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## Current estimates of freshwater flux through Arctic and subarctic seas

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### Abstract

As the world warms, the expectation is that the freshwater outflows from the Arctic Ocean to the North Atlantic will strengthen and may act to suppress the rate of the climatically-important Atlantic meridional overturning circulation. Hitherto, however, we have lacked the system of measurements required to estimate the totality of the freshwater flux through subarctic seas. Though observations remain patchy and rudimentary in places, we piece-together the results from recent large-scale observational programmes together with associated modelling, to establish preliminary maps of the rates and pathways of freshwater flux through subarctic seas. These fluxes are calculated according to two reference salinities,  $S = 34.8$  to conform with the majority of estimates reported in the literature, and  $S = 35.2$ , the salinity of the inflowing Atlantic water, to calculate the freshwater balance of the 'Arctic Mediterranean'. We find that 148 mSv of freshwater enters the Nordic Seas across its northern boundary. There it is supplemented by around 54 mSv of freshwater from Baltic runoff, Norwegian runoff, P – E and Greenland ice melt, so that the total freshwater contribution to the Nordic Seas from all sources is 202 mSv. Of this, around 51 mSv of freshwater is estimated to pass south to the deep Atlantic in the dense water overflows leaving an assumed balance of 151 mSv to leave the Nordic Seas in the upper water export through Denmark Strait. The corresponding estimate for the freshwater outflow west of Greenland is 103 mSv relative to 35.2 so that the total freshwater flux reaching the North Atlantic through subarctic seas is around 300 mSv.

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**Acronyms**

ADCP	acoustic doppler current profiler
ASOF	Arctic-Subarctic Ocean Fluxes study
A-W-I	Alfred-Wegener-Institute for polar and marine research
AR7W	Atlantic repeat hydrography section 7 west
BGC	Bristol Glaciology Centre
BIO	Bedford Institute of Oceanography
BSH	Bundesamt für Seeschifffahrt und Hydrographie
CAA	Canadian Arctic Archipelago
CEFAS	Centre for Environment, Fisheries and Aquaculture Science, Lowestoft
DAMOCLES	EC integrated project on Developing Arctic Modelling and Observing Capabilities for Long-term Environmental Studies.
ECHAM-5	5th generation of an ECMWF climate model developed in Hamburg by Max-Planck-Institut für Meteorologie, Deutsches Klimarechenzentrum, and Meteorologisches Institut der Universität Hamburg
ECMWF	European Centre for Medium Range Weather Forecasting
EC-FP6	Framework Programme 6 of EC
ERA-40	40-year reanalysis data set by ECMWF covering the period 1957–2002
HADCM3	coupled atmosphere–ocean general circulation model developed at the Hadley Centre
IMR	Institute of Marine Research
IOS	Institute of Ocean Sciences
IPCC SRES	Intergovernmental Panel on Climate Change Special Reports on Emission Scenarios
MOEN	Meridional Overturning Exchange with the Nordic Seas; a subset of ASOF
MOC	meridional overturning circulation of the North Atlantic
M-P-I OM	Max-Planck-Institute for Meteorology, Ocean Model
MRI	Marine Research Institute
NAO/AO	North Atlantic Oscillation/Arctic Oscillation
NAOSIM	hydrodynamic coupled ice-ocean model designed by A-W-I and OASys to examine the variability of ice and ocean systems in the Arctic/subarctic domain.
NCC	Norwegian Coastal Current
NPI	Norwegian Polar Institute
NPS	Navy Postgraduate School Monterey
OASys	Ocean Atmosphere Systems
P – E	precipitation minus evaporation
UoB	University of Bergen
UCL-CPOMS	University College London, Centre for Polar Observation and Modelling
UHH	University of Hamburg
UW	University of Washington
VEINS	Variability of Exchanges In Northern Seas
WHOI	Woods Hole Oceanographic Institution

**1. Introduction: budgets, change and ‘system’ in subarctic seas**

In their landmark report, [Aagaard and Carmack \(1989\)](#) provided the first modern and complete accounting of the freshwater budget of the Arctic Ocean, the regional variations in the storage of freshwater within the Arctic Ocean itself and the two-way exchanges of freshwater with subarctic seas. However, budgets change and in view of their likely climatic significance, it is important that we develop an improved understanding of the magnitude of freshwater exchanges between the Arctic and North Atlantic and the extent to which these have varied or are expected to vary. This brief review has the simple aim of collating the many new estimates

of freshwater flux, calculated with respect to a common reference salinity, in order to provide an updated assessment of the total freshwater flux passing through subarctic seas.

Apart from the freshwater that enters as runoff and precipitation, any attempt to estimate the freshwater transports through our northern seas in a comparative way presupposes that we can select a reference salinity that is of some general applicability. Here, we use two such reference values. In reviewing the historic point-estimates of freshwater flux, these are expressed *relative to 34.8*, largely because – following Aagaard and Carmack (1989) – this is the estimated mean salinity of the Arctic Ocean and the reference most commonly adopted and reported in the literature.

In calculating the freshwater balance of the Arctic Mediterranean however (the Nordic Seas and Arctic Ocean), we employ a reference salinity of 35.2. As the northernmost extension of the North Atlantic, the water masses of this region derive mainly from Atlantic water that crosses the sill of the Greenland–Scotland Ridge. This warm saline inflow becomes progressively diluted by mixing with freshwater, from river runoff, from net precipitation and from ice melt. Freezing and brine release can locally lead to a salinity increase but the waters leaving the Arctic Mediterranean are less saline than the inflowing Atlantic water. In forming a freshwater balance, therefore, the natural reference salinity to choose is the salinity of the inflowing Atlantic water, i.e. 35.2 (Hansen and Østerhus, 2000). This has the advantage that the freshwater and the volume transports are consistently in the same direction for all individual streams. It also permits us to calculate the transport of freshwater out of the Arctic Mediterranean in the deep overflows. On the negative side the freshwater transport will in some instances be opposite to that derived using the 34.8 reference.

Whichever reference is applied, the fact that the Arctic–subarctic system is both under-sampled and rapidly-changing is likely to prove problematic. Put simply, the large variability in the supply, storage and transfer of freshwater throughout this system may prevent us from discriminating inconsistency from change in our estimates. Thus Rothrock and Zhang (2005) correctly point out that there is no such thing as a ‘steady state’ for sea-ice mass, that different estimates may simply reflect temporal differences, and the same should also hold true for any imbalances that we may detect when establishing freshwater budgets. The same might be said of undersampling, even in the case of our longest time-series. Thus in reporting larger variations of inflow in the Shetland branch than in the Faroe branch of inflow, Østerhus et al. (2005) admit that ‘*It is not clear whether this reflects reality or indicates differences in the precision of the estimates*’.

Temporal change may also act to undermine the basis for our choice of reference salinity, as it has certainly done in the present case. Though a value of 35.2 had seemed soundly-established as an appropriate reference value for the Atlantic inflow to the Arctic Mediterranean (Hansen and Østerhus, 2000), the continued salinification of these Atlantic water branches to a record maximum in 2003 (Hansen et al., 2004) has meant that 35.25 would have been a more appropriate mean for the three inflow branches in recent years with an even larger mean value (>35.3) in the case of the Shetland branch (Østerhus et al., 2005).

Despite such uncertainties, the positive justification for this task lies in the fact that the ocean exchanges connecting the Arctic and North Atlantic do have identifiable full-latitude ‘system’, and that coherent large-amplitude changes are increasingly observed (or simulated) to pass through and modulate that system on decade-to-century time scales. The reality of this statement is argued by a wide diversity of recent studies: for example, in investigating the competition between ocean dynamics and ice formation/melt on the Arctic basin freshwater balance, Hakkinen and Proshutinsky (2004) find that ‘*changes in the Atlantic water inflow can explain almost all of the simulated freshwater anomalies in the main Arctic basin*’. In their model experiments using ECHAM5 and the MPI-OM, Jungclauss et al. (2005) show that while ‘*the strength of the overturning circulation is related to the convective activity in the deep water formation regions, most notably the Labrador Sea, ... the variability is sustained by an interplay between the storage and release of freshwater from the central Arctic and circulation changes in the Nordic Seas that are caused by variations in the Atlantic heat and salt transport*.’ From their study using HadCM3, Wu and Wood (submitted for publication) conclude that the recent observed freshening trend of subpolar seas may be due not so much to large-scale changes in the surface freshwater flux but to a redistribution of freshwater due to ocean circulation changes in the subarctic seas that are ultimately triggered by deep convection in the Labrador Sea. In their careful analysis of the historic hydrographic record, Curry and Mauritzen (2005) show that an estimated 19,000 km<sup>3</sup> of ‘extra’ freshwater was brought south to freshen the watercolumn of the Nordic Seas and NW Atlantic (the ‘headwaters’ of the global thermohaline circulation) between the mid 1960s and late 1990s. And west of Greenland, Zweng

and Münchow (2006) show that areas affected by Arctic inflow to Baffin Bay have shown a ‘marginally significant freshening of up to about  $0.086 \pm 0.039$  psu per decade’ over most of the past century (see also Houghton and Visbeck, 2002). In each of these cases, it seems clear that the full-latitude ‘system’ of Arctic and subarctic seas is at work. It seems equally plain therefore that explaining the dynamics – and possibly the interactions – of these linkages may hold valuable clues to understanding the role of the northern seas in climate.

## 2. Freshwater inputs, and fluxes through subarctic seas relative to $S_{\text{ref}} = 34.8$

### 2.1. Arctic Ocean Sea-ice

The question of long-term thinning, redistribution and retraction of Arctic Sea-ice has been extensively discussed in the literature for example, Rothrock et al. (1999), Johannessen et al. (1999), Wadhams and Davis (2000), Holloway and Sou (2002), Laxon et al. (2003), Rothrock et al. (2003), Comiso (2002) and Sroeve et al. (2005). The most recent of these (Stroeve et al.) shows that the downward trend in ice extent at the annual minimum (September) has been  $-7.7 \pm 3\%$  per decade since 1979 with evidence of a reinforced downturn in the most recent years. Applying the mean thickness quoted by Laxon et al. (op cit), this amounts to a loss rate of  $\sim 3$  mSv over the last decade.

### 2.2. River inputs

Though only 1% of the world ocean by volume, the Arctic Ocean receives 11% of world river runoff (Shiklomanov et al., 2000). Following the general relation between catchment area and discharge (Prowse and Flegg, 2000) the total river inflow to the Arctic Ocean is dominated by contributions from Eurasia (Peterson et al., 2002; Lammers et al., 2001), with six rivers (Yenesei, Lena, Ob, Pechora, Kolyma and Severnaya Dvina) draining 2/3rds of the Eurasian Arctic landmass. Despite a widespread decline in hydrological monitoring since 1986 (Shiklomanov et al., 2002), Peterson et al. still conclude that the average annual discharge from these six rivers has increased by 7% from 1936 to 1999, equivalent to an increase in discharge rate from 58 mSv to 62 mSv (‘RR’ in Fig. 1). The increase is linked both to Arctic warming (empirically determined to be  $+7$  mSv per  $^{\circ}\text{C}$ ) and to the amplifying NAO/AO since the 1960s. HadCM3 has proven successful in simulating many aspects of observed 20thC climate change including the trend in Arctic sea-ice (Gregory et al., 2002) and the freshening trend in subpolar seas (Wu et al., 2004). In the most recent of these ensemble simulations, Wu et al. (2005) report an attempt to reproduce the trend in Arctic riverflow using an ‘all forcings simulation’ in which the model was forced with both anthropogenic (greenhouse gases, sulphate aerosols and ozone) and natural (solar and volcanic) external factors. The simulated change in Arctic river flow shows a consistent upward trend from the 1960s, ‘in very good agreement’ both in timing and amplitude with the trend reported by Peterson et al. (2002) from river monitoring data. The trend is strong in the ensemble simulation with only anthropogenic forcings but is not evident in the parallel set of experiments using only natural forcings. Wu et al. therefore describe it as likely that the upward trend in circumarctic riverflow is real, part of an anthropogenic intensification of the global hydrological cycle.

A value for North American runoff (‘AR’) in Fig. 1 may be obtained simply by subtracting the Peterson et al. mean Eurasian river discharge of  $\sim 60$  mSv from a modern value of circumarctic total runoff. Serreze et al. (2006) provide such a value, reviewing estimates based on aerological P – E over land, river-gauge data supplemented by hydrological modelling and two land surface models driven by ERA-40 outputs to suggest that a value of 95 mSv for total runoff seems soundly-based. This sets the N American contribution very approximately at 35 mSv. Dery and Wood (2005) provide a very similar estimate of 40 mSv based on the measured discharge from 64 Canadian rivers between 1964 and 2003 (adopted in Fig. 1). They report a 10% downward trend over this period which is at variance with the increasing trend reported for Eurasian rivers by Peterson et al. (2002), most of which occurred since the 1960s. If real, such a geographic difference in trend might be thought to reflect the influence of the amplifying NAO rather than anthropogenic change (see Section 5 below).

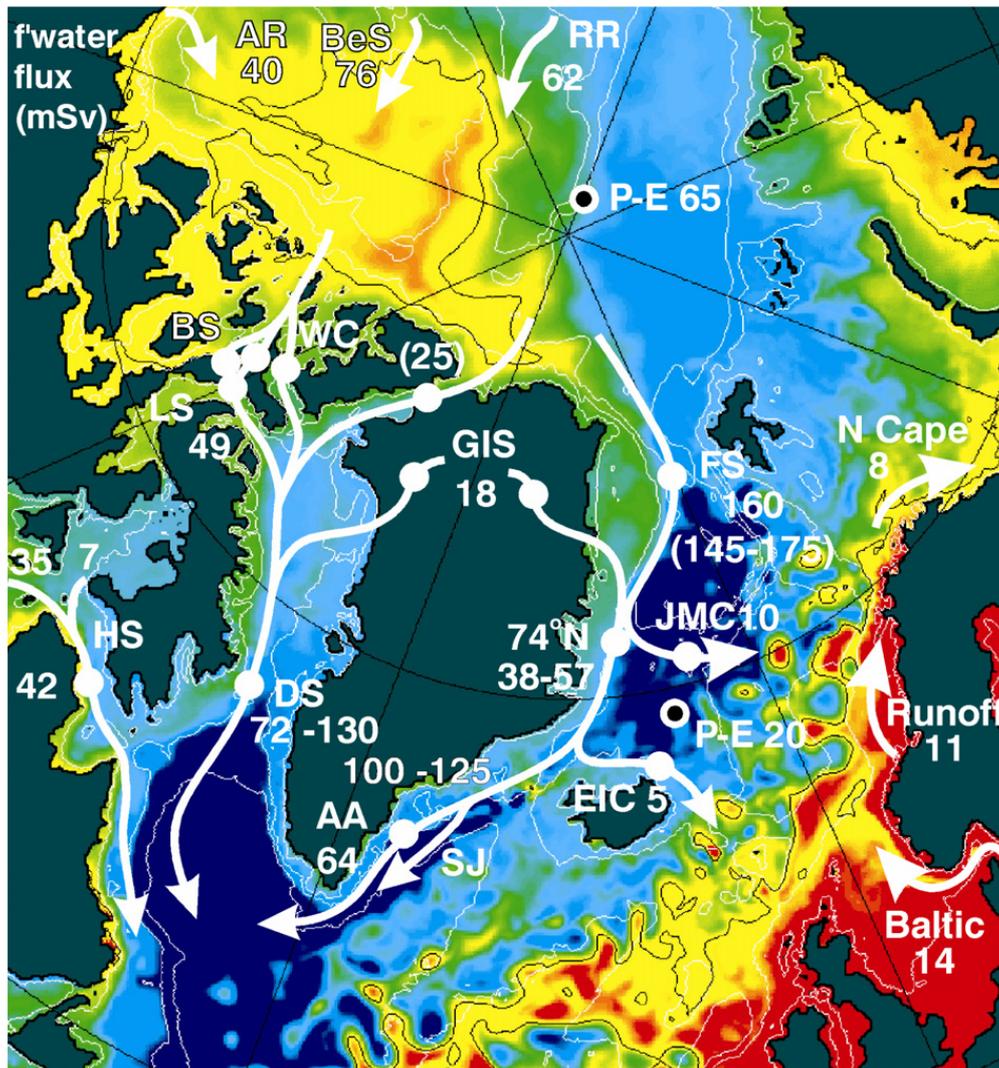


Fig. 1. Estimates of freshwater flux relative to  $S = 34.8$  in Arctic and Subarctic Seas. Units are mSv and the base map is a snapshot of sea surface height from the NPS PIPS 3.0 model kindly provided by Wieslaw Maslowski, Navy Postgraduate School, Monterey. [ $1 \text{ mSv} = 31.546 \text{ km}^3 \text{ y}^{-1}$ ; \* the numbers for P – E, runoff and ice melt are independent of the choice of reference salinity].

Completing the inventory of freshwater inputs to the Arctic Ocean, Fig. 1 includes a value of 65 mSv for P – E based on the modern accounting of Serreze et al. (2006); we note that this is substantially different from the value (29 mSv) derived by Aagaard and Carmack (1989).

### 2.3. Bering Strait

Aagaard and Carmack (1989) estimated the Bering Strait freshwater flux as 53 mSv relative to 34.8, assuming an annual mean transport of 0.8 Sv and salinity of 32.5. Moorings have now been deployed in the Bering Strait since 1990 at two or more locations, to monitor flow through its western and eastern channels. The results [see <http://psc.apl.washington.edu/HLD/Bstrait/bstrait.html>] indicate a pronounced annual cycle in both temperature and salinity with an average peak-to-peak amplitude of salinity variation of about 1.5. Although long-term moored measurements and ship-based observations confirm a volume flux of around 0.8 Sv, Woodgate and Aagaard (2005) now conclude that the earlier values of freshwater flux had been seriously underestimated. By including a contribution of  $\approx 400 \text{ km}^3 \text{ y}^{-1}$  from the fresh Alaskan coastal current, and a further  $\approx 400 \text{ km}^3 \text{ y}^{-1}$  from the effects of seasonal stratification and ice transport, the total freshwater flux relative to 34.8 is raised to about 76 mSv, with a suggested uncertainty of  $\pm 10 \text{ mSv}$ .

#### 2.4. Canadian Arctic Archipelago throughflow

The early estimates of the transports through the Canadian Arctic Archipelago were given as salt transports (Aagaard and Greisman, 1975; Stigebrandt, 1981; Rudels, 1986). Translated into freshwater fluxes relative to 34.8 they ranged between 36 and 63 mSv. Direct freshwater transports were given by Aagaard and Carmack (1989), Steele et al. (1996), Ingram and Prinsenberg (1998) and by Loder et al. (1998). Understandably since we are dealing with vigorous flows in a remote complex of narrow ice-covered passageways where the scales of motion are small, where moving ice and icebergs pose a hazard to moored gear and where proximity to the magnetic pole complicates even the measurement of flow direction, comprehensive simultaneous coverage has been difficult to achieve. Nonetheless pioneering direct measurements have been recovered from the main passageways of the CAA, since 1998 by Prinsenberg in the case of Lancaster Sound and Barrow Strait, by Melling in the case of Wellington Channel, Cardigan Strait and Hell Gate, and by Falkner, Melling and Münchow (currently) in Nares Strait/Kennedy Channel. On the basis of these direct measurements, Melling (2004) and Prinsenberg and Hamilton (2005) provide up-to-date accounts of the volume and freshwater fluxes through the CAA. Giving particular weight to the eastbound flow through the south side of their array, Prinsenberg and Hamilton (2004, 2005) calculate a 3-year mean freshwater flux through Lancaster Sound of  $45 \pm 15$  mSv relative to 34.8 (recently adjusted to 48 mSv plus an ice-flux of 1.3 mSv using the extended record to 2004; Prinsenberg pers. comm. 2006). They then extend that calculation by comparing the direct estimate with simulations performed by a regional model of the CAA (Kliem and Greenberg, 2003; Greenberg and Kliem, 2004) and at the Naval Postgraduate School (NPS; Maslowski et al., 2000, 2003). The Kliem and Greenberg (2003) model indicates that fluxes through Lancaster Sound make up 35% of the freshwater fluxes through the total CAA, while with different grid size and processes, that of Maslowski (2003) suggests that the flux through Lancaster Sound can be 50% of the total. Prinsenberg and Hamilton therefore suggest a value between 90 and 110 mSv for the freshwater flux through the Archipelago as a whole which they point out is similar to present values in the literature (Melling, 2000 and Prinsenberg and Bennett, 1987). The flow through the Straits seems related (Greenberg and Kliem) to the elevation difference across the CAA between the Arctic Ocean and Baffin Bay; the 10 cm increase in sea-level slope along the NW Passage in summer is consistent with the seasonal variability of the flow, strong in summer, weaker in winter (Prinsenberg and Bennett, 1987).

Provision of the first long-term direct measurements of freshwater and mass fluxes through Nares Strait will await the recovery of the present comprehensive array of bottom-mounted ADCPs, conductivity–temperature strings and pressure recorders. Meanwhile, Münchow et al. (2006) have used high resolution ship-based ADCP and hydrographic casts obtained during surveys of the Strait in August 2003 to calculate ‘snapshot’ values of the volume and freshwater fluxes through the Strait of  $0.8 \pm 0.3$  Sv and  $25 \pm 12$  mSv, respectively. As these authors strongly emphasise, these estimates are representative of only a few days of observation without any understanding of the scale or dynamics of their forcing. Their freshwater flux estimate is included, bracketed, in Fig. 1.

#### 2.5. Davis Strait

Intercepting the CAA freshwater outflow further south, Loder et al. (1998), Tang et al. (2004) and Cuny et al. (2006) provide estimates of 120 and 130 mSv for the annual mean freshwater flux passing south through the Ross Moored Array at 66°N in Davis Strait in the late 1980s (Ross, 1992). Making allowance for a north-going component on the West Greenland Shelf, Cuny et al. suggest that the net southward flux through the Strait reduces to 92 mSv. Maslowski (2003, and pers. comm.) later modelled the gross southgoing freshwater flux through the Davis Strait as 76 mSv (65 mSv net) over the period 1979–2001, in reasonable agreement with the observation-based estimates just described. Though first results from the comprehensive SeaGlider- and mooring-based advanced monitoring array across the Davis Strait by Craig Lee of U Washington have yet to be analysed, freshwater fluxes of 72, 102, and 115 mSv between the 400 m isobaths are reported for the first three SeaGlider transects of Davis Strait in 2004–5 (Lee, 2006 pers comm.). As short-term snapshots, the latter are likely to be more variable than the annual mean values reported for the 1980s. Nonetheless, bracketing all of these studies, a provisional range of 72–130 mSv is used in Fig. 1.

## 2.6. Hudson Bay and Strait

Ingram and Prinsenberg (1998) describe the transport of northern source water from the CAA to Hudson Bay via Fury and Hecla Strait and Foxe Basin as small – with a total transport of 0.04 Sv in late winter, 0.1 Sv in summer. The freshwater component of this small inflow is estimated by Straneo (pers. comm. 2006) to be 7 mSv. Regarding the freshwater outflow through Hudson Strait, Saucier et al. (2004) used a 3-D coastal ice-ocean model with 10 km horizontal resolution and with ‘realistic’ tidal, atmospheric, hydrologic and oceanic forcing to propose a mean annual freshwater flux (relative to 34.8) to the Labrador Sea of 35 mSv, 29 mSv in liquid form +6 mSv as ice; this they describe as consistent with the freshwater fluxes at the atmospheric and land boundaries of their model domain (see Prinsenberg, 1977, 1988). The first collaborative direct measurements of the outflow through the Strait have now been recovered by Straneo (WHOI) and by Saucier’s Group at University of Quebec suggesting a higher estimate of 57 mSv for the freshwater efflux along the southern side of the Strait. This is certainly where we may expect the outflowing mix of ice melt, runoff and Pacific water to be concentrated (Ingram and Prinsenberg, 1998) but by including an unknown import from the Baffin Current is likely to provide an overestimate of the flux. Until the full Strait is instrumented therefore (underway), a figure of 42 mSv is suggested as realistic for the net freshwater flux (Straneo, pers comm. 2006), 35 mSv from the river discharge through Hudson Bay +7 mSv from the Canadian Arctic Archipelago through Fury and Hecla Strait; this is the estimate used here (Fig. 1).

## 2.7. Labrador margins

The AR7W repeat hydrographic section from the Labrador shelf to the southern tip of Greenland provides us with our only indication of recent trends in the transport and character of fresh near-surface (0–150 m) waters passing south around the western margins of the Labrador Sea. There (see Dickson et al., 2003) we observe a 40-year increasing trend in the offshore density gradient between the Labrador shelf and upper slope, roughly in parallel with the increase in the NAO Index. In fact, in the absence of direct current measurements, we are unable to determine whether this trend is the result of a 20% increase in south-going transport, some cross-slope change in the hydrographic character of the flow or both. If based on such widely varying assumptions, the resulting estimated change in freshwater transport would be understandably very wide (22–127 mSv) – too wide to be worth including in Fig. 1 – but even such a poorly based estimate makes the point that the freshwater flux around the western margin of the Labrador Sea may be large, variable and may act as a partial bypass of the Labrador–Irminger Basin [See below under ‘Labrador Sea’]. The degree of exchange between the Hudson Strait outflow and the Labrador Current is likewise unknown.

## 2.8. Fram Strait

Recent annual estimates for the efflux of sea-ice through western Fram Strait range from 70 mSv (Kwok et al., 2004), through 76 mSv (Widell et al., 2003) to 92 mSv (Vinje, 2001) for the 1990s. Though these results are consistent, all use the same in situ data (Østerhus pers comm.) and may have the same offset. Earlier, Aagaard and Carmack (1989) reported 88 mSv for the ice-flux. Thus a mean of 82 mSv for the 4 observation-based estimates of annual sea-ice-flux, which is in reasonable agreement with the value of 96 mSv for the 1990s mean efflux of ice through Fram Strait in the NAOSIM model (Karcher et al., 2005).

The liquid freshwater flux passing 79N during the 1990s is still an open question. Actual estimates of the freshwater component range upwards from 20 mSv from a version of the NPS model at 9 km resolution (Maslowski, pers comm.), through a hydro-based estimate of 28 mSv by Aagaard and Carmack (1989), and a preliminary\* mooring-based estimate of 32 mSv through 79N by Holfort and Hansen, 2004 [but relative to a salinity of 34.9] to values of 63–95 mSv derived by Meredith et al. (2001) from two years of  $\delta^{18}\text{O}$  measurements. Estimates of freshwater flux from the NAOSIM model are almost identical to the latter values at 78 mSv for the 1990s and a mean of 65 mSv for the entire 1948–2002 period, with reference salinity 34.8 (Karcher et al., 2005). [\*Holfort and Hansen point out that their estimate is biased high since they use the low near-surface salinity values from summer, and low since they miss part of the barotropic velocity compo-

nent; NAOSIM also suggests that 2/3rds of the freshwater flux may pass south on the East Greenland Shelf where the mooring array lacks coverage].

We might expect the estimates of the liquid freshwater component to be diverse, strongly time- and depth-dependent and subject to a large but unknown recirculation. We also note that the dominant pathway for the outflow of freshwater from the Arctic is Fram Strait in the NAOSIM model and the Canadian Archipelago in the NPS model (Maslowski, 2003, and pers. comm.). It remains to be investigated which factors determine the relative importance of each pathway and if those factors are time-dependent (see discussions in ‘Modelling’ and the ‘Future’).

Jonsson (2003) based his estimate of 125 mSv for the total freshwater flux leaving Fram Strait largely on the Aagaard and Carmack (1989) estimate. Here we place the greater weight on the later estimates in assuming a total freshwater flux of approx 160 mSv (80 mSv as ice +65–95 mSv as freshwater).

### 2.9. Freshwater production from the Greenland ice sheet

The Greenland ice sheet (GIS) is the largest single freshwater reservoir in the northern hemisphere. At present Vanden Broeke (University of Utrecht) and co-workers estimate that the GIS contributes an annual freshwater flux of 18 mSv (of which 10 mSv is through ablation, 8 mSv via calving of icebergs), which they describe as small but non-negligible compared with the discharge of the six largest Eurasian rivers (60 mSv). HADCM3 derives an essentially similar figure for the net freshwater loss from Greenland (pptn minus ablation = 17.4 mSv; Gregory and Lowe, 2000). Jeff Ridley’s analysis (pers comm.) based on HadCM3 puts the flux at 23 mSv, 13 mSv from melt and 10 mSv from calving under present day conditions (1× CO<sub>2</sub> scenario). These estimates of calving are essentially identical to the total ice discharge of 11 mSv measured by Rignot and Kanagaratnam (2006) using satellite radar interferometry. Thus although it is anticipated by Fichfet et al. (2003) that the annual mean total freshwater flux from Greenland will increase by 15 mSv between 1970 and 2080, based on a mix of modelling techniques (see ‘Future’, Section 5 below), a value of 18 mSv seems appropriate to the present annual freshwater flux from Greenland and is used in Fig. 1.

### 2.10. East Greenland Current at 74N

Holfort and Meincke (2005) provide a first measure of the total freshwater flux (ice and liquid water) passing south along the outer part of the East Greenland shelf at 74N. From a mix of direct current profiles by bottom-mounted ADCP and continuous year-long salinity records from Microcat sensors mounted within so-called ‘tube moorings’, they calculate a total freshwater transport of 40–55 mSv in 2001–2002, and 38–57 mSv in 2003–4, with the transport in the form of ice being as large as that of freshwater, though with an opposite seasonality (maximum ice transport in winter).

### 2.11. Jan Mayen Current

From the literature, Jonsson (2003) estimates the diffuence of freshwater from the East Greenland Current into the Jan Mayen Current as 10 mSv or only 8.8% of his estimate for the flux through Fram Strait (above).

### 2.12. East Icelandic Current

The equivalent figure for the diversion of freshwater from the East Greenland Current into the East Icelandic Current has been estimated from Icelandic standard hydrographic sections (Langanes mainly) and direct current measurements as only 5 mSv (Jonsson and Briem, 2003).

### 2.13. Denmark Strait

Based on the Icelandic Standard Section program and moored sensors, Jonsson (2003) estimates the liquid freshwater flux passing Denmark Strait as 108 mSv, equivalent to 87% of his assumed flux leaving Fram Strait but now also incorporating a component of Greenland runoff. The A-W-I NAOSIM model gives the equiv-

alent of 39 mSv for the mean ice-flux plus 86 mSv for liquid freshwater relative to 34.8, hence a total freshwater flux of 127 mSv for the period 1948–2002 (Karcher et al., 2005).

#### 2.14. SE Greenland Shelf and ‘Spilljet’

Though a prototype freshwater flux array across the SE Greenland Shelf is being developed and extended by UK, German and Finnish collaborators within ASOF-EC (W) and has so far returned salinity time-series of up to 5 years duration from its ‘tube moorings’, it has so far lacked the ADCP or current meter data required to complete a flux estimate. Based on the 4-month period with the most complete data set – two ‘tube moorings’ carrying 6 SBE-37 Microcat salinity sensors plus Valeport current meter coverage in late summer 2001 –, Dye and Holfort (pers comm.) provide a preliminary and partial estimate of 64 mSv for the liquid freshwater flux passing south through this part of the SE Greenland shelf. Questions of seasonality, longer-term change and an apparent seasonal encroachment of the shelf-edge front will present major unknowns until further expansions of the array can be made in both space and time (now planned under the DAMOCLES Integrated Project of EC FP6). No estimate of freshwater transport by the so-called ‘spilljet’ of the upper Slope has yet been made (Pickart et al., 2005) and where the spilljet has a salinity above 34.8, it would not in any case contribute to this accounting unless the higher reference salinity is employed (see below). [Further south at Cape Farewell, Bacon et al. (2002) provide a first snapshot estimate of freshwater flux through the narrowest part of the S Greenland Shelf; their estimate is 60 mSv but relative to an ‘Atlantic mean salinity’ of  $S_{\text{ref}} = 34.956$ ].

#### 2.15. Baltic

Fig. 1 also provides a preliminary estimate for the freshwater contributed by Baltic outflow to northern seas, together with that due to runoff from Scandinavia, and after losses en route, the net amount that rounds North Cape in the Norwegian Coastal Current. The basis for these estimates is as follows: following Bergstrom and Carlsson (1994) we take the annual mean runoff to the Baltic to be  $\approx 14 \text{ mSv} \pm 15\%$ ; since the mean  $P - E$  for the Baltic is essentially zero ( $0 \pm 50 \text{ mm y}^{-1}$ ), the net freshwater discharged to the Skagerrak is also taken to be  $\approx 14 \text{ mSv}$ . The figure of 11 mSv for Scandinavian runoff is due to Mosby (1962). From Blindheim’s (1989) figures for the mean transport and salinity of the North Cape Current (0.7 Sv and 34.4, respectively) we derive a freshwater transport in the NCC of 8 mSv relative to 34.8, which was the figure adopted by Aagaard and Carmack (1989).

#### 2.16. Labrador Sea

The Labrador Sea is a critical location for the Earth’s climate system. In its upper and intermediate layers, annual-to-decadal variations in the production, character and thickness of Labrador Sea water (LSW) directly determine the rate of the main Atlantic gyre circulation (Curry and McCartney, 2001). Through its deeper layers pass all of the deep and bottom waters that collectively form and drive the abyssal limb of the Atlantic MOC. Around its margins pass the two main freshwater flows from the Arctic Ocean to the North Atlantic via the Canadian Arctic Archipelago and East Greenland Shelf which have been implicated in model experiments with slowing down the MOC. Over the past 3–4 decades, affected both by an increased freshwater transport around the Labrador margins and by a freshening of the dense overflow system (Dickson et al., 2002, 2003), the entire watercolumn of the Labrador Sea has undergone radical change. From 1966 to 1992, the overall freshening of the watercolumn of the Labrador Sea has been equivalent to mixing-in *an extra 6 m of freshwater* at the sea surface (Lazier, 1995; and pers comm.), part of the largest full-depth change ever observed in the modern instrumental oceanographic record. The Labrador Sea is the main receiving volume in which ocean fluxes from northern seas are stored, recirculated, transformed and discharged, and it is of interest in the present context to determine the total change in freshwater ‘received’ in relation to the freshwater ‘lost’ from northern seas. From their analysis of HYDROBASE (see <http://www.whoi.edu/science/PO/hydrobase>), Curry and Mauritzen (2005) provide us with a modern answer. They calculate that from the mid 1950s to the mid 1990s, the freshwater storage in the watercolumn of Nordic Seas increased by  $4000 \text{ km}^3$  and the fresh-

water loading of the subpolar gyre, including the Labrador Sea, increased by  $15,000 \text{ km}^3$ . Thus in four decades, the total freshwater stored in the watercolumn across this whole domain increased by some  $19,000 \text{ km}^3$ , equivalent to a supplement to the freshwater supply of  $\sim 15 \text{ mSv}$  over 40 years (though in fact, heavily dominated by the Great Salinity Anomaly of the late 1960s and early 1970s; Dickson et al., 1988). This figure may not be a full accounting of the increased efflux of freshwater from the Arctic to Atlantic over this period. The East Greenland shelf was omitted from the HYDROBASE analysis due to the lack of data; equally, the E Greenland component of the southbound flux may continue south around the Labrador margins, thus bypassing also the HYDROBASE data set for the Northern Gyre; and it remains an open question how much variations in the meridional salt flux by the North Atlantic Current have contributed to the changes observed.

### 3. The freshwater balance in the ‘Arctic Mediterranean’ relative to 35.2

As mentioned above, the freshwater balance for the Arctic Mediterranean can be described as a progressive dilution of Atlantic water from its point of entry to its points of exit. One of the most significant results of the EU VEINS and MOEN programmes has been the consistent estimate of a northward flow of  $8 \text{ Sv}$  of Atlantic water with salinity 35.2 across the Greenland–Scotland Ridge. Recent observations might suggest a larger inflow and also a recent increase in salinity (Østerhus et al., 2005; see Section 1 above), but this will not be taken into account here. The dilution of the Atlantic water is primarily due to the addition of freshwater from river runoff, net precipitation and ice melt, but also by mixing with less saline Pacific water entering through Bering Strait. The transformed Atlantic water then returns to the North Atlantic as low salinity upper waters as well as dense overflow waters. Hansen and Østerhus (2000) estimate that most ( $6 \text{ Sv}$ ) of the Atlantic water returns as less saline ( $34.9$ ), colder and denser overflow water. The salinity of the overflow water varies somewhat, especially in Denmark Strait, but even there the variability (mostly  $<0.1$  about a mean of  $34.9$ ) is sufficiently small to be ignored for the present purpose.

To form a freshwater balance, then, we must consider not only the pure freshwater sources (and sinks) but also the volume balance for the import and export of waters of different salinities. Following Aagaard and Carmack (1989) we consider the Nordic Seas and the Arctic Ocean separately.

A crude representation of the freshwater balance in subarctic seas is provided in Fig. 2, intended as a counterpart (but relative to 35.2) to the Aagaard and Carmack (1989) schematic for the Arctic Ocean, and to Fig. 1.

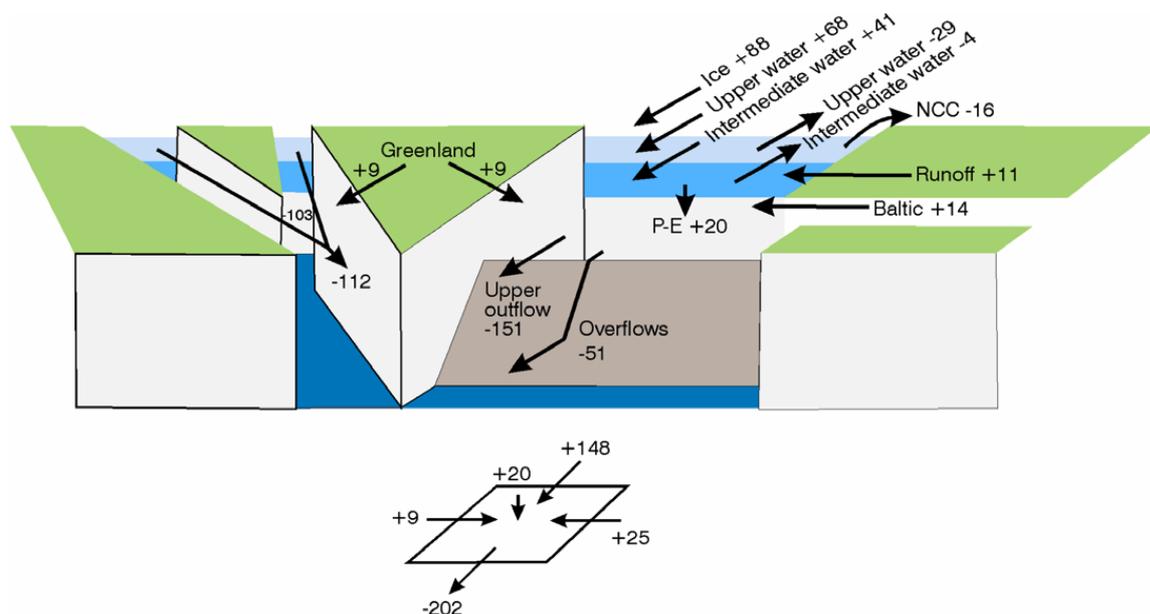


Fig. 2. Schematic breakdown of the freshwater balance in mSv through Subarctic Seas relative to 35.2, the salinity of the most saline inflow component (Atlantic water). The net balances for the Nordic Seas are summarised in the inset diagram.

Table 1  
Estimates of the freshwater balance of the Nordic Seas and Arctic Ocean

Nordic Seas freshwater balance				The Arctic Ocean freshwater balance			
Contribution	Volume (Sv)	<i>S</i>	Freshwater (mSv)	Contribution	Volume (Sv)	<i>S</i>	Freshwater (mSv)
Atlantic inflow	8.0	35.2	0	Runoff	0.1	0	102
Runoff Baltic	0.01	0	14	P – E	0.07	0	65
Runoff Norway	0.01	0	11	Bering Sea inflow	0.8	31.49	84
P – E	0.02	0	20	Barents Sea AW import	1.5	35.05	6
Greenland ice melt	0.01	0	9	Barents Sea NCC import	0.7	34.4	16
Barents Sea Atlantic water export	–1.5	35.05	–6	Fram Strait AW import	4	35.0	23
Barents Sea NCC export	–0.7	34.4	–16	Fram Strait Int. water Import	0.5	34.91	4
Fram Strait Atlantic water export	–4	35.0	–23	Fram Strait Int. water Export	–5	34.91	–41
Fram Strait Intermediate water export	–0.5	34.91	–4	Fram Strait Ice export	–0.09	4	–88
Fram Strait Intermediate water import	5	34.91	41	Fram Strait Upper water Export	–1.15	33.1	–68
Fram Strait ice import	0.1	4	88	CAA upper water (& ice) export	–1.44	32.7	103
Fram Strait Upper water import	1.15	33.1	68				
Overflow export	–6	34.9	–51				
Denmark Strait Upper water (& ice) export	–1.5	31.66	–151				

Note that the volume transports given in the first column of the table are the total transports through the passages, not just the transports of the 35.2 (Atlantic water). The other column gives the freshwater flux relative to 35.2 in mSv.

The component terms in the freshwater balance are listed in Table 1 and are compiled as follows: we separate the freshwater export in the dense overflow water from the freshwater export taking place in the upper, less dense layers. This part of the export from the Arctic Mediterranean occurs not only over the Greenland–Scotland Ridge but also through the straits in the Canadian Arctic Archipelago. Freshwater is added to the Nordic Seas from the Baltic (14 mSv), as runoff from the Norwegian coast (11 mSv) and as melt water from the Greenland icecap (9 mSv). The net direct precipitation to the Nordic Seas is according to Mosby (30 mSv), 30% higher than the estimate given by Aagaard and Carmack (1989). However, Mosby's value includes the Barents Sea, which is already considered in the Arctic Ocean net precipitation estimate given by Serreze et al. (2006). We therefore adopt the 20 mSv suggested by Aagaard and Carmack (1989). Recent work in the EU VEINS and ASOF programmes indicates an outflow of 1.5 Sv of Atlantic water with salinity 35.05 to the Arctic Ocean over the Barents Sea shelf (Ingvaldsen et al., 2004a,b). The VEINS array did not cover the Norwegian Coastal Current and we use the estimate given by Blindheim (1989), 0.7 Sv with salinity 34.4. This corresponds to freshwater exports through the Barents Sea of 6 mSv and 16 mSv relative to 35.2 in the Atlantic water and NCC, respectively. The export of Atlantic water, salinity 35.0, to the Arctic Ocean via Fram Strait has recently been estimated as 4 Sv from the VEINS/ASOF observations (see e.g. Schauer et al., 2004). The exchanges of the deeper waters likely involve a large recirculation but at least some Arctic Intermediate Water and some Nordic Seas deep water pass north of the Yermak Plateau into the Sofia Deep and thus into the Nansen Basin (Rudels et al., 2000; Rudels et al., 2005). We estimate this inflow to be 0.5 Sv with a salinity of 34.91. This gives a freshwater export through Fram Strait to the Arctic Ocean of 23 mSv in the Atlantic layer and 4 mSv in the deep water.

In the Arctic Ocean, freshwater is added by runoff (102 mSv) and by net precipitation (65 mSv; Serreze et al., 2006). Taking the Aagaard and Carmack (1989) estimate for the ice export, used in our earlier discussion, slightly more than half of this freshwater accession (88 mSv) is exported through Fram Strait as ice. Freshwater is also supplied by the low salinity inflow from the Pacific. Woodgate and Aagaard (2005) accept the earlier Aagaard and Carmack estimate of the volume flux through Bering Strait (0.8 Sv) but derive a higher freshwater transport relative to 34.8 of 76 mSv, giving a mean salinity of the Bering Strait inflow of 31.49. Translated into the 35.2 reference salinity, this becomes 84 mSv. Most of the water that exits the Arctic

Ocean leaves as Arctic–Atlantic, intermediate and deep water masses through Fram Strait, the only deep passage. This includes a substantial part of the Barents Sea inflow and the intermediate and deep waters produced on the Arctic Ocean shelves. Since the mixing processes in the Arctic Ocean only dilute a smaller fraction of the Atlantic water entering through Fram Strait most of this water also returns in the deeper layers. A transport of 5 Sv thus appears to be reasonable for the deeper southward flow through Fram Strait. The upper part of the Arctic Atlantic water is less saline than the inflowing Arctic Intermediate Water and the Nordic Seas Deep Water but the denser intermediate and deep waters of the Arctic Ocean are more saline than their Nordic Seas counterparts, so we are justified in adopting a mean salinity of 34.91, the same as the inflowing AIW. This implies that the freshwater export from the deeper layers of the Arctic Ocean becomes 41 mSv. The rest, approximately 1.7 Sv, of the Nordic Seas inflow is transformed into less saline water, which leaves the Arctic Ocean together with the 0.8 Sv of Pacific water in the upper layers. Taking the Pacific inflow, the deep water outflow and the ice export into account the mean salinity of the  $\sim 2.6$  Sv of upper water leaving the Arctic Ocean is 32.9, and if we separate the upper layer water into freshwater and Atlantic (35.2) water the outflow of freshwater becomes 171 mSv and the outflow at 35.2 is 2.42 Sv.

To assess how the freshwater export is distributed between Fram Strait and the Canadian Arctic Archipelago, we note that if the mean salinity of the outflowing upper layer is different in the two passages the transports of freshwater and volume through both passages can be determined using salt and volume balances, providing the mean salinities of the outflows are known. To determine the mean salinities might, however, be just as difficult as determining the transports directly. Nevertheless, to try this approach we assume that the salinities of the outflowing waters are symmetrically distributed around the mean and that the water passing through the Canadian Arctic Archipelago, being closer to the Bering Strait freshwater source, is less saline than that passing through Fram Strait. Tentatively we assume the mean salinity in the Archipelago to be 32.7 and the mean salinity in Fram Strait to be 33.1. If  $X$  Sv of Atlantic water and  $Y$  Sv of freshwater pass through the Archipelago then  $2.42 - X$  Sv of Atlantic water and  $0.171 - Y$  Sv of freshwater leave through Fram Strait. We have;

$$(x + y) \times 32.7 = x \times 35.2$$

$$\{(2.42 - x) + (0.171 - y)\} \times 33.1 = (2.42 - x) \times 35.2$$

which gives  $X = 1.34$  Sv and  $Y = 0.103$  Sv. The outflow of freshwater through the Canadian Arctic Archipelago then becomes 103 mSv and the total volume flux becomes  $\sim 1.44$  Sv. The corresponding volume outflow through Fram Strait is  $\sim 1.15$  Sv, 1.08 Sv of Atlantic water and 68 mSv of freshwater. These numbers are very sensitive to the freshwater input. If the Serreze et al. value for runoff of 95 mSv is used instead of our figure of 102 mSv (Fig. 1), the transports through the Archipelago become 0.87 Sv and 62 mSv, while the transports through Fram Strait increase to 1.72 Sv and 102 mSv, respectively. On the other hand, keeping the symmetry but decreasing (increasing) the salinity in the Archipelago (Fram Strait) would only slightly decrease (increase) the volume transport but increase (decrease) the freshwater flux through the Archipelago. [Note that in Table 1, the volume transports given in the first column of the table are the total transports through the passages, not just the transports of the 35.2 (Atlantic) water. The other columns give the mean salinity and the freshwater flux relative to 35.2 in mSv].

This approximate volume balance between the Nordic Seas and the Arctic Ocean indicates that 6.7 Sv enter the Arctic Ocean and 6.15 Sv return, 5 Sv as intermediate and deep water and 1.15 Sv as upper layer water. Summing all the freshwater contributions to the Nordic Seas gives 202 mSv. In calculating the freshwater balance, there is really no need to delineate the different pathways of the freshwater within the Nordic Seas, e.g. the diffidence of flow from the East Greenland Current into the Jan Mayen and East Iceland Currents. The important point is how much is brought by the overflows directly into the deep North Atlantic, freshening the North Atlantic Deep Water, and how much is exported in the low salinity surface water, which can influence the downstream conditions and the convection in the Labrador Sea and in the Irminger Sea. From the estimate of 6 Sv by Hansen and Østerhus (2000) we find that 51 mSv leaves the Nordic Seas in the overflow waters, and assuming that the ice exported through Fram Strait melts in the Nordic Seas the remaining 151 mSv are mixed into the  $\sim 1.35$  Sv of 35.2 water exiting in the less saline upper layer. The mean salinity of the upper water leaving the Nordic Seas then becomes 31.7. This is a lower limit since ice is exported through Denmark Strait, at least during winter. Using the ice export derived from the NAOSIM model gives

a salinity of 32.4. The freshwater transport in the upper layers through Denmark Strait is close to that found by the NAOSIM model and is 50% higher than the freshwater export through the Archipelago. Much of the low salinity waters of the East Greenland Current will interact with, and dilute, the Irminger Current at the east Greenland shelf and slope. The freshwater export taking place in the overflow water is 1/3rd of that in the upper layers and contributes one quarter of the total freshwater crossing the Greenland–Scotland Ridge.

The East Greenland Current and a part of the Irminger Current become the West Greenland Current west of Cape Farewell. The West Greenland Current passes north through Davis Strait to supply the saline, mid-depth “Atlantic layer” in Baffin Bay. [The relatively low salinity of that Atlantic layer (34.5–34.6) is a clear indication of mixing en route with the freshwater outflow from Fram Strait.] Because of the low salinity in Baffin Bay it is somewhat awkward to give the freshwater export through Davis Strait relative to 35.2. The corresponding Canadian Arctic Archipelago freshwater export relative to 34.8 is 87 mSv, and to obtain the total export through Davis Strait to the Labrador Sea, 9 mSv of melt water from the Greenland icecap have to be added. The resulting estimate of the freshwater outflow west of Greenland of 96 mSv (excluding Hudson Bay) is in the lower range of the numbers mapped in Fig. 1 and less than what is commonly found in the literature (e.g. Loder et al., 1998; Tang et al., 2004; Prinsenbergh and Hamilton, 2004 and Cuny et al., 2006).

Of the two types of flux estimate for the CAA, those calculated to a salinity reference of 34.8 (Fig. 1) are based on fluxes *actually observed* in the CAA, whereas those referred to 35.2 (Fig. 2) reflect the freshwater balances across the entire Arctic Ocean. Only the pure freshwater sources (runoff and  $P - E$ ) are the same for both. The difference between the two approaches could reflect an error of a few tenths in the assumed mean salinity of the outflow or an additional freshwater source of the order of 15 mSv or some combination of the two, but in any event, this difference is rather small, well within the present range of uncertainties.

A ‘balance’ approach is also adopted by Bacon (1997) for his synoptic sections between Cape Farewell and Ireland. Using his observed section-average Atlantic water salinity (34.946), and taking literature values for the transports of salt through the Bering Strait and Canadian Archipelago, and of ice through the Denmark Strait, Bacon estimates the balancing accession of freshwater to the Arctic to be  $170 \pm 60$  mSv. This value corresponds well to the total of the freshwater accession terms listed in Table 1 (the total of Circumarctic and Norwegian Runoff, Baltic outflow and Arctic  $P - E = 205$  mSv).

#### 4. Modelling in support of observations

In the previous sections we have made reference to cases where a direct comparison between models and observations of flux has proved instructive. Nonetheless, as the large-scale system of freshwater flows through subarctic seas evolves into a coherent pattern, large knowledge gaps certainly remain. In part these are due to the difficulties of observing ocean fluxes in the harsh northern environment, and in part due to the financial limitations of covering all gateways with adequate spatial and temporal resolution. Numerical models have an actual or potential role in alleviating both of these problems, and their use in this role might range from process studies to large-scale models covering the Arctic/subarctic domain or entire globe. The potential benefit is that the models would present a consistent dynamical system, but difficulties remain too, arising either from technical issues or from the ambiguities of using observational data for initializing or forcing.

We have little knowledge of the long-term temporal development of the Arctic Ocean’s freshwater content due to a lack of observational data (see e.g. Swift et al., 2005). While numerical models can be used to investigate this topic, the lack of data also poses a specific problem for models since numerical experiments will depend on the initial hydrographic conditions. Therefore numerical experiments hindcasting the last five decades or so should be interpreted with caution particularly when the parameters in question have a renewal time of decades. The numerical hindcast experiments mentioned here all start in the late 1940s. Due to the typical renewal timescale for the Polar Mixed Layer on the order of two decades, the period from the late 1940s to the late 1960s should be left out of consideration in interpreting trends. In hindcast experiments performed with different models of the NAOSIM hierarchy there is a long-term decrease of liquid freshwater amounting to about  $10,000 \text{ km}^3$  between the late 1960s and late 1990s with a parallel gain of about  $3000 \text{ km}^3$  for the Nordic Seas; the remainder drains to the North Atlantic (Karcher et al., 2005; Koeberle and Gerdes, 2007). These numbers are close to the results of the HYDROBASE analysis by Curry and Mauritzen (2005), mentioned above, which estimated an increase of  $4000 \text{ km}^3$  in the freshwater content of the Nordic Seas and

15,000 km<sup>3</sup> in the subpolar North Atlantic over the same period. According to the analysis of Koeberle and Gerdes (2007) the long-term decline of Arctic freshwater content is a result of an imbalance between surface and lateral fluxes since increased fresh water export overcompensates an increased melting of sea-ice that is mostly thermodynamically induced.

Superimposed on these trends, the interannual variability induced by changes in atmospheric forcing can be considerable also. Using different models or model set-ups, Karcher et al. (2005), Koeberle and Gerdes (2007) and Hakkinen and Proshutinsky (2004) all agree in finding a decrease of ~5000 km<sup>3</sup> in the freshwater content of the Arctic between the 1980s and the late 1990s followed by an increase of similar magnitude until the early 2000s. According to the analysis by Karcher et al. (2005), the peak of the liquid release in the mid to late 1990s was associated with a shift of the Arctic upper ocean flow through more-cyclonic atmospheric conditions in the period 1989–1995 (Polyakov and Johnson, 2000), when the NAO/AO was at an extreme high-index state. The resulting changes in the spreading pathways of river runoff, as demonstrated by the NPS model (Maslowski et al., 2000) and an increased inflow of Atlantic water led to an observed and simulated restructuring of the eastern Eurasian and Makarov basins ('retreat of the cold halocline': Steele and Boyd, 1998; Karcher et al., 2005). This restructuring is evident also in the observed (Schlosser et al., 2002) and simulated  $\delta^{18}\text{O}$  distributions in the Arctic (Karcher et al., 2006). Similar changes (shift of river water, freshening of the Beaufort Sea, increased ice and liquid outflow) have been found in a sensitivity study by Zhang et al. (2003) for 10-year periods of high NAO vs. low NAO forcing. Following the large release of the 1990s, the accumulation of about 4000 km<sup>3</sup> of freshwater in the Arctic by the early 2000s (Hakkinen and Proshutinsky (2004); Karcher et al., 2005; Koeberle and Gerdes, 2007) is consistent with the reduction in the AO index since the mid 1990s and a switch to strong anticyclonic conditions in the Arctic (Proshutinsky et al., 2002).

These results are first steps towards a better understanding of the storage and release dynamics of freshwater in the Arctic Ocean, combining models and observations. Open questions remain, for example on the relative importance of different factors such as local wind forcing, lateral fluxes and surface fluxes (P – E, ice processes) in reality as well as in model experiments. The Arctic freshwater balance may also be part of regional feedback loops. Proshutinsky et al. (2002) and Dukhovskoy et al. (2006) discuss the hypothesis of freshwater storage (release) in the Beaufort Gyre in periods of dominant anticyclonic (cyclonic) atmospheric motion (Proshutinsky and Johnson, 1997) in the context of an auto-oscillatory system which involves atmospheric heat and oceanic freshwater fluxes between the Arctic and the Nordic Seas.

The issue of which pathways out of the Arctic are preferred for the exchange of freshwater with the southern basins, both now and in future, remains an open question. Different models show contradictory results (e.g. Maslowski et al., 2000; Karcher et al., 2005; Jungclaus et al., 2005; Koeberle and Gerdes, 2007). Observations are at present insufficient to solve this discrepancy nor do we understand the reason for this difference between model experiments. Details of forcing and representation of topography will certainly play a role.

New and promising results come from other areas of model-data interaction. For example, the enhanced observational activity in the Fram Strait during the recent ASOF-N project has resulted in an improved coverage of freshwater outflow over the East Greenland Shelf and slope. The first available multiyear time-series of freshwater flux estimates from mid 1997–2004 suggested a flux of 32 mSv (rel. to 34.9) on a section from 1.4 to 7.7 W, but omitted the western part of the shelf and parts of the western Fram Strait east of the array. Sub-sampled for the area covered by the moorings, the NAOSIM model (Karcher et al., 2005) gave practically the same mean value and a very similar seasonal amplitude (Hansen et al., 2006). When calculated across the entire strait, the model experiment approximately doubles the freshwater export, and suggests that about 2/3rds of the additional contribution not covered by the moored array stems from the inner shelf and 1/3rd from the western Fram Strait.

A different approach of combining models and data in a beneficial way is the use of inverse methods or data assimilation techniques to provide improved data fields. The advantage of these approaches is that the improved data fields do comply with a dynamically consistent model framework and observations within errors previously assigned to each data point. Recent attempts in the Arctic and subarctic domain have been performed for the volume and heat fluxes in Fram Strait (Losch et al., 2005) and have involved the variational data assimilation of hydrographic data into a general circulation model of the Nordic Seas (Stammer and Koehl, pers. comm.). In a large-scale approach covering the entire Arctic and subarctic domain a variational

data assimilation effort is presently established, based on the NAOSIM code in the framework of the DAM-OCLES project ([www.damocles-eu.org](http://www.damocles-eu.org)).

## 5. The future

Five main factors seem to dominate predicted changes in the freshwater budget at high latitudes.

- Arctic Ocean Sea-ice is generally expected to decrease in volume over the 21st century. This is the conclusion drawn from various studies with coupled climate models, including different scenarios for future greenhouse gas emissions (e.g. Lindsay and Zhang, 2005). An integration of the annual coupled simulations performed at the School of Earth and Ocean Sciences University of Victoria, B.C. Canada. [[http://wikyonos.seos.uvic.ca/movies/ice\\_annual.html](http://wikyonos.seos.uvic.ca/movies/ice_annual.html)] for example, is based on a 1% CO<sub>2</sub> increase per year after 1998. The analysis gives a long-term steady decrease from 18,500 to 10,000 km<sup>3</sup> (46%) in total sea-ice volume over the period 2000–2100 (Weaver, Eby and Wiebe, pers. comm., 2004). Such a loss of 8500 km<sup>3</sup> in 100 years is equivalent to an additional freshwater input to the ocean of 2.7 mSv over this period (Weaver pers. comm. 2004), similar to the annual losses of ice-volume (~2.8 mSv) estimated by Rothrock and Zhang (2005) for the past half-century (1948–1999). It has to be stated, however, that sea-ice parameters like thickness, extent and representation of the seasonal cycle from presently published coupled climate models differ greatly between themselves and in comparison with observations (Flato et al., 2004; Hu et al., 2004), as do atmospheric parameters such as air temperature (Holland and Bitz, 2003) and sea-level pressure. In their 52-year simulation, Rothrock and Zhang (2005) find that the wind-forced component of sea-ice volume has no substantial trend but the temperature-forced component has a significant downward trend of –3% per decade. An additional uncertainty is the change to be expected in the separation of water into sea-ice and brine in the Arctic Ocean at times of low ice volume. Depending on the distribution of the sea-ice, a low volume does not necessarily mean small ice production in winter (e.g. Köberle and Gerdes, 2003). In the latter study the air temperature distribution and the wind field both play essential roles. Köberle and Gerdes suggest that the advective time scales and distribution patterns of freshwater may be different for the liquid form as compared to sea-ice, implying that details of the ice formation and melt will have an impact on the future fluxes of freshwater from the Arctic to the subarctic seas.
- The evidence of trend in river discharge to the Arctic Ocean is equivocal. On the one hand, the discharge of the major Eurasian rivers to the Arctic Ocean is expected to increase with Arctic warming. Peterson et al. (2002) have derived an empirical relationship between recent warming and past Russian riverflow amounting to +7 mSv per °C. Citing present predictions of a global rise in surface air temperature of between 1.4 and 5.8 °C by 2100 (IPCC, 2001), they therefore predict that the “*discharge from the six largest Eurasian rivers alone would increase by 10–40 mSv by 2100*”. From their HadCM3 simulations, Wu et al. (2005) also project an increase of between 10 and 20 mSv for river discharge into the Arctic by 2050, according to the IPCC SRES B2 and A2 forcing scenarios. Dery and Wood (2005) however show a clear declining trend in Canadian river discharge to the Arctic Ocean amounting to 10% since the mid-1960s. The routing of atmospheric moisture flux to the Arctic is known to change markedly between the negative and positive phases of the NAO, tending to pass north through the Canadian Arctic during NAO-negative conditions but overwhelmingly through the Nordic Seas during NAO-positive extrema (Dickson et al., 2000; see their Fig 5). Thus this apparent geographic difference in the trends of river discharge to the Arctic may simply reflect the amplification of the NAO/AO since the 1960s; both Peterson et al. and Dery and Wood report statistically-significant links between the discharges they cover and the NAO/AO. However, a geographic difference in the trends of discharge may be harder to explain in terms of anthropogenic forcing, despite the clarity of this result in the HadCM3 experiments (Wu et al., 2005).
- The freshwater flux from Greenland is expected to increase. As one recent example based on a mix of modelling techniques, simulations by Fichet et al. (2003) indicate that the annual mean total freshwater flux from Greenland will increase by 15 mSv over the period 1970–2080. Jeff Ridley (Hadley Centre pers. comm. 2004) suggests a slightly smaller increase based on the A-2 scenario (+10 mSv).

- The freshwater exchanges between the Arctic Ocean and subarctic seas can all be expected to change dramatically. Though the precise nature of the changes are likely to be model- and scenario-dependent, we illustrate their general sense from the recent experiments of the M-P-I group who have coupled the M-P-I Climate Model ECHAM-5 (T63 resolution; 31 levels) with the M-P-I Ocean Model (1.5°, 40 levels) including an embedded dynamic/thermodynamic sea-ice model, to provide a modern set of projections for each gateway (Haak et al., 2005; Koenigk et al., 2005). Briefly, by 2100: (1) the Bering Strait freshwater inflow will increase by 37% (a large increase in the inflow of liquid freshwater combined with a small decrease in the efflux of ice), (2) the combined Fram Strait–Barents Sea outflow will hardly change as a large increase in liquid freshwater outflow is matched by the dwindling to near-vanishing point of the sea-ice export. (3) The combined freshwater outflow through the CAA is expected to increase by almost 50% as the dwindling of its smaller sea-ice component is insufficient to counter the expected large rise in liquid freshwater export. By the end of their experiment, the M-P-I Group suggest that CAA freshwater export may reach Fram Strait values.
- It seems inevitable that global warming will be accompanied by an acceleration of the global water cycle. A circumpolar freshening of the intermediate waters of the world ocean may already be underway (Wong et al., 2001; Curry et al., 2003).

## 6. Summary and conclusions

A large observational effort over recent decades in some of the most intractable conditions on Earth has permitted the construction of a first map of the freshwater fluxes connecting the Arctic with the North Atlantic through subarctic seas. Our present rough ‘snapshot’ of fluxes (Fig. 1) suggests that the annual accession of freshwater to the Arctic basin (see Serreze et al. for the most modern accounting) is balanced by an outflow of approx 100–125 mSv to the Atlantic either side of Greenland, with a further contribution of 46 mSv or so from the runoff and ice melt of Hudson Bay. None of these numbers are yet established with any confidence, though the key measurement sites have been identified and modern state-of-the-art arrays are now in place at each of them.

Though our direct measurements are too brief as yet to comment, the analysis of the historic hydrographic record and modelling appear to agree that the efflux of freshwater from the Arctic has increased steadily over recent decades (Curry and Mauritzen, 2005), and model-based projections suggest that this will continue, reflecting increased circumarctic river inputs (Wu et al., 2005), a progressive thinning and retraction of the Arctic perennial sea-ice (eg Comiso, 2002), an increase in the freshwater production of Greenland (Fichefet et al., 2003) and increases in the combined ice and freshwater outflows through the CAA and Fram Strait (eg Haak et al., 2005).

While coupled climate models predict major increases in the liquid freshwater flux through both of these gateways (+75% through the CAA and +158% through Fram Strait by 2070–2099 according to the experiments by Haak et al., 2005), the dwindling of the ice-flux is expected to have a disproportionately large effect in Fram Strait, where the sea-ice component currently dominates. As a result, the major future increase in the *total* freshwater flux is expected to take place through the CAA in these simulations (+50%) while the total flux through Fram Strait may hardly vary (+3%). Model results by Steele et al. (1996) had earlier prompted a similar conclusion; that in the case of an increased melt of the Arctic sea-ice, the freshwater excess would be exported mainly through the CAA rather than Fram Strait.

While there is agreement that an increasing freshwater flux through Fram Strait to the North Atlantic is likely to be of climatic significance, we remain uncertain as to whether the impact on climate will result from *local* effects on overflow transport (e.g. from the changing density contrast across the Denmark Strait sill; Curry and Mauritzen, 2005), from the *regional* effect of capping the water column of the NW Atlantic (leading to a reduction in vertical mixing, water mass transformation, and production of North Atlantic Deep Water), or from *global-scale* changes in the Ocean’s thermohaline fields and circulation arising from an acceleration of the Global Water Cycle (Curry et al., 2003). Results are much more equivocal regarding the impact of an increased flux of freshwater through the CAA. Myers (2005) using a high resolution regional model suggests that very little (6–8%) of the freshwater exported from the Canadian Arctic gets taken up in the Labrador Sea

water of his model; yet coarser resolving global models find a 10% decrease of the MOC between closing and opening the CAA (Goosse et al., 1997), or a decrease in the MOC which could reach even 35% (Wadley and Bigg, 2002).

The more specific point regarding the effects of freshwater fluxes on the MOC stems from the recent reanalysis of HadCM3 output by Vellinga (2005) which demonstrated that MOC slowdown was more effectively achieved if a given freshwater increase was delivered to the surface of the NW Atlantic than if spread to depth (for example by overflows; Dickson et al., 2002). The prediction that the major future increase in freshwater flux from the Arctic will happen west of Greenland (above) is of some climatic significance therefore since it is much less likely to be incorporated in, and spread to depth by, the dense water overflows from Nordic Seas.

We conclude that the resolution of these key climate issues will be most effectively addressed by extending our series of direct freshwater flux measurements to the point where we understand *directly* how they vary, why their pathways change, and how they are forced, in collaboration with a parallel effort in model development.

When constructing a freshwater *balance* for the Nordic Seas it becomes convenient to make the calculation with respect to 35.2, the salinity of the most saline water (Atlantic water) entering the system. From the difference between a 197 mSv southbound flux of ice and freshwater through Fram Strait and a 49 mSv northbound flux of freshwater to the Arctic Ocean via Fram Strait and the Barents Sea, a net import of 148 mSv of freshwater enters the Nordic Seas across its northern Boundary. There it is supplemented by around 54 mSv of freshwater from Baltic runoff, Norwegian runoff, P – E and Greenland ice melt, so that the total freshwater contribution to the Nordic Seas from all sources is 202 mSv. Of this, around 51 mSv of freshwater is estimated to pass south to the deep Atlantic in the dense water overflows (Hansen and Østerhus, 2000), and we assume that the balance of 151 mSv leaves the Nordic Seas in the upper water export through Denmark Strait. The corresponding estimate that we derive for the freshwater outflow west of Greenland (103 mSv relative to 35.2 or 96 mSv relative to 34.8) is in the lower range of estimates in the literature (eg Loder et al., 1998; Tang et al., 2004; Prinsenberg and Hamilton, 2004 and Cuny et al., 2006).

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