and sea ice export events through the Canadian Arctic Archipelago and Fram Strait, which led to the Great Salinity Anomalies (GSA) of the 1970s, 1980s, and 1990s (when large freshenings were observed in the subarctic seas; Belkin et al., 1998), had their source in the Arctic. Karcher et al. (2005) found that due to the change of the Arctic Ocean's thermohaline structure in the early 1990s, a large volume of freshwater exited through CAA and Fram Strait to enter the North Atlantic deepwater production regions. Uncertainty with respect to what determines the partition of freshwater export

between CAA and Fram Strait still exists (Gerdes et al., 2008). Jahn et al. (2010) suggest that the dominant freshwater source for the western CAA is the Pacific and runoff from North America, while in Fram Strait the freshwater is mainly composed of Eurasian runoff and Pacific-derived water. A recent modeling study by Houssain and Herbaut (2011) explains the links between freshwater export through the Canadian Archipelago and surface atmospheric forcing in the western Arctic (wind stress curl) and in the Labrador Sea (air-sea flux anomalies).

Improved knowledge of the variability in the export of freshwater and the properties of the overflows are important to better understand past and future changes in the MOC (Holland et al., 2007). Model studies indicate, for example, that the transport of water to the North Atlantic may substantially influence the MOC (e.g., Mauritzen and Hakkinen, 1997; Holland et al., 2001), as do changes in the routes of freshwater out of the Arctic (Koenigk et al., 2007; Rennermalm et al., 2007).

## SUMMARY AND OUTLOOK

This paper reviews our current state of knowledge of the flow through gateways into and out of the Arctic Ocean. Although much of the recent IPY data is still being analyzed, results already show the benefits of new technologies, especially with respect to sampling near the ice-ocean interface. One next challenge is to design a sustainable Arctic Observing Network for the Arctic gateways. What are the most efficient approaches for quantifying, with known uncertainties, oceanic fluxes of volume, heat, and freshwater? A decadal-scale observing system will likely operate under severe resource constraints, forcing difficult choices.

What approach should be used to optimize such a system?

In many of the straits, flow and water properties vary on the Rossby radius (usually ~ 10 km or less). For a sustained, decadal-scale effort, it is impractical and too costly to instrument the straits with moorings at this spacing. Statistical relationships have been exploited to infer fluxes from measurements made at a small number of critical sites. However, such empirical models rely on stationary statistics. Will such relationships remain static despite the dramatic changes observed in the Arctic and subpolar seas? Because the use of such empirical models may mask signals of Arctic change, the straits would have to be instrumented regularly at sufficient resolution to re-assess their validity. Only by understanding the physics controlling the throughflows and by using this understanding to develop physical (rather than statistical) models of the throughflow can useful and affordable long-term monitoring systems be developed. Physical understanding must therefore be a primary goal of current research.

New technologies and approaches could be exploited to improve existing arrays (Lee et al., 2010). For example, in the deep straits (Fram and Davis Straits), autonomous gliders can provide access to the region near the ice-ocean interface and resolve the horizontal Rossby radius, although at coarse temporal resolution (7–10 days). Combining glider data with moored measurements at coarse spatial resolution allows us to refine flux estimates and better define flow composition in water mass classes. For broad, ice-covered shelves and shallow straits (such as Bering Strait and CAA), where gliders cannot safely operate due to high currents or lack of depth for


maneuvering, a variety of new technologies are gradually providing moored access to the upper water column and the ice-ocean interface.

Wider use could also be made of observational data by incorporating in situ and satellite data in high-quality computer models of the regions (i.e., data assimilation efforts such as that of Pantelev et al., 2010). These techniques are still being developed. Such a system, employing mooring observations, acoustic tomography, gliders, and floats (see Figure 2) together with an eddy-resolving numerical model, was proposed for Fram Strait during the European DAMOCLES study. Implementation of this system started last year under the subsequent European ACOBAR program (Sagen et al., 2010).

Fundamentally, however, optimization of a decadal-scale, sustained observing system for characterizing the Arctic gateways will first require a careful, quantitative assessment of goals. What improvements in estimates of fluxes are required and can be made? To what accuracy must they be quantified in order to resolve long-term changes in the ocean and climate system of the Arctic? What biological and biogeochemical parameters should also be studied, and at what spatial and temporal scales? Careful consideration of our goals will provide guidance for the difficult decisions that lie ahead.

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