Arctic Freshwater Export: Status, Mechanisms, and Prospects

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Abstract

Large freshwater anomalies clearly exist in the Arctic Ocean. For example, liquid freshwater has accumulated in the Beaufort Gyre in the decade of the 2000s compared to 1980–2000, with an extra $\approx 5000 \text{ km}^3$ —about 25%—being stored. The sources of freshwater to the Arctic from precipitation and runoff have increased between these periods (most of the evidence comes from models). Despite flux increases from 2001 to 2011, it is uncertain if the marine freshwater source through Bering Strait has changed, as observations in the 1980s and 1990s are incomplete. The marine freshwater fluxes draining the Arctic through Fram and Davis straits are also insignificantly different. In this way, the balance of sources and sinks of freshwater to the Arctic, Canadian Arctic Archipelago (CAA), and Baffin Bay shifted to about $1200 \pm 730 \text{ km}^3\text{yr}^{-1}$ freshening the region, on average, during the 2000s. The observed accumulation of liquid freshwater is consistent with this increased supply and the loss of freshwater from sea ice. Coupled climate models project continued freshening of the Arctic

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during the 21st century, with a total gain of about 50000 km³ for the Arctic, CAA, and Baffin Bay (an increase of about 50%) by 2100. Understanding of the mechanisms controlling freshwater emphasizes the importance of Arctic surface winds, in addition to the sources of freshwater. The wind can modify the storage, release, and pathways of freshwater on timescales of O(1–10) months. Discharges of excess freshwater through Fram or Davis straits appear possible, triggered by changes in the wind, but are hard to predict. Continued measurement of the fluxes and storage of freshwater is needed to observe changes such as these.

Keywords: Arctic Ocean, hydrological cycle, oceanography, climate change

1. Introduction

Large changes have been seen in the Arctic Ocean freshwater system in recent years, particularly as the observational database ballooned during the International Polar Year (2007–2008). Moreover, oceanographic measurements of freshwater leaving the Arctic through the Canadian Arctic Archipelago (CAA) and Nordic Seas now span a decade. With further widespread changes forecast, the time is ripe for a review and synthesis of knowledge on the freshwater system of the Arctic and Subarctic Ocean. That is our task here. We review current understanding on the status, mechanisms, and prospects for Arctic freshwater, focusing on freshwater export to the Atlantic Ocean. Where possible we synthesize this knowledge to draw new conclusions. The overall goal is to *describe recent changes in the Arctic Ocean freshwater system, to attempt to understand the mechanisms causing these changes, and, on this basis, to speculate about future prospects, especially for the oceanic export of Arctic freshwater*. In particular, we consider the budget of Arctic freshwater, quantifying the storage of freshwater, and the various sources and sinks (section 2). In section 3 understanding of the mechanisms controlling the Arctic freshwater budget is discussed. The prospects for changes in the budget and export fluxes in the coming years and decades are covered in section 4.

Why is the Arctic freshwater system important for global planetary change? The principal reasons are these: First, the Arctic freshwater system is one terminus of the global atmospheric cycle that carries water from low to high latitudes. Evaporation from the warm tropics leads to condensation, precipitation, and accumulation over the cold poles. Second, freshwater plays a leading role in Arctic climate dynamics and climate change. Freshwater as ice reflects solar radiation because of its relatively high albedo; freshwater as liquid forms a thin boundary layer (the halocline) that separates the warmer water below from the atmosphere. Recent changes in the Arctic freshwater system, such as the large decrease of sea ice in summer, support the view that Arctic anthropogenic change is amplified with respect to the global average. Finally, the

31 Arctic freshwater system impacts manifold physical and biological processes,
32 both within the Arctic itself, and at lower latitudes. Many of these processes
33 influence human activities.

34 Figure 1 shows the region of interest. We consider freshwater, as both liquid
35 and ice, in the Arctic Ocean, CAA, and Baffin Bay. This control volume is
36 closed by oceanic sections at Bering Strait (50 m deep, 85 km wide), Fram Strait
37 (2600 m deep, 580 km wide), Davis Strait (1030 m deep, 330 km wide), and
38 the Barents Sea Opening (480 m deep, 820 km wide). The fluxes of mass, heat,
39 and freshwater have been monitored across these sections with oceanographic
40 mooring arrays and ship-based surveys. The export of freshwater through Davis
41 Strait captures the branch of the Arctic outflow through the CAA. The export
42 through western Fram Strait is the other major export pathway, draining the
43 central Arctic of liquid freshwater and sea ice. In the discussion of these fluxes,
44 we emphasize low-frequency changes, contrasting the decade of the 2000s with
45 the 1980s and 1990s. Seasonal and inter-annual variations are not discussed in
46 detail.

47 This paper emerged from a meeting of the Arctic-Subarctic Ocean Flux
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49 delle Scienze Marine, Lerici, Italy.

50 **2. Status of Freshwater Storage and Export**

51 Freshwater in the Arctic Ocean exists in the solid form as sea ice (frozen
52 seawater) and in the liquid form. Liquid freshwater dilutes the upper layers of
53 the Arctic Ocean to create the ubiquitous halocline (the 10–50 m thick near-
54 surface layer of strongly-increasing salinity with depth). Understanding Arctic
55 freshwater involves quantifying where these two phases are stored, and how they
56 are transported and redistributed. Quantifying storage requires knowledge of
57 the distribution in space of liquid freshwater and sea ice¹. Quantifying trans-

¹Freshwater storage in the Arctic is quantified by the amount of zero-salinity water required to reach the observed salinity of a seawater sample starting from a particular reference salinity.

Figure 1: (Continued on the following page.)

Figure 1: Map of the Arctic and subpolar North Atlantic oceans and the Canadian Arctic Archipelago (CAA). The main oceanic flux monitoring sites are indicated with thick red lines. Thin red lines in the main map are used to delimit the Arctic Ocean (and the boundaries of *Serreze et al.*'s (2006) domain; see section 2.1). The freshwater budget discussed in section 2 considers the Arctic, CAA, and Baffin Bay contained within the Bering, Davis and Fram straits, and the Barents Sea Opening (BSO). The bathymetry (topography; from *Terrain-base* (1995)) is shown with blue (gray) colors in meters above sea level.

port requires knowledge of the fluxes of freshwater into and out of the Arctic Ocean. Knowledge of freshwater storage and transport allows the construction of a freshwater budget. The goal in this section is to review the current knowledge of freshwater storage and fluxes and to update the freshwater budget for the Arctic Ocean. The two overarching questions are: *What is the current state of freshwater storage in the Arctic Ocean/CAA and freshwater exchange with neighboring reservoirs?* And, *Where are the greatest uncertainties in freshwater storage and export given the present and anticipated observing efforts?* A summary and a graphical view of the past, present and anticipated future freshwater budget of the Arctic Ocean are provided in Table 1 and Fig. 2. The left hand panel of Fig. 2 shows the budget for, nominally, 1980–2000, the period we

In a similar way, transport of freshwater is quantified as the equivalent flux of zero salinity water. Specifically, liquid freshwater content m (in meters) is estimated as

$$m = \int_D^\eta \frac{S_{\text{ref}} - S}{S_{\text{ref}}} dz \quad (1)$$

for salinity S (all salinities are on the practical salinity scale). The reference salinity S_{ref} equals 34.80, following *Aagaard and Carmack* (1989) and *Serreze et al.* (2006), unless otherwise stated. It is close to the mean salinity for the region of interest (*Tsubouchi et al.*, 2012). The integration with depth z is performed over the fresh upper levels between the S_{ref} isohaline surface, whose depth is D , and the sea surface at height η . Occasionally, D is taken as the depth of a different isohaline surface (for example, *Rabe et al.* 2011 take $S_{\text{ref}} = 35$, but D is the depth of the 34 isohaline). Integrating m over horizontal area yields the total liquid freshwater content (a volume, or inventory, that we quote in km^3).

69 initially consider. The right panel shows the budget for the 2000s (sections 2.2,
70 2.3, 2.4).

71 2.1. Pre-2000 Freshwater Budget

72 *Aagaard and Carmack* (1989) provided the first modern account of the com-
73 plete freshwater budget of the Arctic Ocean. Budget estimates change over
74 time, however, for two reasons: First, new data are collected and the historical
75 databases grow and, second, the system itself changes. For these reasons, this
76 pioneering overview was updated by *Lewis et al.* (2000); *Peterson et al.* (2006);
77 *Serreze et al.* (2006); *White et al.* (2007); *Dickson et al.* (2007, 2008); *Rawlins*
78 *et al.* (2010) and *Beszczyńska-Möller et al.* (2011). These assessments synthe-
79 sized information from individually-published studies on the different compo-
80 nents of the Arctic freshwater system. For example, *Serreze et al.* (2006) used
81 the long-term Polar Science Center Hydrographic Climatology (PHC, version
82 3.0, updated from *Steele et al.* (2001)) to estimate the total annual-mean liquid
83 freshwater content of the Arctic Ocean to be 74000 km³ (see also *Serreze et al.*
84 2008). This volume includes all basins and the surrounding shelves in a domain
85 defined by lines across Fram Strait, the Barents Sea Opening, Bering Strait, and
86 the northern entrance to the CAA (see Fig. 1). Freshwater storage is distributed
87 unevenly; more than half resides in the Canadian Basin, with about 25% in the
88 Beaufort Gyre². The CAA (as far as Hecla and Fury straits) and Baffin Bay (as
89 far as Davis Strait) store about 19000 km³ extra freshwater based on the PHC
90 3.0 climatology for a total of around 93000 km³ (Table 1, Fig. 2).

91 Freshwater storage as solid sea ice is the component of the Arctic freshwater
92 budget with most uncertainty. Horizontal sea ice extent is relatively well known
93 from direct satellite observations, at least since 1979. The uncertainty in sea ice
94 volume is due to sparse information on the spatial and seasonal distribution of

²The Beaufort Gyre circulates above the deep Canada Basin (see Fig. 5 below). The Canadian Basin includes the Canada and Makarov basins. The Beaufort Sea includes the shelf and slope region north of Alaska and northwest Canada (Fig. 1).

	1980–2000 ^a	2000–2010 ^b	21st century ^c
<i>Freshwater reservoirs (km³)</i>			
Liquid freshwater	93000	101000	150000 by 2100 (Fig. 9)
Beaufort Gyre	18500	23500	Increases (with fluctuations?)
As seasonal sea ice ^d	13000	13400	Increases?
As multiyear sea ice ^e	10900	7400	Decreases
As average sea ice	17800	14300	3000 by 2100
Total freshwater volume	110800	115300	~ 150000 by 2100
<i>Freshwater fluxes (km³yr⁻¹)</i>			
Runoff	3900 ± 390	4200 ± 420	5500 by 2100
Bering Strait (liquid)	2400? ± 300+ ^f	2500 ± 100	> 2500
Bering Strait (in sea ice)	140 ± 40	140 ± 40	?
Precip.-Evap.	2000 ± 200	2200 ± 220	2500 by 2100
Greenland flux	330 ± 20	370 ± 25	430 by 2025
Davis Strait (liquid)	-3200 ± 320	-2900 ± 190	-4000 by 2070; -3500 by 2100
Davis Strait (in sea ice)	-160±?	-320 ± 45	?
Fram Strait (liquid) ^g	-2700 ± 530	-2800 ± 420	-6000 by 2100
Fram Strait (in sea ice)	-2300 ± 340	-1900 ± 280	-600 by 2100
Barents Sea Opening	-90 ± 90	-90 ± 90	?
Fury and Hecla straits	-200±?	-200±?	?
<i>Total Fluxes (km³yr⁻¹)</i>			
Inflow sources	8800 ± 530?	9400 ± 490	≳ 11000 by 2100
Outflow sinks	-8700 ± 700	-8250 ± 550	-10000 by 2100
Residual	100 ± 900?	1200 ± 730	~ 1000 by 2100?

Table 1: Arctic/CAA freshwater reservoir volumes and fluxes computed with respect to a reference salinity of 34.80 (positive fluxes freshen the Arctic; see also Fig. 2).

^aTaken from *Serreze et al. (2006)* with some modifications (see section 2.1).

^bSee sections 2.2 and 2.3.

^cSee section 4 and *Vavrus et al. (2012)*. These projections are uncertain.

^dSeasonal sea ice is the winter minus summer sea ice volume (Fig. 3).

^eMultiyear sea ice is the sea ice volume at the end of the summer melt season (Fig. 3).

^fSee section 2.4.2.

^gIncluding the Fram Strait deep water and West Spitsbergen Current.

95 Arctic ice thickness. In constructing their sea ice budget, *Aagaard and Carmack*
 96 (1989) assumed a value of 3 m for the mean sea ice thickness³. This number
 97 multiplied by the 1973–1976 satellite-derived ice coverage, yields a freshwater
 98 volume stored in sea ice of 17300 km³ (in the annual average, and consistent
 99 with the 1980–2000 average of 17800 km³, quoted below in section 2.3)⁴. *Serreze*
 100 *et al.* (2006) followed the same approach; they chose a mean ice thickness of
 101 2 m, reflecting the general thinning taking place across the Arctic. Multiplied
 102 with the 1979–2001 satellite-derived mean sea-ice coverage, *Serreze et al.* (2006)
 103 estimated about 10000 km³ freshwater is stored as Arctic sea ice. This number
 104 is probably too low, at least for a climatology representing the late 20th century.
 105 Consider replacing the *Serreze et al.* (2006) ice thickness estimate of 2 m with
 106 3 m, *Aagaard and Carmack*’s (1989) value and very similar to the estimate of
 107 3.1 m for 1958–1976 by *Rothrock et al.* (1999). The annual-mean freshwater
 108 storage estimate in sea ice is then about 15000 km³, which is much closer to
 109 the *Aagaard and Carmack* (1989) estimate of 17300 km³. As shown below in
 110 section 2.3, the volume of sea ice formed each year is around 13000 km³ (for
 111 1980–2000), most of which melts without leaving the Arctic (section 3.1.1). A
 112 fraction of this seasonal sea ice survives the summer melt to become multiyear
 113 ice, or is exported south.

114 Freshwater is supplied to the Arctic by three principal mechanisms: runoff,
 115 oceanic inflow, and precipitation minus evaporation (P-E). Most important,
 116 runoff from rivers, streams, and groundwater discharge supplies around 3900 ±

³*Kwok and Rothrock* (2009) report mean ice thickness from submarine data for the Central Arctic at the end of the melt season of 3.02 m for 1958–1976. *Laxon et al.* (2003) report a mean winter (October to March) ice thickness of 2.73 m for 1993–2001 from radar altimetry south of 82°N.

⁴Sea ice typically has an average salinity, S_{ice} , of about 4 (*Aagaard and Carmack* 1989; it decreases with age). For ease of comparison, we quote the equivalent liquid freshwater volume stored in sea ice throughout. Namely, we multiply ice volume fluxes by $(1 - S_{ice}/S_{ref})(\rho_{ice}/\rho_w)$, where $\rho_{ice} = 900 \text{ kgm}^{-3}$ is the average density of ice, and $\rho_w = 1003 \text{ kgm}^{-3}$ is the density of seawater with salinity S_{ice} .

117 390 km³yr⁻¹ (assuming 10% error; see below) for 1980–2000 to the Arctic,
 118 CAA, and Baffin Bay⁵. This number is the average of two estimates: First,
 119 the runoff from the ERA-INTERIM atmospheric reanalysis product (*Dee et al.*,
 120 2011) is 4200 km³yr⁻¹ (see *Lindsay et al.* 2014 for a comparison of reanalysis
 121 precipitation products, including ERA-INTERIM). Second, the estimate from
 122 river discharge observations, extrapolated to fill the substantial data gaps, is
 123 3600 km³yr⁻¹. This value is derived from the data shown by *Shiklomanov* (2010)
 124 (his Fig. R1) and adjusted to exclude the Yukon river (about 200 km³yr⁻¹)
 125 and include the contributions from the CAA and Baffin Bay (which add about
 126 500 km³yr⁻¹ more). The average of these two estimates equals 3900±390 km³yr⁻¹
 127 (assuming 10% error), and is reported in Table 1 and Fig. 2 as the 1980–
 128 2000 mean runoff. This number exceeds that of *Serreze et al.* (2006) (3200 ±
 129 320 km³yr⁻¹) for two reasons: First, the present synthesis includes the Arctic
 130 and the CAA as far as Davis Strait, not just the Arctic. Second, the *Serreze*
 131 *et al.* (2006) value comes from observations of river discharge only. The differ-
 132 ence between the estimates from ERA-INTERIM and the discharge data reflect
 133 the combined uncertainty in estimating freshwater runoff from reanalysis prod-
 134 ucts and direct measurements. This error, about 10%, amounts to 390 km³yr⁻¹
 135 in the 1980–2000 runoff estimate. It is consistent with *Lindsay et al.*’s (2014)
 136 estimate of a positive bias in the ERA-INTERIM precipitation fields of about
 137 the same size (their Fig. 3b).

138 The flow through Bering Strait is the next largest source of liquid freshwa-
 139 ter, supplying around 2400 ± 300 km³yr⁻¹ relative to $S_{\text{ref}} = 34.80$ (*Woodgate*
 140 *and Aagaard*, 2005; *Serreze et al.*, 2006). This estimate is based on direct ob-
 141 servations for 1990–2004 of the main-channel flow which accounts for about
 142 1700 km³yr⁻¹. An additional 700 km³yr⁻¹ is added to account for the Alaskan

⁵ We express freshwater fluxes in km³yr⁻¹. To convert to a flux in Sverdrups (Sv) note
 that 1000 km³yr⁻¹ equals 31.7 mSv (1 Sv is 10⁶ m³s⁻¹). Component fluxes are significantly
 affected by different choices of reference salinity S_{ref} , but the net flux for an enclosed region
 is not: see *Tsubouchi et al.* (2012) for a discussion of the effects of choosing different S_{ref} .

Coastal Current and freshwater flux due to seasonal stratification which were not observed throughout this period. Estimating the 1980–2000 average flux is hard because the only years with adequate observations are 1991, 1998, and 1999. In the absence of other data, we quote $2400 \pm 300 \text{ km}^3\text{yr}^{-1}$ in Table 1, mindful of this uncertainty. The Bering Strait ice flux is small in comparison, adding another $140 \pm 40 \text{ km}^3\text{yr}^{-1}$ freshwater into the Arctic (*Travers*, 2012) in 2007, for example.

Finally, the difference between precipitation and evaporation over the region delivers a net flux of around $2000 \pm 200 \text{ km}^3\text{yr}^{-1}$ (from ERA-INTERIM; the 10% error is based on the runoff error above). Noting that about $200 \text{ km}^3\text{yr}^{-1}$ is added over the CAA and Baffin Bay, this estimate is 10% smaller than the ERA-40 (*Uppala et al.*, 2005) value of *Serreze et al.* (2006)⁶. Input of glacial ice as icebergs or glacial meltwater into the Arctic and Baffin Bay is relatively small, around $330 \pm 20 \text{ km}^3\text{yr}^{-1}$, from *Bamber et al.*’s (2012) estimates. Summing each of these sources (Table 1), the total freshwater supply is about $8800 \pm 530 \text{ km}^3\text{yr}^{-1}$.

Freshwater also leaves the Arctic as oceanic liquid freshwater and sea ice. The most important liquid freshwater export route is via the CAA and Baffin Bay at around $3200 \pm 320 \text{ km}^3\text{yr}^{-1}$ through Davis Strait (sea ice adds about 160 km^3 freshwater each year; *Serreze et al.* (2006)). This estimate is inherently uncertain, however, because it is based on 1998–2000 data of the flux at Barrow Strait, not Davis Strait. These observations are then multiplied by a factor of 2–3, from models, to estimate the Davis Strait flux. In comparison, direct measurements of Davis Strait flux were made between 1987 and 1990 using a moored array (*Cuny et al.*, 2005), although the shelves and the upper 150 m were excluded. The liquid freshwater flux was $2900 \pm 1100 \text{ km}^3\text{yr}^{-1}$, extrapolating to estimate the unobserved parts, which is insignificantly different from the $3200 \pm 320 \text{ km}^3\text{yr}^{-1}$ number quoted above.

⁶ It is also derived from the stored reanalysis output fields, rather than the so-called aerological method (*Serreze et al.*, 2006)

Export of both liquid freshwater and sea ice through Fram Strait is also important. The liquid freshwater flux through Fram Strait is around $2700 \pm 530 \text{ km}^3\text{yr}^{-1}$, while export of freshwater as sea ice in Fram Strait is about $2300 \pm 340 \text{ km}^3\text{yr}^{-1}$ (*Serreze et al.*, 2006). The $2700 \pm 530 \text{ km}^3\text{yr}^{-1}$ Fram Strait liquid freshwater flux includes the contributions from the deep water and the West Spitsbergen Current. Freshwater flux across the Barents Sea Opening is relatively weak, $-90 \pm 90 \text{ km}^3\text{yr}^{-1}$, compared to $S_{\text{ref}} = 34.80$ because inflowing salty Atlantic water compensates the inflowing fresh Norwegian coastal current.

The total freshwater export rate for the Arctic, CAA, and Baffin Bay thus sums to about $8700 \pm 700 \text{ km}^3\text{yr}^{-1}$ (including the small flux of about $200 \text{ km}^3\text{yr}^{-1}$, of unknown accuracy, through Fury and Hecla straits based on *Straneo and Saucier* 2008). This flux balances the freshwater sources with a discrepancy that is indistinguishable from zero within the large uncertainty: the residual is about $100 \pm 900 \text{ km}^3\text{yr}^{-1}$ leaving the Arctic (Table 1).

This budget, mainly from *Serreze et al.* (2006), nominally covers the period 1980–2000, roughly speaking before major adjustment in the Arctic hydrological cycle. Since publication of *Serreze et al.* (2006), results from several studies have updated our knowledge of the Arctic freshwater system and how it appears to have changed in the last decade. We now discuss these changes.

2.2. Rapid increase in liquid freshwater storage since 2000

The storage of freshwater in the Arctic Ocean is increasing. The first indication of departure from the climatology of *Serreze et al.* (2006) was provided by *Proshutinsky et al.* (2009). Using data collected in 2003–2007 and historical observations, they found that the freshwater content in the Beaufort Gyre increased by over 1000 km^3 relative to the pre-1990s climatology. The 1990s were also found to be fresher than the climatology of the previous decades. The freshening apparently accelerated during the late 2000s: *McPhee et al.* (2009) found that the freshwater content had increased by 8500 km^3 in the Canada and Makarov basins by 2008. This increase is measured relative to winter climatology (PHC 3.0), and uses extensive aerial surveys carried out in March–April

Figure 2: Schematic Arctic/CAA freshwater budgets. The main reservoirs and fluxes are shown with area proportional to the reservoir volume and the integrated flux in one year, respectively (see the white box for scale). That is, Reservoirs: Liquid freshwater (fw), freshwater stored as seasonal ice and multiyear ice; the liquid freshwater content of the Beaufort Gyre is shown with the circle. Incoming fluxes: precipitation minus evaporation (P-E), runoff, and Bering Strait ocean currents. Outgoing fluxes: Fram Strait (liquid and in sea ice) and Davis Strait. The left panel represents the era before significant Arctic environmental change (1980–2000). The middle panel represents the last decade, and the right panel shows the differences between the two periods. The reference salinity is 34.80. See also Fig. 4 and Table 1.

201 2008. For comparison, it corresponds to about one year’s worth of import (and
 202 export) of freshwater in the 1980–2000 budget discussed in section 2.1.

203 The rapid freshening is evident in other datasets as well. *Rabe et al.* (2011)
 204 used summer salinity profiles from ships, drifting ice stations and autonomous
 205 stations between 2006 and 2008 to estimate the freshwater content for the en-
 206 tire Arctic Ocean with a bottom depth deeper than 500 m. Compared to sum-
 207 mer salinity profiles obtained during the period 1992–1999, they found that
 208 the freshwater content had increased by $8400 \pm 2000 \text{ km}^3$ (in this case relative
 209 to $S_{\text{ref}} = 35.00$). The freshwater content m (eq. (1)) increased across nearly
 210 all of the Arctic between 2006–2008 and 1992–1999. Although the estimated
 211 freshwater content increases of *McPhee et al.* (2009) and *Rabe et al.* (2011) are
 212 similar, one should keep in mind that they are not directly comparable. The
 213 two estimates cover different regions, different times of the year, are based on
 214 different time periods, and use different reference salinities and different lower
 215 levels of integration (*McPhee et al.* 2009 integrated from the depth of the 34.80
 216 isohaline surface, whereas *Rabe et al.* (2011) integrated from the depth of the
 217 34.00 surface: see eq. (1))⁷. Despite these differences, the conclusion is the
 218 same; the Arctic liquid freshwater content increased rapidly during the 2000s
 219 by about 10%. Our estimate of the 2000–2010 average liquid freshwater volume
 220 is therefore 101000 km^3 (Table 1; see also Fig. 7 below).

221 The findings of *McPhee et al.* (2009) and *Rabe et al.* (2011) were corrob-
 222 orated by *Giles et al.* (2012), who used satellite measurements between 1995
 223 and 2010 to show that the dome in sea level associated with the Beaufort Gyre

⁷We can estimate the impact of the last two factors: *Rabe et al.*’s (2011) choice of $S_{\text{ref}} = 35.00$, not 34.80, makes their estimate of the liquid freshwater content *larger* because one integrates a greater salinity anomaly in eq. (1). Their choice of the 34.00 surface, not the S_{ref} surface, as the starting point for integration makes their estimate *smaller* because one integrates over a smaller part of the halocline. These two choices have compensating influence on estimates of freshwater inventory. Using the PHC 3.0 climatology we compute the net effect is a decrease of 1000 km^3 in the *total* freshwater volume. Presumably the effect on the *anomaly* in freshwater volume is less and well within *Rabe et al.*’s (2011) error bars.

inflated and the sea level slope steepened at the edges. They estimated that this inflation corresponds to an increase in freshwater storage of $8000 \pm 2000 \text{ km}^3$ in the western Arctic Ocean. *Rabe et al.* (2014) also recently report that over the period 1992–2012 the liquid freshwater content increased at an average rate of $600 \pm 300 \text{ km}^3 \text{ yr}^{-1}$.

The cause of the inflation, freshening, and increased storage in the 2000s is a wind-driven strengthening of the Beaufort Gyre (see section 3 below for an explanation of this mechanism). The extra freshwater is, at least in part, redistributed from other parts of the Arctic. For example, *Morison et al.* (2012) used a combination of hydrochemistry, hydrography and satellite altimetry and bottom pressure measurements to show that over the period 2005–2008 the dominant liquid freshwater content changes involved an increase in the Canada Basin compensated by a decrease in the Eurasian Basin. The upper waters of the Canada Basin were 1–3 practical salinity units fresher in 2008 than the pre-1990s climatology and 1–2 units saltier in the Makarov Basin. The changes were found to be due to a re-routing of Siberian river runoff associated with changes in the phase of the Arctic Oscillation (see sections 2.5 and 3.1.2).

2.3. Sea ice changes since 2000

Sea ice is the component of the Arctic freshwater cycle with most rapid change. Sea-ice extent is declining, especially in summer. For example, *Vaughan et al.* (2013) show that the linear trend in northern hemisphere monthly-mean sea ice extent is $-3.8 \pm 0.3\%$ per decade for the period November 1979 to December 2012 (considering all months). The corresponding trends for winter, spring, summer, and autumn are $-2.3 \pm 0.5\%$, $-1.8 \pm 0.5\%$, $-6.1 \pm 0.8\%$, and $-7.0 \pm 1.5\%$, respectively. These figures show that the decline in sea ice extent is dominated by loss in summer and autumn. The September sea ice extent reached record-breaking values of $4.3 \times 10^6 \text{ km}^2$ in 2007 and $3.6 \times 10^6 \text{ km}^2$ in 2012 (the 1979–2001 average is $7.0 \times 10^6 \text{ km}^2$).

Sea-ice thickness is also declining. For example, *Kwok and Rothrock* (2009) show that the average ice thickness at the end of the melt season was 3.02 m

Figure 3: Freshwater volume stored as Arctic sea ice from the PIOMAS assimilation product (*Zhang and Rothrock, 2003*). The thin full lines show the seasonally-varying and annual average values. The averages of the minimum (summer) and maximum (winter) volumes are shown with thick dashed lines for the periods 1980–2000 and 2000–2010. Thick full lines show the averages over these periods. The seasonal and multiyear volumes of freshwater stored as ice are shown, from PIOMAS, as is the early average volume estimate of *Aagaard and Carmack (1989)*. The estimates of the recent loss of freshwater from multiyear sea ice by *Kwok et al. (2009)* and *Laxon et al. (2013)* are shown with arrows. See Table 1 and Fig. 2.

254 during the period 1958–1976 (based on submarine data), but just 1.43 m during
 255 2003–2007 (based on ICESat satellite data). Similarly, *Comiso (2012)* shows
 256 that the trend in the extent of multiyear sea ice—which is thicker than first-
 257 year ice—is -16% for 1981–2011 (measured by satellite during winter). This
 258 is a faster rate of decline than for sea ice extent as a whole ($-3.8 \pm 0.3\%$ per
 259 decade, from above), reflecting the preferential reduction of thick, multiyear ice
 260 and hence a decline in average thickness.

261 Concomitant with the declines in sea-ice extent and thickness, sea-ice volume
 262 is shrinking. Perhaps the best estimates of sea-ice volume changes over the last
 263 30 years are from Arctic assimilation products, such as the Polar Science Cen-

264 ter Pan-Arctic Ice Ocean Modeling and Assimilation System (PIOMAS; *Zhang*
 265 *and Rothrock* 2003). The PIOMAS assimilates ice concentration and sea-surface
 266 temperature data, and its sea-ice thickness estimates are validated against satel-
 267 lite products and upward-looking sonars on moorings and submarines. Never-
 268 theless, uncertainty remains in the PIOMAS product, especially for the absolute
 269 ice volume numbers. Figure 3 shows that between 1980 and 2000 the mean PI-
 270 OMAS freshwater volume stored in sea ice is 17800 km^3 , very similar to the
 271 *Aagaard and Carmack* (1989) estimate of 17300 km^3 , and the number quoted
 272 in Table 1. The PIOMAS freshwater volume stored as multiyear ice for 1980–
 273 2000 is 10900 km^3 and the seasonal sea ice is 13000 km^3 (Table 1). For the
 274 decade 2000–2010, the PIOMAS annual mean freshwater volume stored in ice
 275 decreased to 14300 km^3 (with 7400 km^3 as multiyear and 13400 km^3 as sea-
 276 sonal ice). This loss of freshwater stored in multiyear ice agrees, more or less,
 277 with the satellite-based estimate of *Kwok et al.* (2009). For 2011 the PIOMAS
 278 estimate of annual mean freshwater volume in sea ice is 10900 km^3 , a loss of
 279 about 40% compared to the 1980–2000 period. This value accounts for both
 280 sea-ice thinning and sea-ice extent reduction and is similar to *Laxon et al.*'s
 281 (2013) satellite-based estimate (Fig. 3).

282 2.4. Freshwater fluxes since 2000

283 Section 2.1 discusses 1980–2000 conditions. A more updated account on
 284 exchanges through the main oceanic gateways between the Arctic Ocean and the
 285 subpolar seas is provided by *Beszczynska-Möller et al.* (2011). *Tsubouchi et al.*
 286 (2012) present a pan-Arctic flux estimate using mainly hydrographic data and
 287 dynamical constraints, plus some mooring data (relative to $S_{\text{ref}} = 34.66$). Their
 288 estimate is quasi-synoptic because it represents the 32-day period 9 August to
 289 10 September 2005, and so is useful for comparison. Here we discuss the latest
 290 numbers in turn, including variability and trends (see also Fig. 4).

Figure 4: (Continued on the following page.)

Figure 4: Synthesis of ocean freshwater flux timeseries. The upper panel shows runoff and precipitation minus evaporation (P-E) from the ERA-INTERIM reanalysis. An estimate from the river discharge data of *Shiklomanov* (2010) is also shown (see text). The second panel shows Bering Strait fluxes from *Woodgate et al.* (2012). The fourth panel shows fluxes as liquid and stored in sea ice through Fram Strait from *de Steur et al.* (2009) and *Spreen et al.* (2009). The bottom panel shows Davis Strait fluxes from *Curry et al.* (2014). The middle panel shows the net freshwater flux (positive means Arctic freshening) and includes the relatively minor Barents Sea Opening (BSO), Fury, Hecla and Greenland fluxes (light color indicates some missing components). Circles indicate annual mean values with error bars on the 2000–2010 mean values at the right hand side. Shading indicates uncertainties on the annual averages, where available. The data gaps in the Fram Strait sea ice record (black) are filled by reverting to the average seasonal cycle. The values from the 1980–2000 budget in Table 1 are shown with lines on the left. The crosses show the quasi-synoptic flux estimates for summer 2005 from *Tsubouchi et al.* (2012). The stars show the Fram Strait flux estimates from *Rabe et al.* (2013) including the contributions from Pacific and meteoric waters. See also Fig. 2.

291 2.4.1. Runoff and Precipitation minus Evaporation

292 Precipitation over the Arctic has increased in recent years, according to both
 293 atmospheric reanalysis and coupled climate models. For example, using the
 294 ERA-INTERIM product, both runoff into, and P-E over, the Arctic and CAA
 295 were greater in the 2000s than for 1980–2000. Runoff was around $4600 \text{ km}^3\text{yr}^{-1}$
 296 for 2000–2010 compared to $4200 \text{ km}^3\text{yr}^{-1}$ for 1980–2000 (long-term terrestrial
 297 storage effects are small so runoff changes derive from precipitation changes
 298 over land). Similarly, using the adjusted river discharge data from *Shiklomanov*
 299 (2010) to estimate runoff change, we find an increase from 3600 to $3800 \text{ km}^3\text{yr}^{-1}$
 300 between the two periods. Taking the average of these two estimates gives our
 301 estimate of $4200 \pm 420 \text{ km}^3\text{yr}^{-1}$ (Table 1). We roughly estimate the uncertainty
 302 in this value to be 10%, based on the differences between the discharge data

and the ERA-INTERIM reanalysis. ERA-INTERIM P-E was around $2200 \pm 220 \text{ km}^3\text{yr}^{-1}$ for 2000–2010 compared to $2000 \pm 200 \text{ km}^3\text{yr}^{-1}$ for 1980–2000 (also assuming 10% errors; see Table 1 and section 2.1). Freshwater flux from Greenland is also higher; about $370 \pm 25 \text{ km}^3\text{yr}^{-1}$ rather than $330 \text{ km}^3\text{yr}^{-1}$ (*Bamber et al.*, 2012).

It is hard to be sure if these increases in runoff and P-E are real or not. They are both smaller than the nominal uncertainty in Fig. 4 of $\pm 10\%$, based on the differences between the ERA-INTERIM and ERA-40 runoff and P-E numbers quoted in section 2.1. Nevertheless, the ERA-INTERIM product is among the best available. It is one of three out of seven reanalysis products that *Lindsay et al.* (2014) identify as being more consistent with independent observations. They compare the ERA-INTERIM precipitation field with the gridded monthly Global Precipitation Climatology Centre Full Data Reanalysis Version 5 (*Rudolf et al.*, 2010). They find that ERA-INTERIM performs best of all seven models considered in matching the observed precipitation anomalies (the correlation coefficient is slightly less than 0.8). The ERA-INTERIM P-E product is therefore a good choice for our purposes. Moreover, climate models predict increasing precipitation and runoff during the 21st century. For example, *Vavrus et al.* (2012) estimate precipitation increases about 40%, on average, from an ensemble of CCSM4 projections (see section 4.1). Therefore, we suspect that the Arctic precipitation did indeed increase between 1980 and the 2000s. To our knowledge, no study exists that compares Arctic precipitation data from the 2000s with earlier decades, however.

2.4.2. Bering Strait

The Bering Strait import of Pacific (liquid) freshwater amounted to $2500 \pm 630 \text{ km}^3\text{yr}^{-1}$ over the period 1999–2005 (*Woodgate et al.*, 2006). Bering Strait volume flux increased from 0.7 Sv ($22 \times 10^3 \text{ km}^3\text{yr}^{-1}$) in 2001 to 1.1 Sv ($35 \times 10^3 \text{ km}^3\text{yr}^{-1}$) in 2011 with insignificant change in salinity (*Woodgate et al.*, 2012). In consequence, the freshwater flux increased from around $2000\text{--}2500 \text{ km}^3\text{yr}^{-1}$ in 2001 to $3000\text{--}3500 \text{ km}^3\text{yr}^{-1}$ in 2011 (Fig. 4). The year 2001 exhibited

the lowest freshwater flux at $2200 \text{ km}^3\text{yr}^{-1}$ in the period 1998–2011, however. Compared to the uncertainty in the freshwater flux estimate (around $250\text{--}500 \text{ km}^3\text{yr}^{-1}$) the 2001 to 2011 increase in Bering Strait freshwater flux is significant. In Table 1 and Fig. 2 we estimate the 2000–2010 Bering Strait liquid flux to be $2500 \pm 100 \text{ km}^3\text{yr}^{-1}$. This decadal average is indistinguishable from the estimate of $2400 \text{ km}^3\text{yr}^{-1}$ which, in the absence of a complete data record, we take as the best-available, likely poor, value for the period 1980–2000 (section 2.1). *Tsubouchi et al.*’s (2012) quasi-synoptic estimate of the Bering Strait flux for summer 2005 is $2300 \pm 400 \text{ km}^3\text{yr}^{-1}$, close to the annual average of *Woodgate et al.* (2012) for that year.

2.4.3. Fram Strait

The export of liquid freshwater in the East Greenland Current in Fram Strait was $1960 \pm 760 \text{ km}^3\text{yr}^{-1}$ over the period 1997–2008 (*de Steur et al.*, 2009). The 2000–2010 average was nearly $2100 \text{ km}^3\text{yr}^{-1}$, using an improved method by *de Steur et al.* (2014) to fill data gaps. These estimates, from moorings and model results, exclude the West Spitsbergen Current which carries warm salty water polewards. From the perspective of the budget this flow counts as a southward flux of freshwater relative to $S_{\text{ref}} = 34.8$. *Serreze et al.* (2006) estimate it exports $760 \pm 320 \text{ km}^3\text{yr}^{-1}$ which gives a net of around $2800 \pm 420 \text{ km}^3\text{yr}^{-1}$ for Fram Strait liquid freshwater flux (for years 2000–2010; Fig. 4). This number is essentially unchanged from the 1980–2000 value of $2700 \text{ km}^3\text{yr}^{-1}$ (Table 1, Fig. 2). *Tsubouchi et al.*’s (2012) quasi-synoptic estimate of $2200 \text{ km}^3\text{yr}^{-1}$ for the summer of 2005 is noticeably smaller although within error bars. *de Steur et al.* (2009) report that Fram Strait liquid freshwater flux is lowest in summer, so seasonal variability is the likely explanation for the difference. *Rabe et al.* (2013) also provide liquid freshwater flux estimates (stars on Fig. 4)⁸. They are based on six summer-time ship sections and current meter data and agree

⁸The *Rabe et al.* (2013) flux numbers are decreased by 5% to account for their higher reference salinity from Table 1 of *de Steur et al.* (2009).

with the *de Steur et al.* (2009) values. The *Rabe et al.* (2013) flux estimate for summer 2011 is $3900 \text{ km}^3\text{yr}^{-1}$, noticeably larger than the previous 14 years, however, due to a greater Pacific Water contribution.

The Fram Strait export of sea ice is estimated to have carried $2100 \text{ km}^3\text{yr}^{-1}$ freshwater averaged over the winters of 2003–2008 (winters are defined as October through May; *Spreen et al.* 2009). The annual average for 2000–2010 (1990–2000) is 1900 ± 280 (2000 ± 290) $\text{km}^3\text{yr}^{-1}$ when data gaps are filled using the average seasonal cycle (Fig. 4). The quasi-synoptic value quoted by *Tsubouchi et al.* (2012) is $1250 \text{ km}^3\text{yr}^{-1}$ for summer 2005. This number is about half of the annual average, but is unexceptional in light of the annual cycle in sea ice flux reported by *Vinje et al.* (1998) and visible for some years in Fig. 4. For the period 1990–1999, *Kwok et al.* (2004) estimate the freshwater flux in sea ice to be $1800 \text{ km}^3\text{yr}^{-1}$. Their estimate is significantly lower than that of *Serreze et al.* (2006) ($2300 \pm 340 \text{ km}^3\text{yr}^{-1}$, based on *Vinje et al.* (1998)), but it is unclear which is more accurate. Given the large inter-annual variability in sea ice flux ($400 \text{ km}^3\text{yr}^{-1}$ according to *Kwok et al.* (2004)), and the challenge in observing this variable, there is no evident change in Fram Strait sea ice flux (*Spreen et al.*, 2009).

It is interesting that the Fram Strait sea ice flux is apparently unchanged. Changes have been observed in Fram Strait sea ice properties however. During the 2000s the modal thickness of multiyear sea ice in Fram Strait decreased by approximately one third compared to the 1990s (*Hansen et al.*, 2013). In the 1990s the mean sea ice thickness was 3.4 m; for 2005–2010 it had decreased to 2.5 m with a record low of just 2.0 m in winter of 2010. These changes are consistent with the strong decline of (thick) multiyear sea ice in the Arctic as discussed in section 2.3. As the total freshwater flux (and its liquid and solid components) has not been observed to change, a decrease in sea-ice thickness is consistent with an increase in the area of sea ice exported. *Kwok* (2009) and *Kwok et al.* (2013) report no significant trend in sea ice area export through Fram Strait since 1980, however, albeit with significant inter-annual variations. A possible explanation is that the correlation in sea ice speed through Fram

391 Strait and sea ice thickness has increased (so that more thick ice is exported
 392 than before even though thick ice is less abundant). Alternatively, the absence
 393 of evident change in Fram Strait sea ice area and volume fluxes, despite declining
 394 sea ice thickness, could be explained by observing uncertainty.

395 2.4.4. *Davis Strait*

396 For the period 2004–2010, *Curry et al.* (2014) report $2900 \pm 190 \text{ km}^3\text{yr}^{-1}$
 397 liquid freshwater flux and $320 \pm 45 \text{ km}^3\text{yr}^{-1}$ freshwater flux in sea ice. In the
 398 absence of other data, we assume these values represent the decade of the 2000s.
 399 They include the flux through the whole CAA because the flux south of Baffin
 400 Island through Fury and Hecla straits, and hence through Hudson Strait, is
 401 negligible in comparison (about $200 \text{ km}^3\text{yr}^{-1}$ according to *Straneo and Saucier*
 402 (2008)). No significant trend exists in the Davis Strait freshwater flux over 2004
 403 to 2010, nor a significant difference from the 1980–2000 average of $3400 \text{ km}^3\text{yr}^{-1}$
 404 for both liquid freshwater and ice (section 2.1). Nevertheless, the 2004–2010
 405 liquid freshwater flux is significantly smaller than the 1987–1990 average for the
 406 central part of the Strait: *Curry et al.* (2014) estimate the 1987–1990 liquid flux
 407 to be $4500 \pm 730 \text{ km}^3\text{yr}^{-1}$ for this region, but just $3300 \pm 220 \text{ km}^3\text{yr}^{-1}$ for 2004–
 408 2010. The corresponding quasi-synoptic estimate from *Tsubouchi et al.* (2012)
 409 for summer of 2005 is $3700 \text{ km}^3\text{yr}^{-1}$, similar to these longer-term averages and
 410 consistent with the 2005 data shown in Figure 4.

411 2.4.5. *Sources of Uncertainty*

412 All of these flux numbers are uncertain. These uncertainties are quoted
 413 where possible from the original references or based on intuition from detailed
 414 knowledge of the primary observations involved. The uncertainties on the 2000–
 415 2010 average fluxes appear in Fig. 4 as vertical error bars on the right hand side.
 416 Where flux error estimates are available on annual averages, they are shown with
 417 shading. The sources of uncertainty are discussed here.

418 Uncertainties in estimates of meteoric freshwater supply to the Arctic, ei-
 419 ther as precipitation or runoff, stem from uncertainty in atmospheric reanalysis

420 products. In particular, precipitation estimates are not well known. For exam-
 421 ple, the estimates of P-E from reanalysis output fields are lower than those from
 422 the aerological method (*Serreze et al.*, 2006), at least for the MERRA model
 423 (by about a third; *Cullather and Bosilovich* 2011). This result suggests that
 424 our P-E estimates are biased low. *Lindsay et al.*'s (2014) analysis finds that
 425 ERA-INTERIM precipitation is biased high, however, as mentioned in section
 426 2.1. An assessment of Arctic precipitation estimates from ERA-INTERIM that
 427 compares the reanalysis output fields with the aerological method is needed.
 428 Comparison with direct precipitation observations is also needed. Measuring
 429 solid precipitation is challenging, however, and local variability can make inter-
 430 preting sparse station data difficult (*Lindsay et al.*, 2014). Therefore, the 10%
 431 P-E error in Table 1 is a provisional estimate.

432 For the oceanic fluxes, there are several sources of error: First, moored
 433 instruments are threatened by ice. Often, the salinity of the upper 50 m of
 434 the water column is not monitored because sea ice ridges extend down tens
 435 of meters. In those cases, significant anomalies in freshwater flux associated
 436 with near-surface salinity changes are missed. Moreover, icebergs threaten shelf
 437 moorings, especially in Davis Strait. Second, a significant flux occurs over the
 438 broad East Greenland Shelf in Fram Strait (270 km wide) of which only a
 439 small part is monitored with the mooring array. This flux is estimated to be
 440 $800 \pm 400 \text{ km}^3\text{yr}^{-1}$ (from a numerical model; *de Steur et al.* 2009). Third,
 441 the short intrinsic spatial scales in the velocity and hydrography fields (the
 442 baroclinic deformation radius) mean that moorings must be closely spaced to
 443 obtain reliable total fluxes by interpolation. Obstructed access, due to heavy ice
 444 or clearance issues in territorial waters, is also a problem that makes deploying
 445 or recovering moorings harder and leads to gaps in coverage. The calculation
 446 of annual averages are vulnerable to data gaps because most of the component
 447 fluxes show large seasonal cycles (Fig. 4; the averages reported here are for
 448 a calendar year whenever possible). Similarly, inter-annual variations are also
 449 typically large and missing data make decadal averages uncertain. For the same
 450 reason, quasi-synoptic estimates, like that of *Tsubouchi et al.* (2012), do not

451 represent decadal average fluxes accurately.

452 Efforts to reduce these errors continue and substantial progress has been
453 made in the last 15 years. Two developments are particularly noteworthy. De-
454 velopments in oceanographic instrument technology now permit continuous flux
455 monitoring efforts in many ice-covered straits. For example, moored winch sys-
456 tems (such as the ICECYCLER; *Fowler et al.* 2004) can provide temperature
457 and salinity profiles in the upper part of the water column. An acoustic warning
458 system detects and avoids sea ice and thus prevents damage to the sonde. Other
459 designs are passive (such as the ISCAT; *Beszczyńska-Möller et al.* 2011) and are
460 designed to survive being pushed down by the ice. They measure in the upper
461 water column and have been used in strong currents, for example in Bering
462 Strait, which can defeat moored winches. These systems make it possible to
463 determine the freshwater content close to the surface, where it is concentrated,
464 and improve estimates of freshwater flux. Seaglidors, autonomous vehicles that
465 measure hydrographic properties among other variables, are now capable of op-
466 erating under ice (*Webster et al.*, 2014). They are used in wide deep passages
467 that cannot be monitored effectively with traditional moorings. The under-ice
468 capability expands the coverage so that fluxes in Davis and Fram straits can
469 be observed on the shelves. Seaglidors are unable to operate effectively in shal-
470 low straits with strong currents, however, such as Bering Strait. The second
471 noteworthy development concerns numerical circulation models of the Arctic
472 and sub-Arctic seas. They have gained resolution and fidelity since the end
473 of the last century. Models now include processes and dynamical scales rele-
474 vant to observational oceanographers (for a recent review of Arctic models see
475 *Proshutinsky et al.* 2011). Realistic models are used to fill data gaps, quantify
476 variability, for instance in freshwater fluxes, and elucidate the causes of change.
477 Examples include the PIOMAS model mentioned in section 2.1 and *de Steur*
478 *et al.*'s (2009) use of the North Atlantic/Arctic Ocean Sea Ice Model to fill the
479 East Greenland shelf data gap mentioned above.

480 2.5. *Freshwater Origins and Pathways*

481 Along with salinity, measurements of chemical tracers, such as nitrate, phos-
482 phate, oxygen isotopes and alkalinity, reveal the origins of different freshwater
483 sources in the Arctic. Contributions from Pacific Water, meteoric water (runoff
484 and precipitation) and sea-ice melt can all be estimated, as can their changes
485 over time (*Schlosser et al.*, 1994, 1995; *Bauch et al.*, 1995). Pacific Water and
486 river water dominate in the Canadian Basin although their contributions vary.
487 Pacific Water entering through the Bering Strait is found throughout the Cana-
488 dian Basin. Its spread is bounded by two paths: across the central Arctic with
489 the Transpolar Drift or east along the boundary (*Jones et al.* 1998; *Steele et al.*
490 2004, see also section 3.1.2). Meteoric water consists mostly of river water arriv-
491 ing from the Laptev Sea and East Siberian Shelves and flows polewards near the
492 Lomonosov and Mendeleyev ridges (*Ekurzel et al.*, 2001). Pacific Water can
493 be found down to 300 m depth in the southern Beaufort Gyre while river water
494 occurs mostly in the upper 50 m (*Jones et al.*, 2008). Melt water from sea ice
495 is only found in summer in a surface layer: in the halocline there is a negative
496 melt water contribution indicating brine formation from freezing (*Macdonald*
497 *et al.*, 2012).

498 In the early 1990s the front between Pacific and Atlantic derived waters
499 shifted east from the Lomonosov to the Mendeleyev Ridge (*Ekurzel et al.* 2001;
500 *McLaughlin et al.* 1996; *Swift et al.* (2005) discuss evidence of earlier variations).
501 This shift is associated with a change from anticyclonic to cyclonic circulation
502 (section 3.1.2). By 2004 the front had shifted back to the Lomonosov Ridge,
503 returning Pacific Water to the central Arctic (*Alkire et al.*, 2007). Moreover,
504 from the first half of the 1990s to 2005 the inventory of runoff water in the central
505 Arctic increased (*Jones et al.*, 2008; *Newton et al.*, 2013). Data from the 1980s
506 and 1990s show a tight relation between river water and brine which suggests
507 a common source on the continental shelves. By 2005 this relation had broken
508 down, likely associated with the general retreat of summer sea ice (section 2.3)
509 so that brine production from freezing now also occurs in the central Arctic
510 (*Newton et al.*, 2013).

511 Freshwater leaving the Arctic through the CAA consists mostly of Pacific
 512 Water (*Rudels and Friedrich, 2000; Jones et al., 2003*). The total volume flux
 513 through the Archipelago is about twice, perhaps even more, as large as the
 514 Bering Strait inflow, however (see Table 1 of *Beszczyńska-Möller et al. 2011*).
 515 Therefore, a substantial fraction of Atlantic water must also pass through the
 516 CAA and in particular through Nares Strait, the easternmost gap. *Bailey (1956)*
 517 noticed that the deep and bottom water in Baffin Bay has similar properties
 518 to the water at 250 m in the Arctic Ocean and proposed that a deep inflow
 519 through Nares Strait could be the source. *Rudels et al. (2004)* showed that
 520 the properties of the Baffin Bay deep water are similar to those of the lower
 521 halocline in the Canada Basin, which can be traced to the Barents Sea winter
 522 mixed layer. Therefore, they suggested that the Barents Sea inflow branch of
 523 Atlantic water makes the largest contribution to the CAA outflow, both in the
 524 deep outflow and, by mixing with Pacific-derived water, the upper layer outflow
 525 to Baffin Bay.

526 Freshwater leaving through Fram Strait consists mostly of meteoric water
 527 (*Falck et al., 2005; Jones et al., 2008; Dodd et al., 2009, 2012; Rabe et al., 2013*).
 528 Brine dominates over sea-ice melt and the Pacific Water contribution is small
 529 and variable (*Taylor et al., 2003; Falck et al., 2005*). *Rabe et al. (2013)* show
 530 that on average 50% less freshwater was extracted by freezing from the water
 531 present in Fram Strait in the summers of 2009 and 2010, compared to 2005 and
 532 2008. There was on average 30% less meteoric water in 2009 and 2010 compared
 533 to 2005 and 2008. In 2011, nearly four times more Pacific Water contributed to
 534 the freshwater flux compared to the average from 2008, 2009, and 2010. There
 535 was a similarly high fraction of Pacific Water in 1998. These changes can be seen
 536 in Fig. 4 where the Pacific and meteoric water components are plotted (stars)
 537 from *Rabe et al. (2013)*⁹. The extra melt and extra Pacific Water that reached

⁹The brine contribution to the Fram Strait liquid freshwater flux is not plotted but can be deduced as the (positive) flux that must be added to the Pacific and meteoric fluxes (small stars) to equal the total flux (large stars). The brine contribution equals the amount of

538 Fram Strait is likely related to a freshwater anomaly seen in the Lincoln Sea
 539 between 2007–2010 (*de Steur et al.*, 2013). Clearly the rates and/or pathways
 540 of Arctic freshwater transport are changing: mechanisms behind these changes
 541 are discussed in section 3.1.2.

542 2.6. Summary of Freshwater Status and Export

543 Straightforward interpretation of the information in Table 1 suggests the
 544 following: Freshwater sources to the Arctic and CAA have increased in the
 545 2000s compared to the 1980–2000 period. Both runoff and P-E have increased
 546 by about 10%. The freshwater sources sum to $9400 \pm 490 \text{ km}^3\text{yr}^{-1}$ for the
 547 2000s rather than $8800 \pm 550 \text{ km}^3\text{yr}^{-1}$ for 1980–2000. The freshwater sinks
 548 sum to $8250 \pm 550 \text{ km}^3\text{yr}^{-1}$ for the 2000s rather than $8700 \pm 700 \text{ km}^3\text{yr}^{-1}$
 549 for 1980–2000. The 1980–2000 budget therefore sums to $100 \pm 900 \text{ km}^3\text{yr}^{-1}$
 550 freshening the Arctic; the 2000s budget sums to $1200 \pm 730 \text{ km}^3\text{yr}^{-1}$ freshening
 551 it¹⁰. Therefore, these estimates suggest that the Arctic and CAA accumulated
 552 an extra $12000 \pm 7300 \text{ km}^3$ freshwater due to unbalanced fluxes over the decade
 553 of the 2000s (see also section 3.4 and Fig. 7 below). In light of the uncertainty,
 554 this extra freshening is significant, but not strongly. Maintaining the existing
 555 boundary mooring arrays, and adopting the improved observing technologies
 556 described in section 2.4.5 where possible, will likely detect future changes in the
 557 Arctic freshwater system.

558 Another likely explanation for the increased storage in the Beaufort Gyre
 559 liquid freshwater reservoir is the smaller sea ice reservoir (section 2.3). Ac-
 560 cording to chemical tracers in the study by *Yamamoto-Kawai et al.* (2009), an
 561 extra 2.7 m per unit area of sea ice melted in the central Canada basin in 2006
 562 and 2007. Satellite data suggests that melting of multiyear ice in the Beaufort
 563 Gyre accumulated up to 1100 km^3 freshwater between 2004 and 2009 (*Kwok*

freshwater that was extracted by freezing to make sea ice.

¹⁰The *Serreze et al.* (2006) budget sums to $700 \text{ km}^3\text{yr}^{-1}$ salinifying the Arctic, but excluded the CAA and used ERA-40 reanalysis product, not ERA-INTERIM which has a greater precipitation estimate.

564 *and Cunningham, 2010*). Multiyear ice volume also decreased because of less
 565 replenishment from first year ice. Indeed, this is the main reason for recent
 566 decreased total sea ice volume. Over the whole Arctic Ocean, freshwater stored
 567 as ice dropped by approximately 4300 (2800) km³ between the autumns (win-
 568 ters) of 2004 and 2008 (*Kwok et al., 2009*). This extra liquid freshwater is a
 569 substantial fraction of the observed increase.

570 A third possibility exists, albeit less likely: The extra freshwater could come
 571 from a redistribution within the Arctic Ocean, driven, for example, by a change
 572 in the wind (see section 3). The studies claiming increased liquid freshwater
 573 volume in the western Arctic (section 2.2) do not comprehensively sample the
 574 entire Arctic, CAA, and Baffin Bay. Some type of extrapolation to unsampled
 575 areas is unavoidable. Therefore, it is conceivable that freshwater missed in early
 576 inventory estimates was sampled and recorded in the decade of the 2000s. In
 577 this way, the increase in liquid freshwater reservoir volume could be due to a
 578 redistribution from unsampled to sampled areas without there actually being
 579 any real change in the total volume. The size of this effect still needs to be
 580 quantified.

581 On this basis, the state of knowledge of the Arctic freshwater budget is as
 582 follows (see Figs. 2–4, Table 1, and the cited sections for details):

- 583 • Nearly all the Arctic freshwater reservoirs are changing. Liquid freshwater
 584 stored in the Arctic is significantly higher in the 2000s compared to 1980–
 585 2000 (section 2.2). Multiyear sea ice storage is lower (section 2.3). The
 586 most uncertain reservoir term is the sea ice volume, reflecting the challenge
 587 of measuring sea ice thickness.
- 588 • It is hard to detect changes in freshwater fluxes. Nevertheless, general cir-
 589 culation models suggest precipitation increased for the decade of the 2000s
 590 compared to the estimate for 1980–2000 (section 2.4). Similarly, models
 591 and river discharge data show increased runoff. Despite flux increases from
 592 2001 to 2011, it is uncertain if the marine freshwater source through Bering
 593 Strait has changed, as observations in the 1980s and 1990s are incomplete.

Estimates of Fram Strait sea ice and liquid fluxes are unchanged, within error bars, since measurements began in the 1990s (section 2.4.3). The ice is thinner and the area export flux is apparently unchanged, however, suggesting that thick ice is being exported faster, or that the ice volume flux has in fact decreased without being detected. The Fram Strait liquid freshwater contains more ice melt. Observations of Davis Strait liquid fluxes are shorter in duration, and show no obvious changes. The liquid freshwater flux in the central part of the strait was reduced by 26% for 2004–2010 compared to 1987–1990, however. The total net freshwater flux to the Arctic has apparently increased in the 2000s compared to 1980–2000 (Fig. 4, Table 1). Measuring oceanic freshwater fluxes remains a challenge although technology now exists for this purpose (section 2.4.5).

- A shift in the balance of sources and sinks can explain the increase in liquid freshwater stored in the western Arctic although the significance of the shift compared to the total uncertainty is not very high (section 2.6; see also section 3.4 and Fig. 7). A smaller reservoir of sea ice is also probably important. Internal redistribution of freshwater and insufficient sampling of the freshwater reservoirs may also contribute to the observed freshwater increase.

3. Freshwater Mechanisms

Here we discuss mechanisms relevant to storage and export of freshwater from the Arctic. We consider observations, numerical models and theory, where possible. The overarching question is: *What processes govern Arctic Ocean freshwater storage and export?*

3.1. Storage and Distribution

Mechanisms controlling how Arctic freshwater is stored—as ice or liquid—and distributed in space—both horizontally and vertically—are central to understanding the Arctic’s role in the hydrological cycle (*Carmack and McLaughlin, 2011*).

Insight into these mechanisms can be found by first asking, why, in its basic state, is the Arctic Ocean so fresh?

3.1.1. Fresh basic state

As described in section 2, the sources of Arctic freshwater are river runoff, the influx of fresh surface waters through Bering Strait, and the regional imbalance of P-E. These are relatively large sources. For example, the Arctic basin contains approximately 1% of the global ocean volume, but receives 11% of the global river runoff (*Shiklomanov et al.*, 2000). The total annual supply of freshwater (relative to $S_{\text{ref}} = 34.80$) is around 8800 km³ (Table 1). With a total surface area of 9.7×10^6 km² (excluding the CAA and Baffin Bay), this implies that 0.91 m of freshwater is added to the Arctic Ocean each year, similar to high values of P-E in the equatorial Atlantic ocean (*Schmitt et al.*, 1989).

The large seasonal cycle in sea ice also promotes a fresh upper layer. Freezing in winter produces very fresh ice and rejects salt which drains away from the surface as dense brine. Melting in summer returns freshwater to the surface thus distilling, namely un-mixing, the freshwater from the sea (*Aagaard and Carmack*, 1989). About 13400 km³ of freshwater freezes each winter, and about 11300 km³ of freshwater is produced by melting each summer, accounting for the fraction that is exported (Table 1). Therefore, about 1.2 m of freshwater is temporarily added to the surface of the Arctic Ocean by melting, on average, each summer.

Moreover, the Arctic is a place where freshwater tends to remain fresh and concentrated in a small part of the water column. The reason is that the density, and thus stratification, of the Arctic Ocean is primarily a function of salinity rather than temperature (a regime referred to as a β -ocean; *Carmack* 2007). Therefore freshwater tends to remain near the surface and is vertically separated from underlying saltier waters (*Rudels et al.*, 2004). Indeed, the Arctic halocline is strongly stratified, stronger than the typical subtropical stratification above a few hundred meters depth and stronger in summer than the typical equatorial stratification in the upper 30 m. The strong halocline suppresses mixing.

652 Wind-driven mixing and upwelling is further weakened by ice coverage which
 653 reduces the wind's fetch and rate of injection of turbulent kinetic energy. For
 654 these reasons turbulent vertical diffusion of heat and salt across the halocline is
 655 weak. The vertical diffusivity is around $10^{-6} \text{ m}^2\text{s}^{-1}$ in the central Arctic from
 656 a salinity analysis by *Rudels et al.* (1996). This value is ten times larger than
 657 the molecular diffusivity of heat and ten times smaller than that observed in the
 658 quiescent thermocline of the eastern subtropical Atlantic (*Ledwell et al.*, 1993).
 659 Similarly, turbulence measurements in the central Arctic by *Fer* (2009) imply a
 660 halocline diffusivity (of heat) in the range 10^{-6} – $10^{-5} \text{ m}^2\text{s}^{-1}$. This range implies
 661 a negligible diffusive loss of liquid freshwater content of $O(10^{-3}$ – $10^{-2}) \text{ myr}^{-1}$
 662 across the base of the S_{ref} surface, based on the salinity stratification from the
 663 PHC 3.0 climatology.

664 3.1.2. *Wind-forced variability*

665 Given that Arctic freshwater exists primarily near the sea surface, the fresh-
 666 water storage and distribution are strongly influenced by the wind. Here we
 667 briefly summarize the main features of Arctic atmospheric flow involved. Then
 668 we discuss wind-forced variability distinguishing between the western Arctic and
 669 the central and eastern Arctic.

670 The main mode of variability in the Arctic troposphere is the Arctic Oscilla-
 671 tion, or Northern Annular Mode (*Thompson and Wallace*, 1998)¹¹. This mode
 672 involves sea-level pressure variations that strengthen or weaken the pressure
 673 difference between the polar and middle latitudes. The positive phase brings
 674 relatively low sea-level pressure to the Arctic and high pressure in mid-latitudes.
 675 The Arctic Oscillation is a pressure anomaly pattern that depends mainly on
 676 latitude but it is not exactly symmetric about the pole. Instead, the variability
 677 in the central and eastern Arctic, and the Nordic Seas, exceeds that in the west

¹¹The Arctic Oscillation is defined by the first empirical orthogonal function (EOF) of non-
 seasonal sea-level pressure north of 20°N. It is closely related to the North Atlantic Oscillation
 (NAO) which characterises the sea-level pressure difference between the Azores High and the
 Icelandic Low.

Figure 5: (Continued on the following page.)

Figure 5: Atmospheric drivers of Arctic freshwater variability. Each panel shows sea-level pressure (colours; hPa) from the NCEP/NCAR reanalysis product (Kalnay *et al.*, 1996) for the periods indicated. The white lines show the surface flow moving with the average sea ice velocity (from the Polar Pathfinder Sea Ice Motion dataset; Fowler *et al.* 2013). The small red circles show the starting points for the sea ice trajectories which last two years.

(Morison *et al.*, 2012). This asymmetry reflects the mean sea-level pressure field which shows low pressure in the Barents Sea and high pressure in the Canada Basin (Serreze and Barrett, 2011)¹².

The Beaufort High is a prominent anticyclone in mean sea level pressure north of Alaska. A strong Beaufort High is correlated with the summer-time negative phase of the Arctic Oscillation when air pressure is high across the whole Arctic. It is also associated with the Pacific-North American pattern, and (less strongly) with the Arctic dipole anomaly and the Pacific decadal oscillation¹³. The Beaufort High is a center-of-action (that is, a region of high variance) for all these modes of atmospheric variability (Serreze and Barrett, 2011).

Now consider the surface ocean and ice circulation driven by these winds and the impact on freshwater pathways (see also section 2.5). Figure 5 (upper panel) illustrates this flow by showing trajectories of surface particles moving with the 1980–2000 average sea ice velocity (from the Polar Pathfinder dataset; Fowler *et al.* 2013). In the eastern Arctic, including the Barents and Kara Seas, the flow is to the north and/or west. Liquid freshwater and sea ice move into deep water above the Makarov and Eurasian basins forming the Transpolar Drift over

¹²The asymmetry is greater in winter than in summer (Ogi *et al.*, 2004).

¹³Loosely speaking, the Pacific-North American pattern is based on a variance analysis of the height of the 500-hPa surface north of 20°N, the Arctic dipole anomaly is the second EOF pattern in polar sea-level pressure, and the Pacific Decadal Oscillation is the leading EOF of North Pacific sea-surface temperature. Serreze and Barrett (2011) provide details and cite the primary literature.

696 the pole towards Fram Strait (see section 2.5). In the western Arctic the anti-
 697 cyclonic Beaufort Gyre is prominent. The surface ocean and ice circulation is
 698 mainly aligned with the sea-level pressure contours, consistent with geostrophic
 699 flow in the atmosphere and ocean.

700 This surface circulation varies according to the wind in the central and east-
 701 ern Arctic (*Proshutinsky and Johnson, 1997; Rigor et al., 2002; Rigor and Wal-*
 702 *lace, 2004; Morison et al., 2012*). When the Arctic Oscillation is negative the
 703 sea-level pressure is higher across the whole Arctic, but mainly in the east. At
 704 these times, Eurasian runoff flows directly into the Transpolar Drift near the
 705 Lomonosov Ridge. When the Arctic Oscillation is positive Eurasian runoff flows
 706 further east, penetrating the East Siberian Sea, before leaving the continental
 707 shelf (*Steele and Boyd, 1998*).

708 In the western Arctic the ocean and ice flow is driven into one of two regimes,
 709 either cyclonic or anticyclonic (*Proshutinsky and Johnson, 1997*): The cyclonic
 710 regime involves a weak (or absent) Beaufort High sea-level pressure and weak-
 711 ened anticyclonic winds, (or a shifting to cyclonic winds; see Fig. 5 left panel).
 712 Then, the Ekman convergence rate decreases, the halocline ascends, sea level
 713 drops, and isopycnic (isohaline) surfaces flatten. These changes reduce the fresh-
 714 water volume stored in the weakened Beaufort Gyre. Freshwater is released and
 715 redistributed. Some fraction of this redistributed freshwater flows towards the
 716 export channels and drains to the Atlantic (*Karcher et al., 2005; Condron et al.,*
 717 *2009; Stewart and Haine, 2013*).

718 Also during the cyclonic regime, as during the positive phase of the Arctic
 719 Oscillation, Eurasian runoff penetrates further to the east on the shelves and
 720 enters the Canada Basin (*Steele and Ermold, 2004; Dmitrenko et al., 2008*). The
 721 Transpolar Drift shifts east towards the Mendeleyev Ridge and directs freshwa-
 722 ter stored in the Beaufort Gyre and Canada Basin towards Fram Strait, increas-
 723 ing the fraction of Pacific freshwater exiting there. North American runoff tends
 724 to remain on the shelf and exits through the CAA and not east of Greenland
 725 (*Taylor et al., 2003; Dodd et al., 2009*). The summer of 1989 represents this
 726 regime (Fig. 5 left panel).

727 In contrast, the anticyclonic regime is characterized by a strong Beaufort
728 High sea-level pressure and anticyclonic surface winds in the Canada Basin
729 (see Fig. 5 right panel). These winds drive an Ekman convergence of surface
730 freshwater. The halocline in the Beaufort Gyre is depressed deeper, sea level
731 rises up, and the isopycnic (isohaline) surfaces steepen around the edges. In
732 tandem the ocean currents around the edges are stronger leading to a strong
733 Beaufort Gyre. This strengthening is evident as increased sea ice circulation
734 velocity within the Beaufort Gyre (*Kwok et al.*, 2013). These factors cause
735 anomalously large (small) storage of freshwater in the Canada (Eurasian) Basin,
736 as has been seen for the past several years (section 2.2).

737 During the anticyclonic regime, as during the negative phase of the Arctic
738 Oscillation, Eurasian runoff flows off the shelves and into the Eurasian Basin
739 and Transpolar Drift near the Lomonosov Ridge (*Steele and Ermold*, 2004;
740 *Dmitrenko et al.*, 2008). Eurasian runoff is prevented from entering the Canada
741 Basin and exits directly via Fram Strait instead. At these times, Pacific fresh-
742 water tends to be incorporated into the Beaufort Gyre and Canada Basin and,
743 subsequently reduces the Pacific contribution to Fram Strait export (*Falck et al.*,
744 2005; *Dodd et al.*, 2012). The pathway for North American runoff varies; either
745 exiting through the CAA, or entering the Beaufort Gyre (*Yamamoto-Kawai*
746 *et al.*, 2009). This regime is represented in Fig. 5 by the conditions of winter
747 2007 (right panel).

748 This evidence suggests that the Arctic Oscillation and Beaufort High are
749 sometimes linked, sometimes distinct, atmospheric modes that control inter-
750 annual variability in the freshwater system (*Morison et al.*, 2012; *Mauritzen*,
751 2012). The wind interacts with the sea ice cover to drive the surface circula-
752 tion. The surface circulation redistributes freshwater by changing its pathways
753 and residence times. For example, over the period 2005–2008 the Canadian
754 Basin accumulated freshwater while a compensating freshwater loss occurred
755 in the Eurasian Basin (section 2.2, *Morison et al.* 2012). Evidently, the Arc-
756 tic Oscillation determines the freshwater source (runoff, melt) and delivery to
757 the Canadian Basin while the Beaufort High determines Beaufort Gyre fresh-

758 water storage. Consistent with these ideas, accumulation of freshwater during
759 the 2000s coincides with increased anti-cyclonic wind over the western Arctic
760 (*Proshutinsky et al.* 2009; *Rabe et al.* 2014, section 2.2, Fig. 5 right panel). The
761 western Arctic can clearly accumulate or release freshwater according to the
762 Beaufort High strength independent of changes in freshwater sources (*Stewart*
763 *and Haine*, 2013).

764 It is also true that changes in the sea ice characteristics may change surface
765 circulation and hence affect freshwater accumulation. For instance, *Giles et al.*
766 (2012) argue that the increased freshwater seen between 1995 and 2010 (section
767 2.2) is because looser sea ice allowed a more efficient momentum transfer from
768 the wind to the ocean, not from more anticyclonic winds. This mechanism may
769 become more important in future as the summer ice cover disappears.

770 3.2. Import

771 Clearly, changes to the long-term (decadal) Arctic Ocean freshwater inven-
772 tory involve fluctuations in freshwater sources and sinks, not just the wind. We
773 turn to mechanisms controlling sources and sinks of freshwater next. Figure 6
774 provides a schematic summary.

775 3.2.1. Bering Strait

776 Inter-annual variability in freshwater import to the Arctic through Bering
777 Strait matches or exceeds variability from other sources (*Woodgate et al.*, 2012).
778 For example, the standard deviations of the 2000–2010 annual mean freshwater
779 fluxes shown in Fig. 4 are 200, 270, and 270 km³yr⁻¹ for P-E, runoff, and Bering
780 Strait inflow, respectively. This estimate for Bering Strait variability is probably
781 biased low, however, because it derives from changes in the lower layer only and
782 neglects variability in the surface-intensified Alaskan Coastal Current and in the
783 water column stratification (*Woodgate et al.*, 2012).

784 Mooring measurements show that Bering Strait freshwater variability is
785 strongly correlated with volume-flux variability. Volume-flux changes explain
786 more than 90% of the freshwater-flux changes. In turn, volume-flux changes are

Figure 6: Mechanisms of Arctic freshwater fluxes discussed in sections 3.2 and 3.3. ϕ_{FW} is freshwater flux, ϕ_{vol} is volume flux, ϕ_{ice} is ice flux (positive fluxes are poleward), ΔSSH is the sea-level difference (for example between the Bering Sea and the Chukchi Sea in the case of Bering Strait), and v_a is the along-strait component of the surface atmospheric wind (positive poleward). NAO means North Atlantic Oscillation. The proportionality sign means that fluctuations in the two quantities are highly correlated. Colours show the liquid freshwater content (meters, see eq. (1)) from the PHC 3.0 climatology (Steele *et al.*, 2001) representing, nominally, the period 1980–2000.

787 related to both changes in the local wind (about 1/3 of the volume-flux variabil-
788 ity) and changes in the far-field forcing of the flow (about 2/3 of the variability).
789 The latter is often related to a sea-level difference between the Pacific and the
790 Arctic (*Woodgate et al.* 2012; references in *Woodgate et al.* 2005), which in turn
791 is often attributed to a net atmospheric flux of freshwater from the Atlantic to
792 the (fresher) Pacific Ocean (*Stigebrandt*, 1984).

793 3.2.2. *Runoff and Precipitation minus Evaporation*

794 The inter-annual changes in runoff to the Arctic and in P-E are controlled
795 by the polar troposphere. Both freshwater sources are ultimately related to
796 the atmospheric moisture flux convergence across the domain boundaries. For
797 variability on inter-annual periods and longer, the effects of water storage in land
798 ice, snow, and watersheds are relatively minor. Thus, variability in atmospheric
799 supply of moisture and hence precipitation control water supply variability to
800 the Arctic ocean. In general, positive phases of the Arctic Oscillation correspond
801 to greater precipitation in the Arctic and sub-Arctic (*Serreze et al.*, 2008). For
802 example, *Peterson et al.* (2006) document how positive anomalies in Arctic P-E
803 became more frequent as the NAO changed from mainly negative in the mid
804 1960s to positive in the early 1990s. Nevertheless, the links between runoff, P-
805 E, and the Arctic Oscillation are not straightforward and other large sources of
806 variability exist. For instance, summer-time convective precipitation over land
807 has little to do with the Arctic Oscillation.

808 3.3. *Export*

809 Now consider mechanisms affecting the export fluxes of freshwater from the
810 CAA through Davis and Fram straits. We discuss the individual channels in the
811 CAA because mechanisms controlling them are better understood than mecha-
812 nisms controlling the net flux at Davis Strait¹⁴.

¹⁴*Melling et al.* (2008) and *Beszczyńska-Möller et al.* (2011) discuss observations of fresh-
water flux in the CAA.

813 3.3.1. Canadian Arctic Archipelago

814 Arctic freshwater export through the CAA to Baffin Bay occurs by three
815 main routes; via Barrow Strait to Lancaster Sound, via Nares Strait to Smith
816 Sound, and via Cardigan Strait and Hell Gate (which is narrower) to Jones
817 Sound (Fig. 1). The volume flux through Cardigan Strait is less than half
818 that through Nares or Barrow straits and the freshwater flux is still unobserved
819 (*Melling et al.*, 2008). For this reason we omit Cardigan Strait from the discus-
820 sion.

821 Barrow Strait

822 Since 1998 a current meter array has been deployed in Barrow Strait west
823 of Lancaster Sound. The net volume flux is eastward and concentrates at the
824 southern side of the channel. The volume flux is highly variable, with variations
825 as large as the long-term mean of 0.7 Sv ($22 \times 10^3 \text{ km}^3 \text{ yr}^{-1}$). It is stronger in
826 spring and summer than in autumn and winter, perhaps due to land fast ice in
827 winter which retards the surface flow (*Prinsenberg and Hamilton*, 2005; *Melling*
828 *et al.*, 2008). The long-term mean freshwater flux is $1500 \text{ km}^3 \text{ yr}^{-1}$ (*Prinsenberg*
829 *and Hamilton*, 2005; *Prinsenberg et al.*, 2009; *Peterson et al.*, 2012) accounting
830 for over half of the total Arctic export through the CAA (*Beszczyńska-Möller*
831 *et al.*, 2011).

832 The freshwater flux at Barrow Strait is highly correlated with the volume flux
833 (the correlation coefficient exceeds 0.96; *Prinsenberg et al.* 2009); high-resolution
834 numerical models concur (*Jahn et al.*, 2012; *McGeehan and Maslowski*, 2012).
835 Volume flux is highly correlated with wind conditions in the Beaufort Sea, the
836 latter determining the along-channel sea level difference (the correlation coeffi-
837 cient exceeds 0.80; *Prinsenberg et al.* 2009). In particular, *Peterson et al.* (2012)
838 show that northeastward winds in the Beaufort Sea, parallel to the CAA coast,
839 drive the sea-level difference, and hence the volume transport through Barrow
840 Strait. *Melling et al.* (2008) show that the positive phase of the NAO (and
841 hence the Arctic Oscillation) correlates well with increased freshwater flux at
842 Lancaster Sound with an 8-month delay. These authors suggest that 8 months is

843 the timescale for the Beaufort High to respond to the NAO, weaken the Beaufort
844 Gyre and raise sea level upstream of Barrow Strait.

845 *Nares Strait*

846 Volume and freshwater flux observations in Nares Strait are limited, chal-
847 lenging to acquire, and the flow structure is complicated (*Münchow et al.*, 2006,
848 2007; *Melling et al.*, 2008). Although occasional and short term current ob-
849 servations were made in the 1960s and 1970s (*Day*, 1968; *Sadler*, 1976), Nares
850 Strait was the last Arctic gateway where an extensive current meter array was
851 deployed, and not without hardship¹⁵. The current observations, both from
852 the ship and the moorings, indicate a volume flux at Kennedy Channel (North
853 of Smith Sound) for early August 2003 of about 0.7 Sv ($22 \times 10^3 \text{ km}^3\text{yr}^{-1}$).
854 towards Baffin Bay (*Melling et al.*, 2008). A more recent geostrophic estimate
855 is lower, namely, $0.47 \pm 0.05 \text{ Sv}$ ($15 \pm 2.8 \times 10^3 \text{ km}^3\text{yr}^{-1}$). (*Rabe et al.*, 2012),
856 but it is an average over 2003–2006 and excludes the contribution from the up-
857 per 35 m, where the strongest flow is expected¹⁶. The freshwater flux through
858 Nares Strait, $890 \text{ km}^3\text{yr}^{-1}$ (*Rabe et al.*, 2012), is smaller than that in Lancaster
859 Sound because the Nares Strait outflow is saltier (it carries more Atlantic water;
860 section 2.5). These measurements suggest that Nares Strait provides 30–50%
861 of the total CAA volume flux and a similar fraction of the total freshwater flux
862 (*Beszczyńska-Möller et al.*, 2011).

863 Nares Strait freshwater flux is driven by the along-channel pressure differ-
864 ence (*Münchow and Melling* 2008; also seen in the model of *McGeehan and*
865 *Masłowski* 2012). In summer when the ice is mobile, the along-channel wind is
866 also important in driving the freshwater flux (*Rabe et al.*, 2012). The freshwater

¹⁵The array, comprising 16 moorings, was deployed in 2003 from USCGC *Healy* in a joint US-Canadian experiment and was planned to be retrieved from the ice in spring 2005. Due to a severe storm the recovery ice camp had to be abandoned the same day it was established and the retrieval was postponed for a year (*Melling*, 2011).

¹⁶The mooring deployments neglect the upper $\sim 30 \text{ m}$ to avoid instrument damage by ice keels (*Münchow et al.*, 2006; *Rabe et al.*, 2012).

867 flux is influenced by the state of the sea ice, which is either mobile (in summer)
 868 or land fast (*Samleson et al.*, 2006; *Kwok et al.*, 2007). With mobile ice the
 869 freshwater flux is 20% larger than for land fast ice conditions.

870 3.3.2. Davis Strait

871 Arctic water exported through Baffin Bay ultimately transits Davis Strait
 872 before entering the Labrador Sea and leaving our control volume. The freshwa-
 873 ter flux at Davis Strait does not simply equal the summed CAA fluxes, however.
 874 It includes contributions from sea ice processes (freeze/melt); glacial and river
 875 runoff; precipitation less evaporation; and contributions from the West Green-
 876 land Current. The West Greenland Current enters Baffin Bay at the eastern side
 877 of Davis Strait along the West Greenland shelf, flows cyclonically around Baffin
 878 Bay to merge with CAA outflows, and exits the western side of Davis Strait
 879 as the Baffin Island Current. The West Greenland Current salinity is less than
 880 the reference salinity $S_{\text{ref}} = 34.8$ so it adds freshwater to Baffin Bay relative to
 881 S_{ref} (*Curry et al.*, 2014). Some of this freshwater was earlier exported from the
 882 Arctic through Fram Strait in the East Greenland Current and re-enters the
 883 control volume at Davis Strait. Freshwater processes along the path of the East
 884 Greenland Current, such as east Greenland runoff and sea ice melt, influence the
 885 freshwater content of the West Greenland Current. The net flux across Davis
 886 Strait sums these sources of freshwater. As there are several sources to sum
 887 there are several mechanisms at work and no single mechanism dominates, un-
 888 like in Barrow and Nares straits (see above). Moreover, the relative importance
 889 of each contributing mechanism depends on the choice of reference salinity S_{ref}
 890 (see footnote 5).

891 Some facts hint at the mechanisms controlling the net freshwater flux at
 892 Davis Strait, however. First, most of the freshwater flux through Davis Strait
 893 comes from the near-surface outflow driven by the CAA inflows to Baffin Bay
 894 and the West Greenland Current (*Curry et al.* 2014, their Fig. 9). Second,
 895 observations of the near-surface outflow indicate that freshwater and volume
 896 fluxes peak between August and December (*Curry et al.*, 2014). Barrow Strait

897 has peaks in July and August, with minima in November and December (*Pe-*
898 *tersen et al.* 2012, their Fig. 4a). Nares Strait freshwater flux is greatest when
899 the sea ice is mobile, rather than land fast in late winter and spring (*Rabe*
900 *et al.*, 2012). Finally, high-resolution modeling indicates that the freshwater
901 and volume fluxes at Davis Strait are less well-correlated than at Barrow Strait
902 (*McGeehan and Maslowski*, 2012). This finding suggests that CAA freshwater
903 and volume anomalies de-couple in Baffin Bay and/or that other freshwater
904 sources vary significantly too, disrupting the CAA correlation. Better under-
905 standing is needed of how flux variations at Davis Strait inherit from the CAA
906 and the West Greenland Current.

907 3.3.3. Fram Strait

908 Fram Strait supports flow in both directions. To the west is the East Green-
909 land Current which carries virtually the entire Arctic sea ice export (*Kwok*,
910 2009). To the east the West Spitsbergen Current supplies warm salty water
911 of Atlantic origin to the Arctic, one of the primary Atlantic inflow branches
912 (section 2.4.3). The net freshwater flux at Fram Strait sums these sources.

913 Observations show strong correlation between sea ice flux through Fram
914 Strait, which shows large intra- and inter-annual variability, and the cross-
915 strait air pressure difference, which is a proxy for through-strait southward
916 wind (*Vinje*, 2001; *Kwok et al.*, 2004). The through-strait wind in turn relates
917 to the large-scale atmospheric circulation, in particular the NAO (*Köberle and*
918 *Gerdes*, 2003; *Kwok et al.*, 2004). Positive NAO phases correspond to strong
919 Fram Strait winds and thus high sea ice export flux. During times of negative
920 NAO the sea ice flux can be either above or below normal. The flux is high
921 when the Transpolar Drift strengthens and directs ice towards and through
922 Fram Strait (*Kwok and Rothrock* 1999; *Rigor et al.* 2002; *Nghiem et al.* 2007;
923 sections 3.1.2, 3.5). Large sea ice export events require a preconditioning of the
924 upstream sea ice field by the large-scale atmospheric circulation, followed by
925 favourable wind conditions local to the Fram Strait (*Kwok*, 2009). In the last
926 decade it appears that cross-strait air pressure difference increased while sea ice

927 concentration decreased (*Kwok, 2009*).

928 The mechanisms governing the Fram Strait liquid freshwater flux are poorly
929 known. Liquid freshwater is exported through the western end of Fram Strait
930 over the Greenland shelf and shelf break within the East Greenland Current
931 (between 3° – 8° W; *de Steur et al. 2009; Rabe et al. 2013*). The interaction and
932 exchange with the warm salty West Spitsbergen current to the east is hard to
933 observe and not well understood. Observations show that the freshwater and
934 volume fluxes exhibit large inter-annual variability (visible in Fig. 4). A seasonal
935 cycle also exists with flux peaks in September and March (*Jahn et al., 2010; Dodd*
936 *et al., 2012*). During winter the East Greenland Current is mainly barotropic:
937 in summer there is also a baroclinic component (*Aagaard and Coachman, 1968*).
938 Fluctuations in both outflow salinity and speed apparently influence freshwater
939 flux anomalies based on *Jahn et al.*’s (2012) analysis of eight model hindcasts
940 for the last 20–60 years. Unlike the CAA, the correlation between freshwater
941 and volume flux through Fram Strait is weak, however. At times of large liquid
942 freshwater export the halocline deepens over the Greenland shelf in the western
943 Fram Strait and the front with the West Spitsbergen Current steepens (*Rabe*
944 *et al., 2013; Köberle and Gerdes, 2007*).

945 3.4. Rotational Export Control Model

946 We now discuss controls on freshwater outflow due to rotational dynamics,
947 which are relevant to the Arctic straits. Rotational controls lead to a simple
948 model of freshwater export fluxes. The model flux results are compared to the
949 observed fluxes from section 2 (Fig. 4) and provide a context for interpreting
950 the predictions of climate models discussed in section 4 (Fig. 9).

951 Arctic outflow through an opening wider than the first baroclinic Rossby
952 radius is affected by the Earth’s rotation (*Jakobsson et al., 2007*). In these cases,
953 the outflowing layer adheres to the right of the strait (*Werenskiöld, 1935*). The
954 outflow of the East Greenland Current through Fram Strait was described this
955 way by *Wadhams et al. (1979)*. Similarly, *Stigebrandt (1981)* used rotational
956 control in a two-layer model of Arctic outflow. Later elaborations by *Björk*

(1989) and *Rudels* (1989) introduced water mass formation processes on the shelves and *Hunkins and Whitehead* (1992) studied Fram Strait exchanges with a laboratory experiment. Rotational controlled outflows have recently been studied by *Nilsson and Walin* (2010) and *Rudels* (2010).

Salinity controls stratification in the upper Arctic Ocean, and the Rossby radius is determined by the freshwater export flux, lower layer salinity, and Coriolis parameter. The Rossby radius is independent of the mixing rate and volume flux in the upper layer (*Rudels*, 2010). This means that the Rossby radius is controlled by the freshwater thickness at the strait, m_s (see eq. (1) in footnote 1), not by the total depth of the upper layer. The freshwater export flux through the channel is proportional to m_s^2 . If one assumes that m_s equals the average value of m (the liquid freshwater content; eq. (1)) over the Arctic, then the export flux is related to the storage of freshwater in the Arctic Ocean interior. Then, if the total freshwater export flux is known, or can be estimated, it is possible to estimate the freshwater storage, and vice versa (*Rudels*, 2010).

Barrow, Nares, and Fram straits, are wide enough to support a rotational outflow in the upper layer of this type. By taking the lower layer salinities and the upper layer depths to be equal in these passages, *Rudels* (2010) estimated the mean thickness of the freshwater layer in the interior Arctic Ocean to be about 8 m. In the absence of ice export, the freshwater storage must increase to more than 10 m to maintain a freshwater balance (*Rudels*, 2010). The average liquid freshwater thickness over the Arctic Ocean is 8 m, in good agreement¹⁷. The assumption of equal salinity in the lower layers in all three straits is unrealistic. The deep salinity in Fram Strait is about 35, but in the CAA it is 33.5–34. Nevertheless, this discrepancy is compensated by a fresher upper outflow through the CAA than through Fram Strait so that the differences in densities between the upper and lower layers in each passage are similar.

Davis Strait is also wide enough to support rotational outflow and a two-

¹⁷The area of the region is 9.7×10^6 km² (section 3.1.1) and the 1980–2010 average liquid freshwater volume is 97000 km³ (Table 1).

way exchange between Baffin Bay and the Labrador Sea (section 3.3.2). *Rudels* (1986) and *Rudels* (2011) estimated these fluxes using geostrophic balance. To obtain unique solutions he made assumptions about ice formation in Baffin Bay (*Rudels*, 1986) or applied a sea-level difference between the Lincoln Sea and the Labrador Sea that drives the deep outflow through Davis Strait (*Rudels*, 2011). Two effects appeared in this model. First, freshwater input to Baffin Bay (directly or through Davis Strait) freshens the upper layer in Baffin Bay and can reduce volume flux entering through the CAA. The fresh Arctic Ocean upper layer exits the Arctic Ocean through Fram Strait instead. It may eventually arrive in Baffin Bay via the West Greenland Current, thus further freshening the upper layer and reducing the CAA outflow. Second, the Lincoln Sea/Labrador Sea sea-level difference also drives a rotational flow through the passages, but was not considered. For this reason the *Rudels* (2011) estimate of the Arctic Ocean freshwater storage is likely too high.

It is instructive to apply this idealized rotational export model to the flux timeseries shown in Fig. 4. As above, we assume that at Davis and Fram straits the freshwater thickness m_s equals $\bar{m}$, the average value of the freshwater thickness m over the domain. The freshwater budget is:

$$A \frac{d\bar{m}}{dt} = F_{in}(t) + \mathcal{M}(t) - \alpha A^2 \bar{m}^2, \quad (2)$$

$$\frac{dI}{dt} = F_{ice}(t) - \mathcal{M}(t). \quad (3)$$

Here, A is the area of the domain (including the CAA and Baffin Bay), F_{in} is the total inflowing freshwater flux, $\mathcal{M}$ is the freshwater flux due to melting ice, and α is the proportionality coefficient between flux and m_s ¹⁸. Also, I is the volume of freshwater stored in ice and F_{ice} is the export ice flux through Fram Strait (which is negative leaving the Arctic as in Fig. 4; we neglect any inflow of ice). We have estimates of $I(t)$, $F_{in}(t)$, and $F_{ice}(t)$ from Figs. 3 and 4 for the last two decades (F_{in} is the sum of the runoff, P-E, and Bering Strait inflows).

¹⁸ $\alpha = 7 \times 10^{-7} \text{ km}^{-3} \text{ yr}^{-1}$ and ensures balance on average over the period 1980–2000 between F_{in} , $\mathcal{M} = F_{ice}$, and the outflowing liquid freshwater flux.

Figure 7: Idealized outflow model predictions of liquid freshwater (fw) (a) volume (for the Arctic, CAA, and Baffin Bay), and (b) export flux (through Davis and Fram straits). The blue lines show the predictions of the idealized liquid export flux model (see eq. (2) in section 3.4). The red lines show the observed liquid export fluxes (in (b)) and the volume estimate from the integral of the net flux (in (a)); see text). The shading indicates the uncertainty in the 2000–2010 average liquid export flux (in (b)), and the corresponding accumulated volume uncertainty (in (a)). Dashed lines show average freshwater volume and export fluxes. Arrows show the estimates of *Rabe et al.* (2011) and *Rabe et al.* (2014) (section 2.2).

Using them we integrate eq. (2) starting in 1990. Fig. 7 shows the results for the time-varying freshwater volume, $A\bar{m}$, and the total liquid export flux, $\alpha A^2 \bar{m}^2$. The figure also shows the sum of the observed liquid freshwater fluxes through Davis and Fram straits from Fig. 4. Finally, it shows the corresponding liquid freshwater volume obtained by integrating the net flux convergence (that is, by replacing the final term in (2) with the observed export flux; the red line in Fig. 7b). The observed fluxes imply convergence of freshwater in the 2000s. The implied freshwater accumulation is similar to, but somewhat greater than, the independent accumulation estimates of *Rabe et al.* (2011) and *Rabe et al.* (2014). They consider only the Arctic ocean with a bottom depth deeper than 500 m, however, which is smaller than our domain (see section 2). The average for the period 2000–2010 is 101000 km³, the value quoted in Table 1 and section 2.

The idealized outflow model (2) predicts an increasing export flux in the 2000s and hence a smaller increase in liquid freshwater volume. These changes were not observed. A simple explanation for the difference between the predictions of the idealized outflow model and the observed freshwater volumes and fluxes is that the freshwater thickness at Davis and Fram straits did not increase. In other words, m_s was unchanged and was not proportional to $\bar{m}$ in the 2000s. The wind sequestered the extra freshwater in the Beaufort Gyre, away from the drainage channels (as discussed in sections 2.2 and 3.1.2).

Finally, the idealized outflow model connects the export flux to the freshwater thickness squared. This nonlinearity is unimportant in practice, however, because the volume fluctuations are relatively small. Solutions of (2) therefore show nearly exponential relaxation with characteristic timescale $1/(\alpha V_{avg}) \approx 15$ yr, where V_{avg} is the average liquid freshwater volume for 1980–2000 (taken as 93000 km³ from Table 1). This relaxation period is the timescale needed to restore balance between freshwater input and export. It is about ten times longer than the timescales over which the wind modifies the export fluxes through the different channels (*Stewart and Haine, 2013*).

1040 3.5. *Historical freshwater export events: Great Salinity Anomalies*

1041 What can we learn about freshwater mechanisms from the historical record
1042 of variability? Perhaps the most remarkable example of a large freshwater varia-
1043 tion is the Great Salinity Anomaly (GSA) of the late 1960s and 1970s. This event
1044 was a propagating, decadal-scale, surface-intensified freshening of the subpolar
1045 North Atlantic and Nordic Seas (*Dickson et al.*, 1988). It has been attributed
1046 to Arctic freshwater export anomalies of about 10000 km^3 over 5 years (*Curry*
1047 *and Mauritzen*, 2005). Two similar events have been observed: in the 1980s
1048 (*Belkin et al.*, 1998), and 1990s (*Belkin*, 2004). Others may have gone unob-
1049 served (*Wadley and Bigg*, 2004). Indeed, timeseries of salinity in the North
1050 Atlantic reveal several smaller anomalies (*Sundby and Drinkwater*, 2007), and
1051 by “GSA” we refer collectively to the 1970s GSA and other GSA-like events.
1052 On reaching the Labrador Sea, GSAs apparently follow similar cyclonic paths
1053 around the subpolar North Atlantic and Nordic Seas. The 1970s, 1980s, and
1054 1990s anomalies are detectable in salinity data for about a decade.

1055 A link may exist between GSAs and the large-scale wind circulation regime,
1056 especially the NAO, although it is not well understood (*Dickson et al.*, 2000).
1057 Cyclonic winds over the Canada Basin tend to increase freshwater export from
1058 the Arctic and anticyclonic winds tend to retain freshwater there (section 3.1.2).
1059 The 1980s and 1990s events occurred when the winds were cyclonic and both
1060 of these anomalies apparently emerged west of Greenland. The 1970s GSA
1061 occurred when the winds were strongly anticyclonic, however, suggesting that
1062 freshwater should have been strongly retained in the western Arctic. This “para-
1063 dox” (*Dickson et al.*, 2000) can be understood by recalling that the Transpolar
1064 Drift strengthens under anticyclonic wind forcing (as in 2007; Fig. 5 lower right
1065 panel). Fram Strait freshwater export can increase at these times (mainly due
1066 to the export of sea ice) even as Beaufort Gyre freshwater content rises. Hence,
1067 a GSA during cyclonic (anticyclonic) winds likely results from an increased
1068 liquid freshwater export through the CAA (increased sea ice export through
1069 Fram Strait). Recent work suggests that a large wind-driven freshwater release,
1070 around 10000 km^3 in 5 years, can only occur if freshwater storage in the Beau-

1071 fort Gyre is already anomalously high (*Stewart and Haine, 2013*). Otherwise,
1072 the freshwater volume released is significantly smaller.

1073 The mechanism of GSA propagation is an open question. Specifically, the
1074 decadal lifetime is hard to understand. One idea is that anomalous pack-
1075 ets of freshwater are advected passively by the otherwise unchanged currents
1076 (*Belkin et al., 1998*). Numerical models of GSA propagation implicate fresh-
1077 water flux anomalies and/or circulation anomalies, however (*Wadley and Bigg,*
1078 2006). Timeseries observations reported by *Sundby and Drinkwater (2007)* seem
1079 to agree. A positive feedback may be important: the fresh surface damps deep
1080 convection, reducing both sea-surface temperature and ocean-atmosphere heat
1081 flux (*Gelderloos et al., 2012*). This cooling favors precipitation, further reinforc-
1082 ing the fresh anomaly.

1083 3.6. Summary of Freshwater Mechanisms

1084 The main points are summarized as follows (see the cited sections and figures
1085 for details):

- 1086 • The Arctic upper Ocean is relatively fresh because it has a large supply of
1087 freshwater from runoff, Bering Strait inflow, and precipitation compared
1088 to its volume. Seasonal freezing and melting promotes a fresh surface
1089 by ice distillation. Also, the turbulence intensities in the halocline are
1090 exceptionally small, reducing the flux of salt mixed up from below (section
1091 3.1.1).
- 1092 • Wind stress over the Arctic controls ice motion and the surface ocean
1093 currents, and hence determines freshwater pathways and accumulation
1094 (section 3.1.2; Fig. 5). The Arctic Oscillation and fluctuations in the
1095 atmospheric Beaufort High sea-level pressure are particularly influential.
1096 In the last decade there has been an increase in Ekman pumping driven by
1097 the Beaufort High that has increased freshwater storage in the Beaufort
1098 Gyre at least partly by drawing freshwater from other regions.

- 1099 • Convergence of tropospheric moisture, and hence precipitation, controls
1100 the net supply of freshwater to the Arctic from the atmosphere (section
1101 3.2.2). The Arctic Oscillation is an important, but not dominant, influence
1102 on this mechanism.
- 1103 • The marine freshwater inflow through Bering Strait is believed to be con-
1104 trolled by the Pacific-to-Arctic sea level difference and moderated by the
1105 local southward wind (section 3.2.1, Fig. 6). In Bering Strait the fluctua-
1106 tions in volume flux are highly correlated with those in freshwater flux.
- 1107 • Similarly, fluctuations in volume flux and freshwater flux are highly corre-
1108 lated in Barrow and Nares straits (section 3.3.1). In both these channels
1109 the volume flux is highly correlated with the along-channel sea level dif-
1110 ference. In Barrow Strait the along-channel sea level difference correlates
1111 with the Beaufort Sea wind field. In Nares Strait the along-channel south-
1112 ward wind correlates with the volume and freshwater fluxes in summer
1113 when the ice is mobile.
- 1114 • Davis and Fram straits support two-way exchange and several mechanisms
1115 compete because there are several sources of freshwater contributing to the
1116 net flux through these straits (sections 3.3.2 and 3.3.3). Fram Strait ice
1117 flux is driven mainly by the local southward wind. Sea ice decline in
1118 the 2000s is apparently compensated by increased flow speed to maintain
1119 about the same sea ice flux through Fram Strait.
- 1120 • An idealized model of a rotational outflow predicts that the liquid freshwa-
1121 ter flux is controlled by the liquid freshwater content at the export straits
1122 (section 3.4). The characteristic response timescale of the model is 15
1123 years. Over the 2000s, the model predicts an increasing outflow through
1124 Davis and Fram straits (Fig. 7). The observed freshwater fluxes did not
1125 change significantly, however, because the excess freshwater was stored in
1126 the Beaufort Gyre away from the drainage channels. The observed fluxes
1127 (Fig. 4), and loss of sea ice (Fig. 3), suggest accumulation of liquid fresh-

1128 water that is consistent with observations, although the uncertainties are
1129 large (Fig. 7).

- 1130 • Three major freshwater export events seem to have occurred since the
1131 mid-1960s; two through the CAA and Baffin Bay, and one through Fram
1132 Strait (section 3.5). They caused Great Salinity Anomalies that moved
1133 through the subpolar North Atlantic and Nordic Seas in about ten years.
1134 The export mechanism apparently involves wind shifts in the Beaufort
1135 Sea, but the details are not understood. The mechanism behind the long
1136 lifetime is also unknown.

1137 To conclude this section on freshwater mechanisms, the importance of un-
1138 forced, internal variability should not be forgotten. By unforced variability we
1139 mean fluctuations that are not due to changes in forcing from the atmosphere
1140 or due to changes in sources and sinks of freshwater. Instead, the variability is
1141 caused by intrinsic chaotic dynamical processes in the ocean/ice system. It is
1142 hard to quantify the magnitude of such variability, but for the total freshwa-
1143 ter volume it appears to be several thousand km³ based in Arctic/sub-Arctic
1144 Ocean/ice models with steady forcing (*Stewart and Haine, 2013*). Almost cer-
1145 tainly, the decline in sea ice since 2000 is due to anthropogenic climate change,
1146 not internal variability (*Notz and Marotzke, 2012*), but the contribution of nat-
1147 ural variability to changes in the freshwater reservoir volume is unclear. The
1148 forcing and supply mechanisms identified in this section compete with these
1149 unforced internal fluctuations and are often hard to distinguish.

1150 4. Prospects for Arctic Freshwater

1151 Climate model projections of Arctic freshwater variables are diverse and
1152 thus uncertain. Nevertheless, most climate model projections of Arctic fresh-
1153 water variables are similar enough to infer some probable changes in the future
1154 Arctic liquid freshwater storage and export. Here we discuss these prospects and
1155 compare the freshwater changes described in section 2 with climate model pro-

jections. The overarching question is: *How do we expect the freshwater system will change in future?*

4.1. Robust climate signals

There are several robust signals that emerge consistently from climate model projections: First, a warmer climate features a stronger hydrological cycle with greater atmospheric moisture transport to high latitudes. Therefore, precipitation over, and runoff to, the Arctic Ocean is projected to increase based on coupled climate models (see *Kattsov et al. (2007)*, for an overview of CMIP3¹⁹ models and *Vavrus et al. (2012)*, for an example of a CMIP5 model). For example, the Community Climate System Model, version 4 (CCSM4) shows about a 40% increase in precipitation polewards of 70°N over the 21st century (*Vavrus et al., 2012*). The main reason Arctic precipitation increases in CMIP5 models is increased local evaporation (in winter), signifying an accelerated freshwater cycle within the Arctic itself (*Bintanja and Selten, 2014*). Increased atmospheric moisture transport is less important (and peaks in late summer and autumn). Second, sea ice extent declines in the northern hemisphere in all seasons (*IPCC (2007)*, Chapter 10). The rate of decline differs greatly among models and most models underestimate the recently-observed summer sea ice retreat seen in Fig. 3 (section 2.3). The aforementioned acceleration in Arctic precipitation and evaporation is linked to winter-time sea-ice retreat (*Bintanja and Selten, 2014*). Third, ice volume decreases in all CMIP5 models over the 21st century (Julienne Stroeve, pers. comm., 2012). The volume of sea ice at the end of the 20th century and the rate of change in Arctic sea ice volume again varies greatly among models. The declining ice volume results in smaller ice thickness in Fram Strait and decreasing ice export rates (*Holland et al., 2007; Koenigh et al., 2007; Vavrus et al., 2012*). Because both sea ice volume and export flux

¹⁹CMIP is the Coupled Model Intercomparison Project serving coupled climate model projections to the Intergovernmental Panel on Climate Change (IPCC). CMIP3(5) models are from the fourth (fifth) assessment reports in 2007 (2013).

1182 decrease over time, the net (annual-mean) thermodynamic growth rate must
1183 also decline. It is possible, however, that the seasonal cycle might increase with
1184 higher freezing and melting rates in the Arctic Ocean.

1185 *4.2. Consequences for Freshwater Storage and Export*

1186 Increasing freshwater input into the Arctic Ocean through P-E and runoff
1187 and decreasing ice export flux implies either (transient) liquid freshwater storage
1188 or increasing liquid freshwater export rates from the Arctic (or both). Different
1189 models behave differently in these respects. For example, the freshwater content
1190 changes over the 21st century are shown for four CMIP models in Fig. 8. The
1191 upper panels show the differences in liquid freshwater content between the end
1192 (2090–2100 average) and the beginning (2000–2010 average) of the 21st century
1193 for the CCSM3 and ECHAM5-OM-MPI models (both from CMIP3). Only one
1194 realization of the IPCC’s A1B scenario is shown although the decadal variability
1195 can be substantial. We see gains of several meters in CCSM3 in the western
1196 Arctic whereas the eastern Arctic loses freshwater (compare to Fig. 6). In
1197 the ECHAM5-OM-MPI model, freshwater content increases by 10–20 m over
1198 virtually the whole Arctic Ocean. The total liquid freshwater content increases
1199 by 24000 km³ (from 118000 to 142000 km³) in the CCSM3 model over the
1200 21st century. For the ECHAM5-OM-MPI model, the corresponding increase is
1201 63000 km³ (from 110000 to 173000 km³). These models begin the 21st century
1202 with moderately realistic total liquid freshwater volumes: for the decade of the
1203 2000s the estimate from section 2 is 101000 km³ (Table 1).

1204 A likely reason for these striking freshwater differences is different ocean
1205 volume (and freshwater) fluxes between the Arctic and the subpolar North At-
1206 lantic. In the ECHAM5-OM-MPI model, increasing meteoric freshwater input
1207 and reduced sea ice export lead to increased storage of liquid freshwater. The
1208 CCSM3 model responds to these input and ice changes by increasing oceanic
1209 volume exchange with the subpolar Atlantic. For example, the surface salinity
1210 decreases along the export pathways east and west of Greenland and in the west-
1211 ern Labrador Sea (not shown). As a consequence, more saline Atlantic water

Figure 8: Arctic ocean liquid freshwater content differences between the end (2090–2100 average) and the beginning (2000–2010 for the upper panels, 2010–2020 for the lower panels) of the 21st century from four CMIP climate models. Upper panels show two CMIP3 models (A1B scenarios; CCSM3 and ECHAM5-OM-MPI). Lower panels show two CMIP5 models (RCP4.5 scenarios; MPI-ESM-LR and GFDL-CM3).

enters the Arctic, reducing the freshwater content in the eastern Arctic Ocean. The integrated response in Arctic and CAA freshwater storage is smaller in CCSM3 compared to ECHAM5-OM-MPI for these reasons.

Liquid freshwater content differences are shown for two CMIP5 models in the lower panels of Fig. 8 (RCP4.5 scenario). They behave similarly to the CMIP3 calculations with CCSM3 and ECHAM5-OM-MPI. Again, in some places the freshwater content increases, but in others it decreases. The same is true for the surface salinity (not shown). For example, the MPI-ESM-LR model freshens in all deep basins of the Arctic Ocean and increases salinity on most shelves. Enhanced import of saltier Atlantic waters causes salinity to increase below the halocline whereas liquid freshwater content increases in the Beaufort Gyre. The MPI-ESM-LR model accumulates 33000 km^3 (from 125000 to 155000 km^3) liquid freshwater in the 21st century. In contrast, the GFDL-CM3 model accumulates freshwater in the Beaufort Gyre and Lincoln Sea and loses it in the Eurasian Basin; the shelf salinities change only weakly. The GFDL-CM3 model accumulates 36000 km^3 (from 117000 to 153000 km^3) liquid freshwater over the 21st century for this realization.

Vavrus et al. (2012) describe the Arctic Ocean evolution over the 21st century in the CCSM4, the NCAR model used for CMIP5. CCSM4 includes an open Nares Strait, unlike CCSM3, which allows an increased freshwater export through the CAA. *Vavrus et al.* (2012) find a freshening of the surface in the Arctic Ocean over the 21st century. There is a 28% increase in liquid freshwater storage in the Arctic Ocean. This increase is very similar to those for the MPI-ESM-LR and GFDL-CM3 models, quoted above (although the details of the liquid freshwater content calculations differ: *Vavrus et al.* (2012) consider only the upper 250 m of the Arctic Ocean and exclude the CAA and Baffin Bay). The CCSM4 sea ice stores 80% less liquid freshwater by the late 21st century. Together, liquid and sea ice account for a moderate increase of 9% in the total freshwater storage. CCSM4 Bering Strait freshwater flux into the Arctic also increases. The liquid freshwater export through Fram Strait increases through the 21st century (by about $3200 \text{ km}^3 \text{ yr}^{-1}$), whereas sea ice export drops substan-

Figure 9: Twenty-first century liquid freshwater (fw) prospects from CMIP models. The volume of liquid freshwater for the Arctic Ocean, CAA, and Baffin Bay is shown. The estimates from Table 1 and Fig. 7 are shown at bottom left. The CMIP3 (CMIP5) models are realizations of the A1B (RCP4.5) scenario.

tially (by about $1800 \text{ km}^3\text{yr}^{-1}$ to about $600 \text{ km}^3\text{yr}^{-1}$). The liquid freshwater export through the CAA first increases (by about $900 \text{ km}^3\text{yr}^{-1}$) then decreases after 2070 (by about $300 \text{ km}^3\text{yr}^{-1}$) when decreasing CAA volume flux dominates the decreasing salinity. This development is attributed to weakening convection in the Labrador Sea, which grows fresher over the 21st century. If CCSM4 behaves similarly to CCSM3, the downstream freshening would raise sea level and decrease the surface pressure gradient between the Arctic and the Labrador Sea, hence reducing the CAA volume flux. Details of this mechanism remain unclear, however. Finally, *Vavrus et al.* (2012) find strongly increasing temperature of the CCSM4 Atlantic Water layer indicating stronger inflow of Atlantic Water.

4.3. Summary of Freshwater Prospects

Figure 9 condenses the results from these CMIP climate models. It shows time series over the 21st century of the volume of liquid freshwater, and includes the results of the budget analysis in section 2 and the freshwater model in section 3. Given the various sources of error, and variability, we see good agreement in general. The models over-estimate the liquid freshwater volume somewhat, but

the discrepancy is only 10–20% of the total. The increasing freshwater trend inferred from observations in Fig. 7a (the red line) is similar to, but generally greater than seen in the models. Recall that the models underestimate the decline in ice volume, however, compared to the observations. In other words, the models underestimate the increasing liquid freshwater trend in the first few decades of the 21st century for this reason. Finally, the models show consistent increases in freshwater until at least mid-century. After that, some reductions in freshwater volume of 5000–10000 km³ over 5–10 years occur, for the GFDL-CM3 model, in particular. These events may resemble GSAs (section 3.5), although it is presently unknown if the CMIP climate models can realistically simulate GSAs.

Summarizing:

- The CMIP models consistently predict an increasing hydrological cycle with greater precipitation, evaporation, and runoff, polewards of 70°N by 2100 (section 4.1; *Vavrus et al.* 2012; *Bintanja and Selten* 2014). Sea ice extent and volume are projected to decrease, with large variability between models (section 4.1) and loss rates significantly lower than observations. The total liquid freshwater volume is projected to increase by about 50000 km³ between 2000 and 2100 (Fig. 9). Liquid freshwater in the Beaufort Gyre will likely also increase, although there is significant variability among models (Fig. 8).
- The best evidence to date on climate projections of marine freshwater fluxes comes from the CCSM4 model (*Vavrus et al.*, 2012). In CCSM4, Bering and Fram Strait liquid freshwater fluxes increase (section 4.2). The CAA liquid flux increases to 2070 then declines thereafter in this model. Sea ice export through Fram Strait declines substantially through the 21st century.

To conclude this section on freshwater prospects refer to the final column in Table 1. The prospects for the freshwater budget for the 21st century are quantified where possible based on the CMIP climate models. Each value derives

from numbers quoted in section 4²⁰. Ignoring changes in the minor components (which have question marks in Table 1), we anticipate that the sources of freshwater to the Arctic will increase, from about 9400 to perhaps 11000 km³yr⁻¹, by 2100²¹. The sinks of freshwater draining the Arctic will also likely increase, from about 8250 to perhaps 10000 km³yr⁻¹. These numbers indicate that the Arctic freshwater cycle will accelerate in the 21st century with significantly increasing inflow, outflow, and storage of freshwater. It is likely that the freshwater budget in 2100 will not be balanced: the freshwater sources will probably exceed the sinks and the Arctic will continue freshening. These estimates are provisional and uncertain, as discussed above.

5. Conclusions

This paper has reviewed published literature on the status, mechanisms, and prospects for freshwater, and especially freshwater fluxes, in the Arctic and Subarctic Ocean. Where possible, we have synthesized these prior works. The main findings are:

- Freshwater is accumulating in the Arctic, CAA, and Baffin Bay (section 2.2): about 8000 km³ more freshwater was present in the decade of the 2000s compared to the 1980–2000 average (Table 1). Accumulation is mainly in the Beaufort Gyre, where the increase was about 5000 km³.
- Sea ice extent, volume, and age have decreased in the 2000s compared to 1980–2000 (section 2.3, Fig. 3).

²⁰To estimate the runoff increase we take *Vavrus et al.*'s (2012) 40% precipitation increase and reduce it to 30% based on Fig. 1a of *Bintanja and Selten* (2014), which shows precipitation over land increases less than over the ocean. The P-E increase is estimated as 300 km³yr⁻¹ from *Bintanja and Selten*'s (2014) Fig. 2a (inset). The entry for the Greenland flux is based on extrapolating *Bamber et al.*'s (2012) estimates (see section 2).

²¹This estimate excludes changes in the Bering Strait inflow, which increases in CCSM4 but has not been quantified (*Vavrus et al.*, 2012).

- 1311 • The meteoric fluxes supplying freshwater (runoff and precipitation) have
1312 increased in the 2000s compared to 1980–2000 (section 2.4, Table 1, Fig. 4;
1313 most of the evidence comes from models). Despite flux increases from 2001
1314 to 2011, it is uncertain if the marine freshwater source through Bering
1315 Strait has changed, as observations in the 1980s and 1990s are incom-
1316 plete. The total marine flux draining freshwater (liquid and as ice through
1317 Fram and Davis straits) have not changed significantly. The net flux of
1318 freshwater has therefore increased, to about $1200 \pm 730 \text{ km}^3\text{yr}^{-1}$.
- 1319 • The observed increase in liquid freshwater storage in the 2000s is consistent
1320 with the shift in freshwater fluxes and the loss of freshwater as sea ice,
1321 although the uncertainty is large (Fig. 7).
- 1322 • Understanding of the mechanisms controlling Arctic freshwater fluxes and
1323 storage points to the importance of the surface wind field (sections 3.1.2–
1324 3.3, Fig. 6). The wind controls the surface ocean circulation (Fig. 5) and
1325 hence freshwater transport rates and pathways (section 2.5).
- 1326 • The characteristic timescale for changes in the freshwater system due to
1327 source or sink changes is about 15 years (section 3.1.1). The timescale
1328 for export flux changes driven by the wind is much shorter, perhaps $O(1$ –
1329 $10)$ months (section 3.4). Because the wind controls these changes, they
1330 are less predictable than those caused by variability of freshwater sources.
1331 Large freshwater export events, Great Salinity Anomalies (GSAs, section
1332 3.5), have been observed in the last 50 years, probably triggered by changes
1333 in the Arctic surface winds.
- 1334 • Although inherently uncertain, coupled climate models simulate Arctic
1335 freshwater processes in several realistic ways (section 4, Fig. 9). Their
1336 predictions for the 21st century show continued acceleration of the hydro-
1337 logical cycle, with roughly an extra 50000 km^3 liquid freshwater stored by
1338 2100 (section 4.3, Table 1). Climate models predict that the marine ex-
1339 port fluxes of liquid (ice) freshwater will increase (decrease) enough to be

1340 detected by the export monitoring arrays. They underestimate the speed
1341 of sea ice decline, however. Also, it is unclear if they capture GSA mech-
1342 anisms, and therefore may be incapable of simulating rapid freshwater
1343 discharge events.

1344 • The impacts of these changes in the Arctic freshwater system are diverse.
1345 They include effects within the Arctic ocean, such as albedo and upper-
1346 ocean stratification changes, which in turn may affect the heat budget,
1347 mineral nutrient supply to phytoplankton, and the light environment near
1348 the surface. And they reflect the view that climate change in the Arctic
1349 is amplified. A thorough review of these impacts is beyond the current
1350 scope, but would be interesting and valuable (for example, see *Bhatt et al.*
1351 2014 on implications of sea-ice decline).

1352 Future work on Arctic freshwater should continue to focus on the gateway
1353 fluxes through straits. Although no significant changes in export fluxes have yet
1354 been seen, it is likely they will occur, perhaps suddenly, in response to changes
1355 in Arctic wind. Future work should maintain the hydrographic sampling of the
1356 Arctic ocean to determine freshwater storage changes. Chemical tracers are
1357 essential too, in order to distinguish different freshwater origins and pathways.
1358 The mechanisms discussed in section 3 are valuable because they provide a ba-
1359 sis to test and refine the coupled climate models discussed in section 4, and
1360 discriminate between them. Understanding the freshwater processes in these
1361 models is another priority, as is examining their freshwater budgets in detail,
1362 for example using the framework of section 2. Future work to deepen under-
1363 standing of the mechanisms controlling freshwater accumulation and release will
1364 potentially aid in observing strategies. For example, processes controlling sea
1365 level are important because sea level differences are linked to volume fluxes and
1366 hence freshwater fluxes, especially west of Greenland. Finally, it seems likely
1367 that many Arctic freshwater mechanisms will change with the impending loss
1368 of summer sea ice. They include some of the processes that maintain the fresh
1369 basic state of the upper ocean (section 3.1.1). Anticipating, observing, and

1370 understanding those changes is an unprecedented opportunity that will further
1371 elucidate the dynamics of the Arctic freshwater system.

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