Chapter 13
Constraints on Estimating Mass, Heat and Freshwater Transports in the Arctic Ocean: An Exercise

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13.1 Introduction

The ASOF programme, with its study of the transports between the Arctic Ocean and the North Atlantic via the subarctic seas – the Nordic Seas, Baffin Bay and the Labrador Sea –, also provides an opportunity to examine the mass (volume), freshwater and heat budgets of the Arctic Ocean. The exchanges between the two passages between the Arctic Ocean and the Nordic Seas, Fram Strait and the Barents Sea opening between Norway and Bear Island, have been measured continuously since 1997, first in the VEINS programme (Variability of Exchanges in the Northern Seas) and then in ASOF and the observations are presently continued within the DAMOCLES (Developing Arctic Modelling and Observing Capabilities for Long-term Environmental Studies) programme. The transports through two of the three main channels in the Canadian Arctic Archipelago, the Lancaster Sound and the Jones Sound, have been directly measured for a couple of years now (Prinsenberg and Hamilton 2005), and the instruments from the first year-long measurements in Nares Strait have been brought in. The fluxes through Bering Strait have also been studied intensely the last 10–15 years (e.g. Woodgate and Aagaard 2005). The work within ASOF has shown that the transports through Fram Strait and through the Canadian Arctic Archipelago are those most difficult to determine. The Archipelago because of the severe climate, the remoteness of the area and the nearby location of the magnetic North Pole, Fram Strait because of its depth, the transports in both directions, and the presence of baroclinic and barotropic eddies leading to high spatial and temporal variability.

The estimates of the mean transport through Bering Strait obtained since the mid-1980s have ranged around 0.8 Sv (1×10^6 m^3 s^{-1}), but large seasonal variations have been reported, 1.2 Sv in summer and 0.4 Sv in winter (Coachman and Aagaard 1988; Woodgate and Aagaard 2005). The mean transport of Atlantic water to the Arctic Ocean through the Barents Sea opening has been estimated to 1.5 Sv from

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the observations in VEINS and ASOF, but with large short periodic variations (Ingvaldsen et al. 2004a, b). A longer time variation with a period of 3–4 years also appears to be present, causing the transport to change from below 1 Sv to slightly above 2 Sv (ASOF-N Final Report 2006). In addition to the inflow of Atlantic water there is also the contribution from the Norwegian Coastal Current, which amounts to 0.7 Sv with salinity 34.4 (Aagaard and Carmark 1989 based on Blindheim 1989). The Arctic Ocean also receives a freshwater input from runoff and net precipitation amounting to 0.15–0.2 Sv (Serreze et al. 2006). Assuming these estimates to be close to reality, the total transport through Bering Strait and the Barents Sea and the freshwater input, adding up to 3.2 Sv, can be used, together with requirements of mass and freshwater balance, to evaluate the transport estimates derived from the observations in Fram Strait. The passages and the transports are indicated in Fig. 13.1.

We begin by examining some of the estimates obtained in Fram Strait during different phases of the VEINS and ASOF programs and what these transports imply for the Arctic Ocean mass and freshwater budgets. In fact, this exercise was provoked

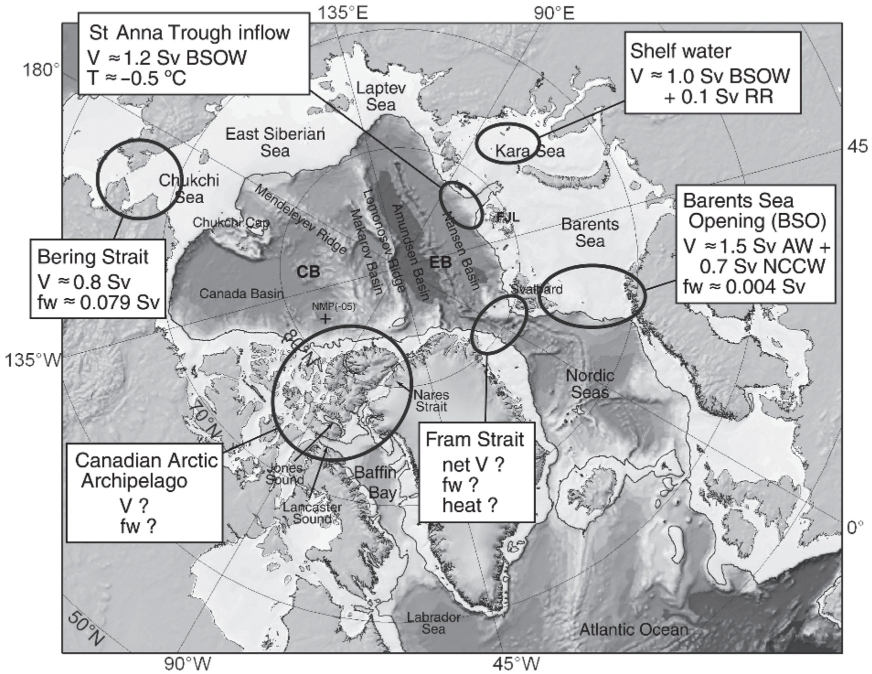

Fig. 13.1 The four main passages between the Arctic Ocean and the world ocean. The Bering Strait inflows are adopted from Woodgate and Aagaard (2005) and the inflows through the Barents Sea Opening (BSO) are taken from Ingvaldsen et al. (2004a) Atlantic Water (AW) and Blindheim (1989) Norwegian Coastal Current Water (NCCW). For the separation of the BSO inflow into a deep inflow via St Anna Trough and a less saline shelf water see discussion in Section 13.6. The freshwater is computed relative to 34.92. The river runoff is taken from Dickson et al. (2007). CB (Canadian Basin), EB (Eurasian Basin), FJL (Franz Josef Land), MNP (Magnetic North Pole). The Lambert equal area projection has been provided by M. Jakobsson

by the report that observations from the current meter array showed a net northward transport persisting for more than 1 year (ASOF-N 2nd Annual Report 2005). Is such result compatible with the transports found through the other passages?

Concentrating on the net transport through Fram Strait, presently ignoring the total northward and southward fluxes, the long-term mean net transport is southward and estimated from the mooring array to be 0.6 Sv (ASOF-N 2nd Annual report 2005). (This value was later adjusted to 1.7 Sv (ASOF-N Final report 2006)). Using 0.6 Sv mass conservation demands a mean outflow of 3.2 – 0.6 = 2.6 Sv through the Canadian Arctic Archipelago. The commonly cited estimates for the outflow through the Canadian Arctic Archipelago range between 1 and 2 Sv converging toward 1.7 Sv (e.g. Melling 2000; Prinsenberg and Hamilton 2005). Can the straits in the Archipelago sustain a mean outflow of 2.5 Sv? Suppose that this is not the case. There is then an imbalance and water is accumulating in the Arctic Ocean at a rate of 1 Sv. The area of the Arctic Ocean is, including the shelves, 10×10^{12} m^2 and imbalances of this order would raise (lower) the sea surface by 25 cm in 1 month. One month is then probably the longest period such an imbalance can prevail.

These speculations can be extended further. A net northward flow (inflow) of 0.4 Sv was estimated from the Fram Strait array in 2002–2003 and this situation prevailed for more than 1 year (ASOF-N 2nd Annual Report 2005). This amounts to a total inflow of 3.6 Sv, which, to maintain mass balance and assuming an ice export of ~0.1 Sv, requires an outflow of ~3.5 Sv through the Canadian Arctic Archipelago. Only water from the upper 250 m can pass through the straits in the Archipelago, and even if there is a net inflow through Fram Strait the East Greenland Current will still carry low salinity upper water out of the Arctic Ocean at a rate of ~1 Sv. This implies a total outflow of ~4.5 Sv of Polar surface water, more than twice the available input of low salinity water from Bering Strait, from river runoff, from the Norwegian Coastal Current, and from the interaction between sea ice and the Fram Strait Atlantic inflow (see below for details). The outflow would reduce a 100 m thick upper low salinity layer in the deep basins by 10 m in 1 year. The net inflow through Fram Strait was observed during a period, when the Barents Sea inflow was close to its maximum (ASOF-N Final Report 2006). A net inflow can therefore not be explained by smaller transport through the Barents Sea.

If the upper layer thickness is to be maintained, sea ice must be melted and mixed into the entering Atlantic water to re-supply the exported low salinity water. To produce the 2.5 Sv of additional upper water with salinity 33.2, assuming this to be a realistic mean value of the salinity of the outflows in the East Greenland Current and through the Archipelago, requires an ice melt rate of 0.12 Sv, taking the Atlantic water salinity to be 35. This is of the same order as the present ice export and implies that the ice volume over the deep basins ($3 \times 5 \times 10^{12}$ m^3), using a mean ice thickness of 3 m, would be reduced by 20–25% in 1 year. The ice melt would also require that 40 TW of the heat entering the Arctic Ocean goes to ice melt. This is about equal to the heat released by cooling 2.4 Sv of Atlantic water (3 °C) to the freezing point. Furthermore, melting sea ice by sensible heat stored in the water column may not be possible without also supplying a substantial amount of heat to the atmosphere (Rudels et al. 1999a). The required heat input would then be even larger.

It should be kept in mind that these numbers and scenarios describe possible responses of the Arctic Ocean to large perturbations and do not represent the present situation in the Arctic Ocean, which is one where about 0.1 Sv liquid freshwater is transformed into ice (equal to the ice export). The excessive ice melt is needed, if the stratification in the Arctic Ocean basins shall be maintained during a major inflow event. A more likely effect is a thinning of the upper layer.

The examples described above show that some questions may still be asked and some insight might still be gained by studying basic mass, heat and salt balances. To be specific; we shall examine the contributions from Fram Strait to the mass (volume), heat and freshwater budgets of the Arctic Ocean using geostrophically determined transports through hydrographic sections obtained in Fram Strait between 1980 and 2005.

The reasons for using geostrophy instead of the results from the current meter array are: (1) Hydrographic observations are easier to work with and to interpret. (2) The time series of the hydrographic observations is considerably longer than the period of direct current measurements. (3) The spatial resolution on the hydrographic sections is finer than for the current meter array and allows for a better identification of water masses. On the other hand, the temporal resolution (about once a year) is considerably worse than that of the array. (4) The geostrophic transports are undetermined with respect to the reference velocity. If the transports do not fulfill obvious required budget constraints, it is then possible, and permissible, to deduce where an error might reside and also to suggest plausible corrections of the computed transports. Such corrections are much more difficult to defend with direct current measurements, which, when treated correctly, should give an optimal estimate.

In Section 13.2 we discuss the assumptions made when estimating the geostrophic transports through Fram Strait (Section 13.2.1) and then determine the exchanges of volume (Section 13.2.2). The choice of reference temperature and reference salinity is presented in Section 13.3. The distribution of the transports in different areas of the strait and the exchanges of different water masses are examined in Section 13.4. The mean Θ–S properties of the in- and outflow of the different water masses are computed for each crossing, and their variations with time and in the different part of the strait are discussed in Section 13.5. The heat transport is studied in Section 13.6 and the freshwater transport in Section 13.7. In Section 13.8 the obtained transports through Fram Strait are used, together with the requirement of mass, heat and freshwater balances of the Arctic Ocean, to examine if they lead to realistic outflows of mass (volume) and freshwater through both Fram Strait and the Canadian Arctic Archipelago. The inflows through Bering Strait (Woodgate and Aagaard 2005) and through the Barents Sea opening (Ingvaldsen et al. 2004a, b), as well as the river runoff and the net precipitation (Serreze et al. 2006; Dickson et al. 2007) are then assumed known (see Fig. 13.1). If not both the freshwater balance and the volume balance are acceptable, we will re-examine and adjust the geostrophic transport through Fram Strait to establish more realistic balances. The results of the study are summarised in Section 13.9.

13.2 Transports

13.2.1 The Geostrophic Calculations

The transports through 16 hydrographic sections taken between 1980 and 2005 are determined using the dynamic method. The first section was obtained in 1980 from the Swedish icebreaker Ymer, and the 1983 and 1984 crossings were made by RV Lance on regular Norwegian Polar Institute cruises. The 1988 and 1993 sections were taken by RV Polarstern on AWI expeditions and from 1997 onwards the sections have been obtained within the VEINS and ASOF programs. The sections taken in the 1980s used Neil Brown CTDs and the station spacing was generally larger than on the sections from 1997 onwards. SeaBird CTDs have been used since 1993. The data quality improved significantly between the 1980s and the 1990s. All sections run along the sill at about 79° N except 1983 which was taken along 79° 15 N (over the Molloy Deep). All sections were obtained in late summer, August–September except 1988 (June) and 1993 (March). For further details see Table 13.1.

On the sections the depth at each station is assumed constant halfway to the neighboring stations and the temperatures and salinities (1 or 2 db average) observed at the station are taken to extend halfway to the neighboring stations. Between stations of unequal depth the method of Jacobsen and Jensen (1926) is used to estimate the density anomaly correction below the deepest common level. Direct current measurements have shown that both the West Spitsbergen Current and the East Greenland Current are largely attached to the continental slope and follow the isobaths with shallow water to the right. To mimic this behavior within

Table 13.1 Information on sections and number of stations

Year	Vessel	Institute/programme	Stations 9° E – 6° W	Stations shelf
1980	IB Ymer	Ymer – 80	15	5
1983	RV Lance	NPI	23	–
1984	RV Lance	NPI	17	3
1988	RV Polarstern	AWI	18	–
1993	RV Polarstern	AWI U. Hamburg	17	–
1997	RV Lance	VEINS	16	2
1998	RV Polarstern	VEINS	20	14
1999	RV Polarstern	VEINS	26	8
2000	RV Polarstern	VEINS	16	20
2000	RV Lance	VEINS	22	9
2001	RV Polarstern	AWI	27	12
2002	RV Polarstern	AWI	49	23
2003	RV Polarstern	ASOF	50	–
2004	RV Polarstern	ASOF	42	7
2005	RV Polarstern	ASOF	50	24

the geostrophic framework we set the velocity to zero at the bottom of that station of the pair, which results in a flow at the deeper station, below the deepest common level, that has the shallower station to the right, looking in the direction of the flow.

A variational approach with auxiliary constraints on the deep-water exchanges is finally applied to the deep part of the strait. The Arctic Ocean is known as a source of dense water, warmer and more saline than the deep-water masses formed in the Nordic Seas (the Greenland Sea). We expect the deep-water formation to have a relaxation time scale comparable to the ventilation times of the deep basins, ranging from about 30 years in the Greenland Sea to perhaps 400 years in the Canada Basin. This is long enough to expect a fairly constant, baroclinic exchange of the deep waters during the observation period. The increase in temperature and salinity observed in the deep waters in the strait, however, suggests that the deep transports might be changing during the period. If so, it is ignored.

The circulation in the deeper layers is largely confined to the Arctic Ocean and the Nordic Seas and the Arctic Ocean appears at present to be a more active source of deep water than the Greenland Sea. We postulate that a production of 0.4 Sv of deep water with a mean salinity of 34.9325 takes place in the Arctic Ocean by brine rejection on the shelves and subsequent sinking of dense saline plumes down the slope, entraining warmer intermediate water on their way to their equilibrium density levels (Rudels 1986; Rudels et al. 1994; Jones et al. 1995). The dense water production by open convection in the Greenland Sea is assumed strong enough to generate an inflow of 0.2 Sv of deep water ($\sigma_\theta \geq 28.06$) with salinity 34.910 from the Nordic Seas to the Arctic Ocean. Since we do not expect that any deep water advected into the Arctic Ocean to be mixed upward into the overlying layers, this implies an outflow of 0.6 Sv with salinity 34.925 through Fram Strait from the Arctic Ocean. The volume and salt constraints on the deep exchanges then become $M = -0.4 \times 10^6\ m^3\ s^{-1}$ and $S = -13.973 \times 10^6\ kg^{-1}$. The flow field with the least added kinetic energy below the density surface $\sigma_\theta = 28.06$, fulfilling these constraints, is then determined.

The minimization of the added kinetic energy below the 28.06 isopycnal leads to a weak flow field, and the constraints on the deep water exchange are mainly introduced to ascertain that the more saline Arctic Ocean deep waters, to the west, leave and the Nordic Seas deep waters, mainly located to the east, enter the Arctic Ocean. A stronger outflow could be obtained by increasing the net deep water export, and a more intense deep circulation would be generated by increasing the salt export while keeping the net volume flux. However, the deep exchanges between the Arctic Ocean and the Nordic Seas as well as the deep-water production in the two areas are essentially unknown and the constraints have therefore been kept small. They force the deep outflow to take place in the west and the inflow to the east consistent with the locations of the East Greenland Current and the West Spitsbergen Current, but, because of the small added barotropic velocities, ~0.01 m s^{-1}, they do not unduly influence the transports in the upper layers, the main concern in this work, which are then essentially geostrophic. If reliable estimates of the deep-water productions in the two areas become available a more realistic barotropic flow field can be determined.

13.2.2 Volume Transports

The geostrophic transports are shown in Fig. 13.2, central panel. The total in- and outflows range from 5 Sv to almost 15 Sv with an average inflow of ~6 Sv and an outflow close to 9 Sv. This is smaller than the transports obtained from the direct current measurements, but not alarmingly so. The net outflow, 2.5 Sv, is, however, larger than that reported from the current meter array (e.g. Schauer et al. 2004; ASOF-N 2nd annual report 2005; ASOF-N final report 2006). The total in- and outflows estimated here include everything that is moving north and south and do not discriminate between eddies and more organized exchanges. The slight increase in total transports that is noticed in recent years might then be due to the closer station spacing on the later sections.

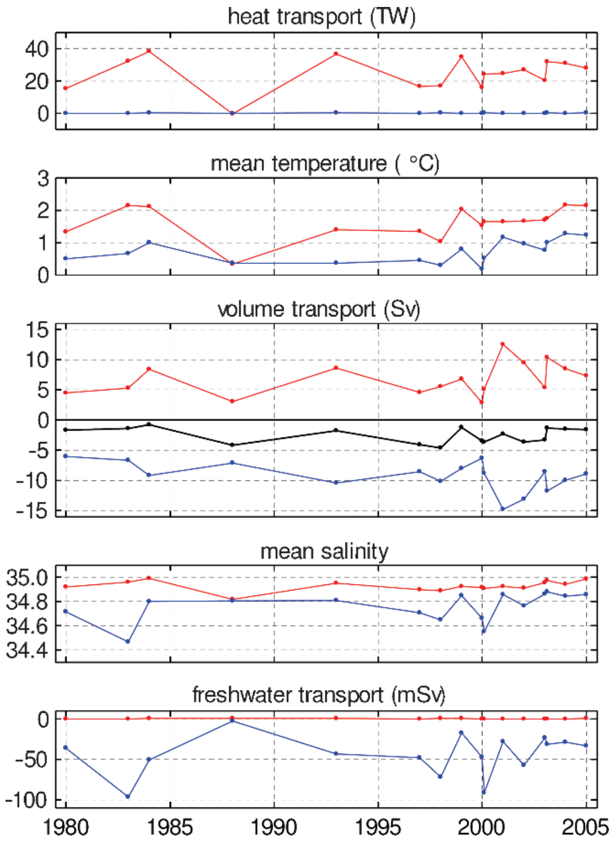

Fig. 13.2 Centre frame: Total in (red), out (blue), and net transports (black) in Sv obtained from the geostrophic computations. Upper frames: Mean inflow temperature (red) and mean outflow (reference) temperature (blue) and heat transport into the Arctic Ocean (red), the heat export (blue) is zero. Lower frames: Mean inflow (reference) salinity (red) and mean outflow salinity (blue) and the liquid freshwater export (blue). The freshwater import (red) is zero

Fram Strait probably contributes more than 60% of the inflow and 80–90% of the outflow volumes, if the deep exchanges are included. Although the Barents Sea inflow supplies intermediate and deep water to the Arctic Ocean, these dense waters are created, by cooling and also by freezing, in the Barents Sea. Similarly a small amount of Pacific water is made dense enough on the Chukchi Sea to enter the Canada Basin deep water. However, Fram Strait is the only passage that allows deep water to enter, and perhaps more important, the only passage that permits an outflow of deep water.

13.3 Reference Temperatures and Reference Salinities

To properly assess the Fram Strait contribution to the heat and freshwater balances of the Arctic Ocean all in- and outflows have to be accounted for, and a mass balance must first be established. Although this is one of the ultimate aims of ASOF, it has, as yet, not been accomplished. Without mass balance the heat and freshwater transports will depend upon the choice of reference temperature and reference salinity. Often these have been set as –0.1 °C and 34.80, taken as representing the mean temperature and the mean salinity of the Arctic Ocean (e.g. Aagaard and Greisman 1975; Aagaard and Carmack 1989; Simonsen and Haugan 1996; Schauer et al. 2004; Serreze et al. 2006). These values were determined in the 1970s, if not earlier, when the observational basis for forming such averages was very slim, and the variability in space and time of the Arctic Ocean water masses that has become evident during the last 10–15 years (e.g. Quadfasel et al. 1991; Polyakov et al. 2005) makes it doubtful that values determined 30 years ago can still be used without qualification.

Acknowledging the fact that we do not have, at present, sufficient observations from the other passages to formulate a mass balance of the Arctic Ocean, and taking into consideration the temporal variations of the Arctic Ocean mean temperature and salinity, we here choose a different approach. In view of the overreaching importance of the exchanges through Fram Strait we deem it sensible to estimate the inflow of heat to the Arctic Ocean and the outflow of freshwater from the Arctic Ocean through each section in Fram Strait relative to the mean outflow temperature and the mean inflow salinity determined on that section. This implies that no heat is transported by the outflowing water and no freshwater is transported by the inflowing water through the sections in Fram Strait. It should be noted that since the outflow is larger than the inflow, these choices give the largest transports of heat and freshwater through Fram Strait, unless reference temperatures, higher than the mean outflow temperature, and reference salinities, higher than the mean inflow salinity, are used. To compare the results obtained here with other estimates using different reference values, the differences in reference values should be multiplied with the net volume transport.

This does not eliminate the necessity to close the mass (volume) budget for the Arctic Ocean to really determine the fate of the heat entering the Arctic Ocean through Fram Strait and to estimate the relative contribution of the momentary export of liquid freshwater through Fram Strait in the Arctic Ocean freshwater budget. To use these varying reference salinities and temperatures might therefore appear a futile exercise. However, they bring the balances down to simple inflow/outflow terms, which makes it possible to discuss the mass imbalance, its origin and what it can reveal about the redistribution of the heat carried by the entering Atlantic water.

Furthermore, by comparing the time series of the heat transport, the reference temperature and the inflow and outflow volumes different factors contributing to the variability of the heat transport can be assessed. In a similar manner the variability of the freshwater export can be related to the variability of the reference salinity and the exchanged volumes (Fig. 13.2). These tasks have not been attempted here. Before we turn our attention to the heat and freshwater fluxes, we shall further discuss the exchange of different water masses through Fram Strait and how the transports are distributed in different parts of the strait.

13.4 Exchanges of Different Water Masses

The obtained estimates do not, so far, say anything about the exchanges of different water masses, nor where in the strait the main transports take place. A detailed water mass definition for the Arctic Mediterranean Sea has been formulated elsewhere (Rudels et al. 2005), but for the transports here we introduce a simplified water mass classification of 6 water masses, Surface water (SW), Atlantic water (AW), dense Atlantic water (dAW), Intermediate water (IW), Deep water I (DWI) and Deep water II (DWII) separated mainly by isopycnals but in the case of dAW and IW by the 0 °C isotherm (Table 13.2 and the Θ–S diagrams in Fig. 13.5).

The net outflow occurs as surface water and in the dense Atlantic water and the intermediate water ranges. It appears reasonable that waters from other passages that leave the Arctic Ocean through Fram Strait create net outflows with properties that at least partially reflect their initial characteristics. The low salinity of the less dense surface outflow (see Fig. 13.5a) reveals that it originates from the part of the Barents Sea inflow, mainly comprising Norwegian Coastal Current water, that stays on the shelves and incorporates most of the Siberian river runoff. Some ice melt might also be present as well as low salinity Pacific water from Bering Strait,

Table 13.2 Simplified water mass classification

Surface water (SW)	$\sigma_\theta < 27.70$
Atlantic water (AW)	$27.70 \le \sigma_\theta < 27.97$
Dense Atlantic water (dAW)	$27.97 \le \sigma_\theta, \sigma_{0.5} < 30.444, 0 < \theta$
Intermediate water (IW)	$27.97 \le \sigma_\theta, \sigma_{0.5} < 30.444, \theta < 0$
Deep water I (DWI)	$30.444 \le \sigma_{0.5}, \sigma_{1.5} < 35.142$
Deep water II (DWII)	$35.142 \le \sigma_{1.5}$

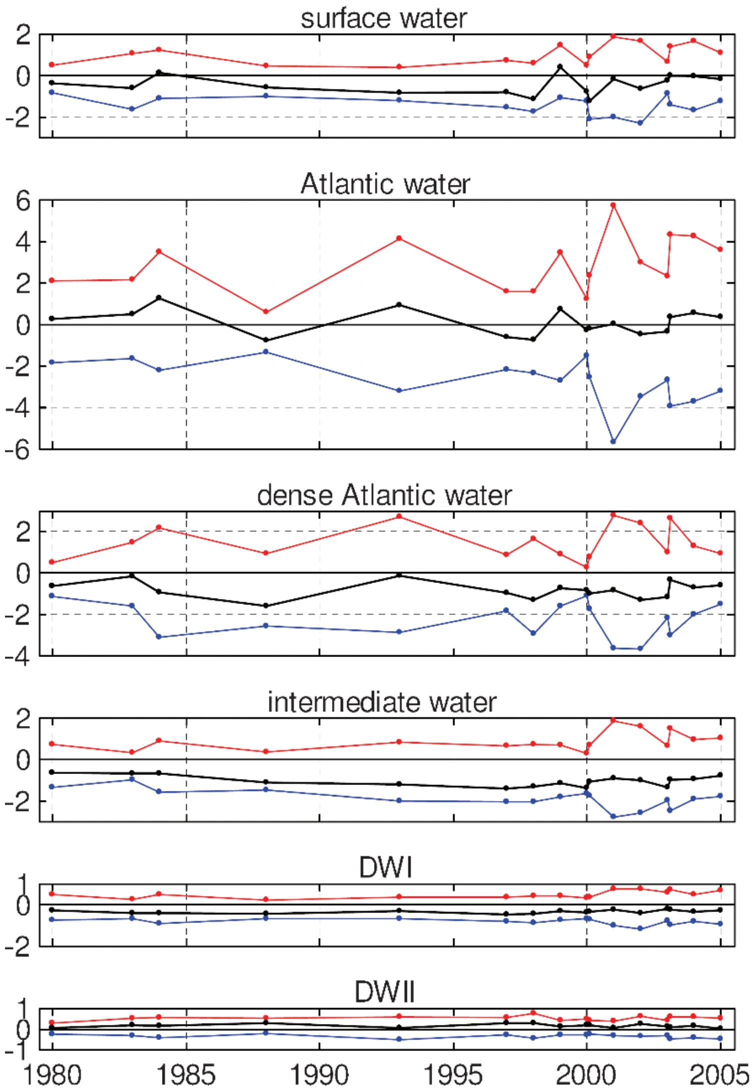

Fig. 13.3 Transports in Sv of different water masses based on geostrophic calculations. Inflow (red), outflow (blue) and net transport (black)

although the Pacific water mainly leaves the Arctic Ocean through the Canadian Arctic Archipelago (Jones et al. 2003). The net outflow in the denser, intermediate water range largely derives from the part of the Barents Sea inflow that enters the deeper Arctic Ocean water column via the St Anna Trough (Fig. 13.3).

The Fram Strait sections are subdivided into five different areas. Four of them, the eastern slope, the eastern deep part, the western deep part and the western slope

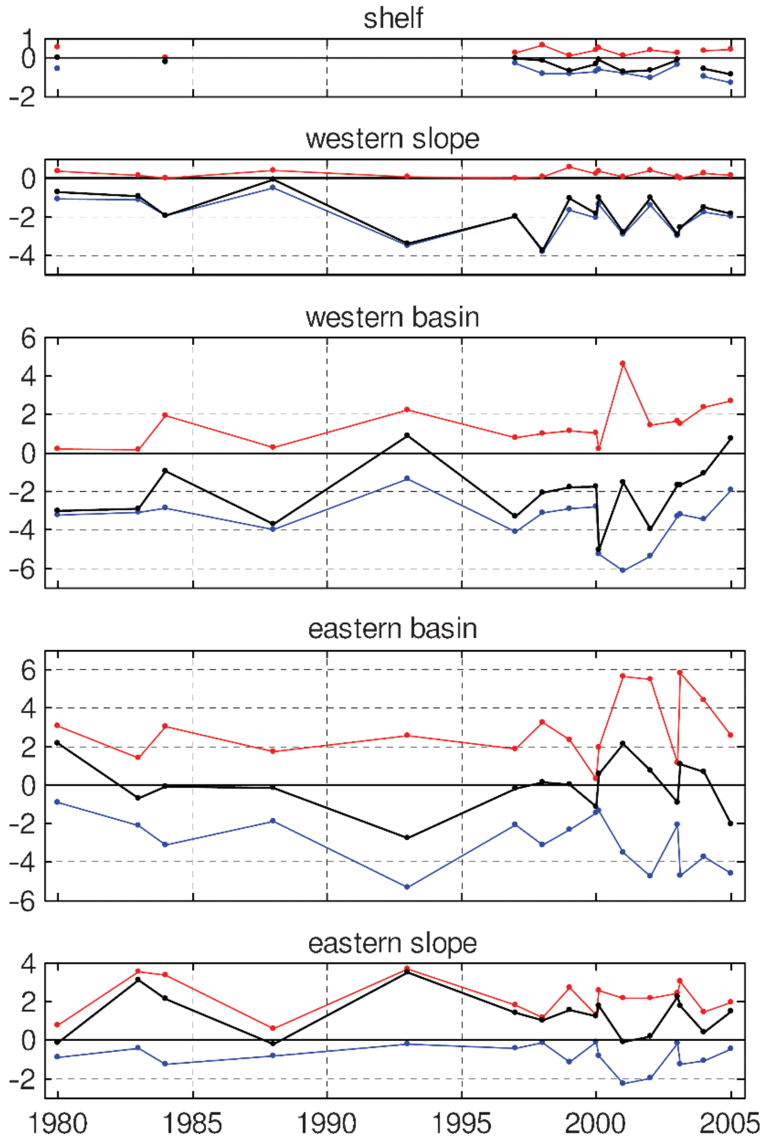

Fig. 13.4 Transports in Sv in different parts of Fram Strait between 6° W and 9° E based on geostrophic calculations. The western and eastern slopes extend down to 2,200 m and the eastern and western basins are separated by the Greenwich meridian. Inflow (red), outflow (blue) and net transport (black). The shelf transports are determined from geostrophic calculations with the velocity set to zero at the bottom

are approximately the same for all sections. East and west are separated by the Greenwich meridian, and the slopes and deep parts by the 2,200 m isobath. The eastern slope area reaches 9° E and the western slope area is taken to extend to 6° W. The fifth part is the shelf area west of 6° W, where the extent of the observations varies from year to year depending upon ice conditions. Here only surface water and occasionally Atlantic water are encountered. The Svalbard shelf east of 9° E was only observed twice. The net northward transport was in both cases less than 0.1 Sv. The transport here can therefore be ignored. The total northward and southward flow and the net transports in each part are shown in Fig. 13.4. The most consistent southward flow occurs over the western slope, and the outflow over the shelf varies between almost zero and occasionally close to 1 Sv. A net inflow is found in the West Spitsbergen Current at the Svalbard slope. The central parts generally indicate outflows, the western part more so. However, when an inflow is observed in the west, the eastern deep part shows a compensating outflow.

13.5 Variations in Water Mass Properties

So far we have considered in- and outflows but not, in detail, examined the characteristics of the water masses involved in the exchanges. Are the exchanges connected with small-scale eddy motions, which practically make the same water mass cross the section in both directions? Is there a systematic recirculation in the strait with most of the inflow taking place in one part, the outflow in another? Are there large differences between the in- and outflow characteristics, suggesting that the water masses have been long enough in the Arctic Ocean for substantial water mass transformations to occur? The total transports shown in Fig. 13.4 suggest that at least in the two central areas the exchanges largely compensate each other, and a northward transport in the west is mirrored by a southward transport in the east and vice versa. The East Greenland Current on the western slope consistently shows an outflow, while an inflow is concentrated to the West Spitsbergen Current in the east.

We only consider the main part of the strait, from 9° E to 6° W, and presently ignore the Greenland shelf, which is occupied mostly by outflowing low salinity water. The Θ–S characteristics of the northward and southward flowing water masses are determined by dividing the heat and salt transports with the volume (mass) transport in each water mass class. The transports of the different water masses in each area are indicated in Θ–S diagrams by bubble plots, where the location of the bubbles gives the Θ–S properties and their size indicates the transport. We have here included additional water masses in the classification. The surface water (SW) is sub-divided into Polar surface water (PSW) and warm Polar surface water (PSWw) by the 0 °C isotherm, and the Atlantic water (AW) is separated by the 2 °C isotherm into the colder Arctic Atlantic water (AAW) present in the Arctic Ocean and warmer Atlantic water (AW) from the south, which partly enters the Arctic Ocean, partly recirculates in Fram Strait. In the deep water ranges water more saline than 34.915 in the DWI class is defined as Canadian Basin Deep Water (CBDW) and in the DWII class as Eurasian Basin Deep Water (EBDW), while the water less saline than 34.915 in both classes is denoted Nordic Seas Deep Water (NDW) (Fig. 13.5).

Because of the widely different potential temperature and salinity ranges of the water masses we present the layers separately. The panels show the Θ–S properties for the upper, the Atlantic and the dense Atlantic, and the intermediate and deep waters respectively in all sub-areas for all years in eight Θ–S diagrams. Two Θ–S diagrams are given for each sub-area, since also the in- and outflows are shown separately. The years are distinguished by colour coding to indicate the temporal variability.

In the upper layers the difference between the areas is very distinct (Fig. 13.5a). The cold, low salinity surface water is located over the Greenland continental slope and is advected southward by the East Greenland Current. The deep western part is also dominated by outflow but the transport of PSW is smaller and the water, although still cold, is more saline and warmer than over the slope. In the eastern deep part the upper waters are warmer still and more saline. The characteristics of the northward flowing water are well clustered, while the southward transports have more varying properties. To the east, over the Svalbard slope, warm, saline and well-clustered inflows are observed, while the southward flow shows slightly more diverse characteristics and are smaller. The transports observed to the east are smaller than those to the west, especially the net transports.

The Atlantic waters over the Greenland slope mainly flow southward (Fig. 13.5b) and the low temperatures and salinities imply that water from the Atlantic layer in the Arctic Ocean here is carried out of the Arctic Ocean by the East Greenland Current. The Svalbard slope, by contrast, is dominated by northward flows and the Atlantic water is warmer and more saline. The transports here appear to be larger than over the Greenland slope.

The dense Atlantic water shows larger Θ–S variations on the Svalbard slope than on the Greenland side, where the Θ–S relations are tight except occasional years, when the recirculating Atlantic Water from the south extends onto the Greenland slope. The transports are northward over the Svalbard slope, southward over the Greenland slope and the net transports are fairly equal. The differences in Θ–S properties indicate that the Atlantic waters have become cooler and less saline, reflecting the mixing, and cooling that the Atlantic water experiences in the Arctic Ocean.

In the central parts the transports are as large as over the Svalbard slope. The range of the Θ–S characteristics between the different years found in the central areas is wider than at the Svalbard slope. The Atlantic water is slightly colder than over the slope and perhaps the western part is colder than the eastern, indicating a weak cooling and freshening from east to west. These differences are, however, smaller than the annual variability, indicating that the temporal variability over most of the strait is larger than the spatial variability across the strait. This suggests that part of the water from the West Spitsbergen Current recirculates westward in the strait on time-scales of months rather than years. In the deep central parts the inflow and outflow are of similar magnitude. The location of the in- and outflows appears to shift in time and often a large inflow in the deep western area is balanced by a strong outflow in the deep eastern area and the opposite. This is consistent with a pattern similar to that seen in the current meter array, where narrow barotropic eddies drift westward along the sill (ASOF-N Final report 2006).

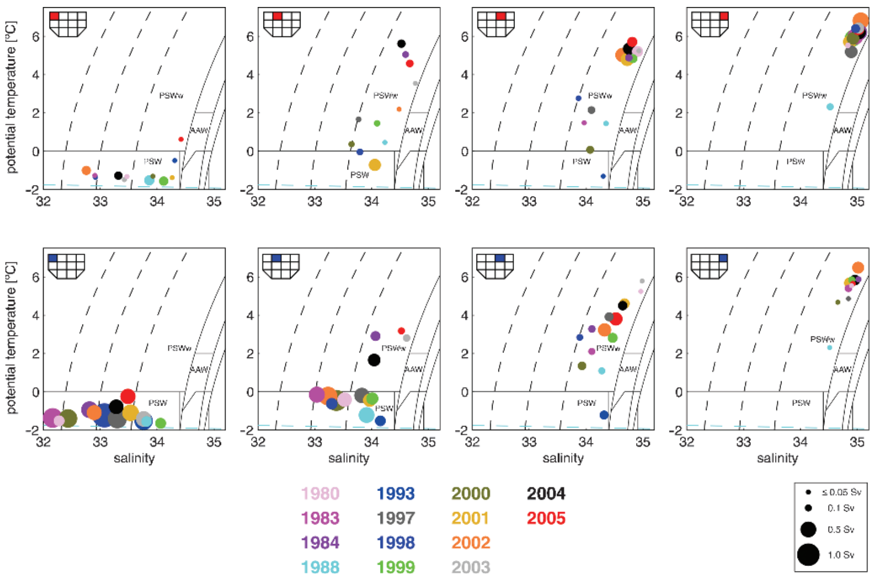

Fig. 13.5a Θ–S characteristics and transports in the surface waters for the different parts of Fram Strait. The upper four diagrams give, from left to right, the inflow over the western (Greenland) slope, the western deep part, the eastern deep part and the eastern (Svalbard) slope. The four lower panels give the outflow for the same areas. The different years are colour coded and the size of the bubbles indicates the transports. All transports ≤0.05 Sv are shown as the same size

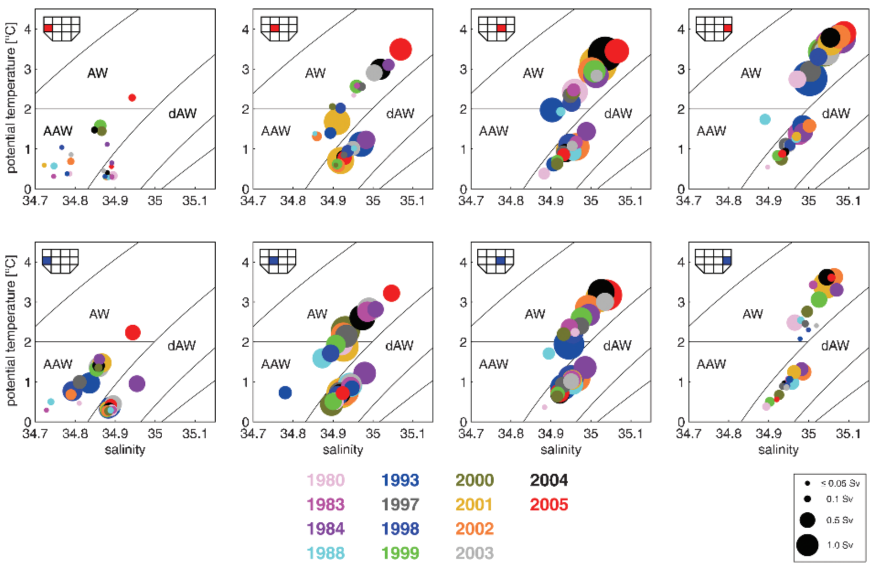

Fig. 13.5b Θ–S characteristics and transports in the Atlantic and dense Atlantic waters for the different parts of Fram Strait. The upper four diagrams give, from left to right, the inflow over the western (Greenland) slope, the western deep part, the eastern deep part and the eastern (Svalbard) slope. The four lower panels give the outflow for the same areas. The different years are colour coded and the size of the bubbles indicates the transports. All transports ≤0.05 Sv are shown as the same size

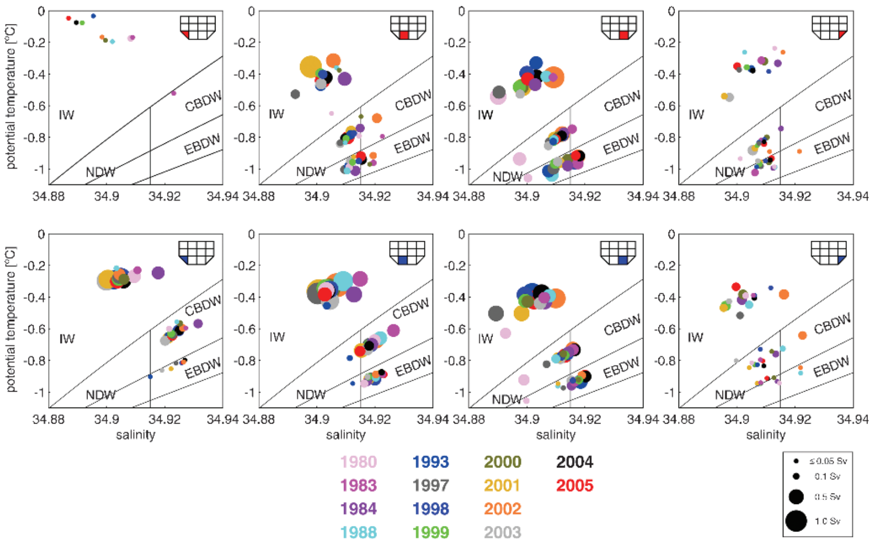

Fig. 13.5c Θ–S characteristics and transports in the intermediate and the deep-water masses for the different parts of Fram Strait. The upper four diagrams give, from left to right, the inflow over the western (Greenland) slope, the western deep part, the eastern deep part and the eastern (Svalbard) slope. The four lower panels give the outflow for the same areas. The different years are colour coded and the size of the bubbles indicates the transports. All transport ≤0.05 Sv are shown as the same size

In the intermediate and deep water ranges a similar but weaker pattern is detected (Fig. 13.5c). In the intermediate range outflow dominates over the western slope, while in the other parts of the strait the in- and outflow are about equal. The southward flows are slightly warmer and more saline, especially over the western slope, which agrees with the upper Polar Deep water (uPDW) of the Arctic Ocean being warmer and more saline than the Arctic Intermediate Water (AIW) of the Nordic seas.

The outflow of CBDW is concentrated to the western part, the western slope and the western deep area. Farther to the east the NDW becomes more prominent. The NDW dominates the inflows but also the outflows at the Svalbard slope, the inflow being stronger. In the deep areas the NSD is more strongly represented in the inflow than in the outflow. The inflow occurs mostly in the eastern but is also fairly strong in the western deep area. The EBDW is present in the outflow in both the eastern and the western deep area. However, it also takes part in the inflow, especially in the deep eastern area, suggesting some recirculation. Because of the small cross sectional areas the deep transports are comparatively small over the slopes, and the strongest deep exchanges occur in the deep areas. This can partly be explained by the larger areas, although the existence of strong, barotropic eddies could also contribute, adding recirculation to the north–south exchanges. However, the deep transports are slightly forced by the volume and mass constraints that have been imposed on the deep exchanges and too much should not be read into smaller features seen in the transports of the deep waters.

13.6 The Heat Transports

The heat transports, except in 1988, vary between above 35 TW and below 15 TW with an average around 25 TW and the mean reference temperature lies around 0.7 °C. The temperature has risen during recent years and the mean outflow temperature for the last 5 years is above 1 °C (Fig. 13.2). The time series is still rather short and contains lots of gaps in the early part of the observation period. We will therefore here not examine the time variation in transport and reference temperatures but concentrate on the mean transports and mean reference temperature. We shall especially discuss the net outflow volume and what that discloses about the distribution of the heat transported into the Arctic Ocean. For this discussion the mean heat transport (25 TW), the mean net volume flux (2.5 Sv) and the mean reference temperature (0.7 °C) are sufficient.

The obtained mean heat transport is clearly less than the >40 TW estimated from the current meter array using −0.1 °C as reference temperature (ASOF-N Final report 2006). If we adjust for the use of different reference temperatures the heat transport obtained here should be reduced by $c \times (0.7-(-0.1)) \times 2.5 \times 10^9 = 8$ TW, c being the heat capacity of sea water (4,000 J kg^{-1} K^{-1}). The difference between the results from the direct current observations and the geostrophic computations thus become larger.

The excess volume leaving the Arctic Ocean through Fram Strait must derive from the inflow over the Barents Sea and/or through Bering Strait. The Barents Sea inflow partly forms, together with the river runoff, the low salinity shelf water that eventually contributes to the low salinity surface water in the Arctic Ocean, partly supplies a denser inflow down the St Anna Trough, which cools the Atlantic water of the Fram Strait branch and forms the bulk of the underlying intermediate water mass, the upper Polar Deep Water (uPDW). The Bering Strait inflow contributes low salinity surface and upper halocline waters, which presently are mainly confined to the Canada Basin (Jones et al. 1998). The entire Pacific inflow, perhaps excluding the Bering Strait Summer Water, and about half of the Barents Sea inflow are eventually cooled to freezing temperatures within the Arctic Ocean. The denser St Anna Trough inflow is cooled at least to below zero in the Barents Sea and we tentatively set this deep inflow to 1.2 Sv with temperature −0.5 °C.

This is slightly larger than the 0.75 Sv of dense water that Schauer et al. (2002) estimated passing between Novaya Zemlya and Franz Josef Land. However, Schauer et al. only give the transports with temperature below 0 °C both for the dense deep water, 0.75 Sv, and the less dense surface water, 0.75 Sv. To have volume balance the rest of the inflow through the Barents Sea opening must either pass between Franz Josef Land and Novaya Zemlya with temperatures above 0 °C, or enter the Arctic Ocean west of Franz Josef Land and the Kara Sea south of

Novaya Zemlya. We have therefore increased the deep inflow estimate given by Schauer et al. (2002) from 0.75 Sv to 1.2 Sv and the less dense part from 0.75 Sv to 1 Sv. The less dense part will eventually be cooled to freezing temperature, and we do not expect any high temperatures to be present in the deep inflow and the postulated −0.5 °C should be a reasonable mean temperature for the denser inflow to the Arctic Ocean over the Barents Sea. The details of these "known" transports are summarized in Fig. 13.1.

The net outflow Y Sv through Fram Strait would then comprise 1.2 Sv of intermediate water with temperature −0.5 °C, since the deep Barents Sea inflow can only exit through Fram Strait, and (Y – 1.2) Sv of surface water at the freezing point (the seasonal heating of the surface water is ignored). To attain the mean outflow temperature – the reference temperature – the temperature of the cold, net outflow Y has to be compensated by a comparably warm return flow of Fram Strait branch Atlantic water. In a heat balance based on the mean outflow temperature in Fram Strait the amount F, $F = c \times (T_{out} - T_f) \times (Y - 1.2) + c \times (T_{out} - (-0.5)) \times 1.2.$, of the inflowing heat has to be used to increase the temperature of the excess volume Y to the mean outflow temperature T_{out}. Again c is the heat capacity of seawater (4,000 J kg^{-1} K^{-1}) and T_f the freezing temperature (−1.8 °C). Taking the mean outflow temperature (0.7 °C) and the mean net outflow volume Y = 2.5 Sv F becomes ~19 TW. If instead all the added water would be upper layer water the heat needed to compensate for the outflow becomes 26 TW and if all added water is upper Polar Deep water 14 TW is required. If choosing a mean temperature of the deeper outflow to 0 °C or −1.0 °C the corresponding heat requirement becomes 17 TW and 22 TW respectively.

A large heat loss of the inflowing Atlantic water occurs in the area just north of Svalbard, the Whalers' Bay. The heat is lost to ice melt and to the atmosphere, and Rudels et al. (1999a) suggested that when ice is melting on warmer water and the air temperature is below the freezing temperature of sea water, the heat loss of the ocean is distributed in such a way that the ice melt rate is a minimum. With a linear equation of state this implies that the fraction, f, of the heat loss that goes to ice melt is given by $f \approx 2\alpha L (c\beta S_A)^{-1}$. S_A is the salinity of the underlying water, L (336,000 J kg^{-1}) is the latent heat of melting and α and β are the coefficients of heat expansion and salt contraction respectively (Rudels et al. 1999a). About one third of the oceanic heat loss then goes to ice melt. The ice melt dilutes the upper part of the inflowing Atlantic water and creates an upper layer with lower salinity, ~34.3, which in the Nansen Basin is cooled to freezing temperature in winter and homogenised down to the Atlantic layer by (mainly) haline convection (Rudels et al. 1996; Rudels et al. 2005). Farther to the east this mixed layer is overrun by less saline and less dense shelf water and becomes the Fram Strait branch lower halocline (Rudels et al. 1996; Rudels et al. 2004).

Untersteiner (1988) estimated the formation of low salinity upper water in Whalers' bay due to ice melt to at least 0.5 Sv. The estimated salinity in the water was less than the ~34.3 normally encountered in the area, and the amount of low salinity water created north of Svalbard is probably larger. We shall assume a formation rate of 0.7 Sv, and using the difference, ~5 K, between the entering Atlantic water temperature T_A ~3 °C and the freezing temperature, the amount of inflowing oceanic heat lost during the initial formation of the lower halocline water can be estimated from $c \times (T_A - T_f) \times 0.7 \times 10^9$ to 14 TW. Here the heat going to ice melt as well as that being lost to the atmosphere is accounted for.

The amount of ice melted, I, can be found in two ways. Either by computing the dilution of the Atlantic water from salt conservation $(0.7 + I) \times 34.3 = 0.7 \times 35$, giving I = 0.015 Sv, or by using the expression from Rudels et al. (1999a) (given above) for the fraction of heat going to ice melt. With $\alpha = 0.6 \times 10^{-4}$ and $\beta = 8 \times 10^{-4}$ f becomes 0.36, and the heat lost to ice melt $0.36 \times 14 = 5$ TW. This also corresponds to a melting rate of 0.015 Sv. By contrast, Untersteiner (1988) deduced a much larger melting rate, 0.06 Sv, in Whalers' Bay, based on ice transport estimates by Vinje and Finnekåsa (1986).

These two heat sinks then use most (all) of the heat advected into the Arctic Ocean. In some years the heat loss is larger, in some years it is smaller than the heat import. This points to a further factor to consider in the Arctic Ocean heat balance, the change in temperature in the Atlantic layer in the Arctic Ocean. The higher temperatures of the Atlantic layer, first noticed in the early 1990s (Quadfasel et al. 1991), suggest an increase in heat storage in the Arctic Ocean. The continued studies in the Arctic Ocean have shown that this warm inflow pulse lasted perhaps close to a decade and gradually spread around the gyres in the different basins. Return flows were encountered in the northern Nansen Basin and in the Amundsen Basin (Rudels et al. 1999b), along the Lomonosov Ridge (Swift et al. 1997). It was observed in the Makarov Basin, first at the Siberian continental slope and at the Mendeleyev Ridge (Carmack et al. 1995), and then around the basin, and presently it is returning along the Lomonosov Ridge from North America towards Siberia (Kikuchi et al. 2005). The pulse also penetrated from the Chukchi Cap into the northern Canada Basin (Smethie et al. 2000). The spreading into the southern Canada Basin appears to occur differently (Shimada et al. 2004), perhaps through interleaving structures (Carmack, 2006) rather than circulating along the continental slope. Similar ideas have been advanced for the spreading of heat from the boundary current into the central Nansen Basin (Carmack et al. 1997; Swift et al. 1997). For the present discussion the spreading mechanisms are of little importance.

The long, warm inflow event was eventually followed by the arrival of colder Atlantic water. A comparison between sections taken in Fram Strait 1984 and 1997 (e.g. Rudels et al. 2000; Rudels 2001) indicate that a cooling and freshening of the Atlantic water has taken place. This is perhaps not so obvious in the time series from Fram Strait (Fig. 13.2) because of the gaps in the time series between 1984 and 1997 and because after 1997 the temperature gradually increases, indicating that the cold pulse has passed. The presence of colder water was noticed at the NABOS moorings north of the Laptev Sea in 2002 (Dmitrenko et al. 2005; Polyakov et al. 2005). Another warm pulse was observed around 2000 in Fram strait (ASOF-N Final report 2006) and a sudden, strong increase in the Atlantic water temperatures was detected at the NABOS moorings in 2004 (Dmitrenko et al. 2005; Polyakov et al. 2005). Still warmer Atlantic water was observed in Fram

Strait in 2004 suggesting the arrival of another warm inflow pulse. This pulse was found to partly recirculate in Fram Strait (ASOF-N Final report 2006).

The temperature increase in the Atlantic layer in the Arctic Ocean is uneven. Some of the warm Atlantic water has already left the Arctic Ocean and the rest is redistributed around the different gyres. Roughly assessing an overall temperature increase of 0.3 °C over a 200 m thick Atlantic layer over 15 years, this corresponds to a storage rate of 2 TW. Polyakov et al. (2004) estimated the change in heat content of the Atlantic layer between 1970s and the late 1990s as 4.3×10^8 J m^{-2}, which corresponds to 2.7–3.4 TW, reasonably close to the back of the envelope calculation above.

13.7 Freshwater Transports

The freshwater export estimated relative to the inflow salinity has three components, the salinity difference between the in- and outflows, and the volume transport, which can be separated into two parts: one part corresponding to the inflow volume, and a second part representing the net outflow volume. The inflow salinities range between 34.8 and 35 but cluster around 34.92. The outflow salinity tends to co-vary with the inflow salinity and averages around 34.8 (Fig. 13.2).

As with the heat transport we can consider the freshwater export partly as a dilution of the inflow, partly as the addition of water from other sources with different freshwater content. The freshwater outflow, excluding 1988, ranges between 0.02 and 0.1 Sv, is highly variable but the mean appears to be somewhere between 0.03 and 0.05 Sv. Almost all the freshwater export occurs in the surface water, suggesting that the Barents Sea inflow, combined with river runoff and ice melt, contributes most of the net outflow volume with occasionally some Bering Strait inflow water added. The dilution of the upper part of the Fram Strait inflow to 34.3 is mainly due to ice melt and creates 0.7 Sv of halocline water (Section 13.6) but only 0.015 Sv. of freshwater is added by this process.

The outflowing Arctic Atlantic water (AAW) is, as expected, less saline than the inflowing Atlantic water. In fact, the crossover point in a Θ–S diagram, where the Arctic Ocean water column changes from being less saline than the entering Nordic Seas water column to becoming more saline than the Nordic Seas water column occurs close to 0 °C, which, according to our water mass definitions, separates dense Atlantic water (dAW) from the intermediate water.

The inflowing deep waters are less saline than the reference salinity and the deep inflow will add freshwater to the Arctic Ocean (Fig. 13.5c). This freshwater is largely re-exported by the outflowing Arctic Ocean deep and intermediate waters. Only if the reference salinity lies between the deep inflow and outflow salinities will both deep transports result in a freshwater flux into the Arctic Ocean. The salinity anomalies are then small and no large deep freshwater transports take place. The freshwater flux below the Atlantic layer is thus small and can safely be ignored.

In the Arctic Ocean the dilution of the entering Atlantic water occurs by ice melt north of Svalbard and perhaps also, but to a much smaller degree, in the entire Nansen Basin. A freshening of the Atlantic layer core takes place through convection of cold, dense shelf water, reaching the Atlantic layer. This occurs north of Svalbard (Rudels et al. 2005) and also at the Barents Sea slope between Svalbard and Franz Josef Land (Rudels 1986; Schauer et al. 1997). This freshwater input is restricted to the Atlantic layer. For the slope convection to reach deeper the initial salinities on the shelf have to be higher than the salinity of the Atlantic water and no freshwater is exported to the deeper layers.

The major freshening occurs downstream of the St Anna Trough. Here the denser part of the Barents Sea branch inflow joins the boundary current. It forms a colder and less saline water column extending from the surface to about 1,200 m. It is initially confined to the slope and depresses the deep isopycnals and the denser underlying Arctic Ocean deep water (Schauer et al. 1997). The upper part derives from the mixed layer in the eastern Barents Sea and the northern Kara Sea. Like the mixed layer in the Nansen Basin, it is initially formed by sea ice melting on warm Atlantic water (Rudels et al. 2004). The denser part of the inflow eventually mixes with the Fram Strait branch, cools and freshens the Atlantic core and creates the intermediate salinity minimum observed in the Eurasian Basin (Rudels and Friedrich 2000).

Farther to the east the river runoff and the rest of the Barents Sea inflow enter the central basins as low salinity shelf water, capping the boundary current and reducing its interaction with the sea surface and the ice cover. The mixed layer of the Nansen Basin and the boundary current deriving from the Fram Strait branch, as well as the mixed layer of the Barents Sea branch, are then covered by less saline water, the Polar Mixed Layer (PML), and become halocline waters. The two lower halocline waters as well as the Atlantic derived part of the Polar Mixed Layer return towards and exit through Fram Strait, although the Barents Sea branch halocline water moves along the North American slope and partly passes through the Nares Strait, contributing to the deep and bottom waters of Baffin Bay (Rudels et al. 2004).

The Bering Strait inflow provides the second largest freshwater source to the Arctic Ocean, larger than the net precipitation and almost as large as the river runoff (Woodgate and Aagaard 2005; Serreze et al. 2006). It supplies most of the water that passes through the straits in the Canadian Arctic Archipelago into the Baffin Bay. Pacific water also exits the Arctic Ocean through Fram Strait (Jones et al. 2003) but not continuously. Some years Pacific water is absent (Falck et al. 2005). The main contributions to the liquid freshwater transport through Fram Strait then come from river runoff and from the Barents Sea inflow, mainly the Norwegian Coastal Current. Some ice melt is exported in the halocline but this is likely to be a smaller part, ~0.015 Sv, if the same estimates as for the heat transport are used.

A considerable fraction of the Arctic Ocean freshwater export occurs as ice and about 90% of the ice export from the Arctic Ocean is estimated to pass through Fram Strait (Vowinckel and Orvig 1970). The passages in the Canadian Arctic Archipelago are narrow and often blocked by landlocked ice. In the northern

Barents Sea the opening between Svalbard and Franz Josef Land usually freezes early in fall, and the ice cover prevents the multi-year ice from the Arctic Ocean to pass into the Barents Sea. Occasionally it happens, but the sea ice, as well as the low salinity water of the East Spitsbergen Current, will be brought northward by the West Spitsbergen Current to the Arctic Ocean and are not really exported. A problem for the volume balance could therefore arise, if this transport is measured and included in the inflow but not accounted for as an outflow. Its contribution might be as large as 1 Sv, at least in winter (Rudels et al. 2005).

Freezing extracts freshwater from the surface water and the sea ice comprises river runoff from the Siberian shelves as well as water drawn from the Pacific water and the runoff from the North American continent. The ice gradually thickens, as it is advected towards Fram Strait indicating that freshwater is extracted from the PML in the entire Arctic Ocean. It is also likely that the net precipitation on the Arctic Ocean mainly falls on the sea ice and ends up in the solid phase, not in the water column.

How the freshwater export is distributed between the liquid and solid phases in Fram Strait has, so far, not been determined. Commonly the ice export has been assumed the largest, and results from ASOF-N indicate that the ice export in Fram Strait could be three times the liquid freshwater export (ASOF-N Final report 2006). However, tracer studies have suggested that the liquid freshwater export could be as large or larger than the ice export (Meredith et al. 2001). The results from the geostrophic calculations here indicate large variability in the liquid freshwater export, ranging from 0.1 Sv, which is close to the most cited value (0.09 Sv) for the ice export, down to 0.01 Sv. The mean value (~0.04 Sv) is close to that obtained by direct measurements in ASOF-N. However, the transport estimates given so far are for the standard section between 9° E and 6° W. The transport over the Greenland shelf has, because of the different extent of the section during different years, to be estimated separately. The transport over the shelf, which only comprises low salinity upper waters, is occasionally almost as large as the outflow of upper water in the rest of the strait, while in other years it is much weaker (compare Figs. 13.3. and 13.4.). Taking the mean of the freshwater transports over the shelf from the existing shelf sections (not shown) we get 0.025 Sv, which, added to the 0.04 Sv obtained for the strait proper, increases the freshwater flux to 0.065 Sv, or almost 75% of the ice export.

The reference salinity has been determined only for the deep part of the strait, and even if some inflow occurs on the shelf, it mainly involves a recirculation of the same low salinity water masses, which derive from passages other than Fram Strait. The choice of reference salinity, based on the inflow salinity, would therefore be the same, also when the shelf transports are included. The fact the transports over the shelf were excluded, when the reference temperature was determined, should also not seriously affect the discussion about the heat balance given above. The transports over the shelf almost exclusively involve waters from other passages than Fram Strait, which have lost their heat to the atmosphere being cooled to freezing temperature within the Arctic Ocean. They therefore say more about the fate of the heat fluxes through the Bering Strait and the Barents Sea than about the distribution of the heat transport through Fram Strait.

13.8 Updated Fram Strait Exchanges

The obtained freshwater transports can be used, together with external information about the freshwater budget, to re-examine the calculated volume transports through Fram Strait. A freshwater budget for the Arctic Ocean and for the Nordic Seas has recently been compiled by Dickson et al. (2007). Taking the values given in Dickson et al. (2007) for runoff, net precipitation, the Bering Strait inflow, the inflow through the Barents Sea opening, the export through the Canadian Arctic Archipelago, and the ice export, recomputed to the mean reference salinity (34.92) applied here, we obtain the transports presented in Table 13.3. The Fram Strait net outflow has been increased to 2.8 Sv as compared to 2.5 from Fig. 13.2 to accommodate the transport over the shelves. It should also be mentioned that the outflow through the Canadian Arctic Archipelago in this estimate is 0.25 Sv lower than the most often cited value, 1.7 Sv (Prinsenberg and Hamilton 2005).

The freshwater budget is balanced to within 2%, while the volume budget indicates a large net outflow through Fram Strait. The use of geostrophy underestimates the transports, since strong barotropic current will not be adequately accounted for. In Fram Strait the exchanges are known to have large barotropic components, in the central part of the strait as well as in the two main currents, the West Spitsbergen Current and the East Greenland Current. The applied constraints, combined with the requirement of minimum added kinetic energy in the deep exchanges, obviously cannot reproduce the barotropic transports.

However, the East Greenland Current is more stratified than the West Spitbergen Current and likely to be more baroclinic and better represented by the geostrophic computations. We therefore hypothesize that the volume imbalance in Fram Strait is due solely to underestimation of the inflow volume. By adding 1.1 Sv with the mean inflow characteristics to the inflow, we obtain an approximate balance also in volume. Since the inflow salinity is the same as the reference salinity this will not affect the freshwater balance, which continues to hold.

Table 13.3 Volume fluxes, salinity and freshwater fluxes Black numbers from Dickson et al. (2007) red numbers from this work

Contribution	Volume (Sv)	Salinity	Freshwater (mSv)
Runoff	0.1	0	102
Net precipitation	0.065	0	65
Bering Strait	0.8	31.49	79
Barents Sea	2.2	34.84	4
Canadian AA	−1.44	32.7	−92
Fram Strait ice export	−0.09	4	−88
Fram Strait net outflow and liquid freshwater export	−2.8	–	−65
–	–	–	–
Net transport	−1.17	–	−5

However, the heat transport through the strait, and the distribution of the heat within the Arctic Ocean will change. The average difference between the inflow temperature (1.6 °C) and the outflow (reference) temperature is 0.9 K. This gives $0.9 \times 1.1 \times 10^9 \times 4{,}000 \approx 4$ TW and the average transport of heat through Fram Strait into the Arctic Ocean increases from 25 TW to 29 TW. The net outflow that has to be heated to the reference temperature is reduced from 2.5 to 1.7 Sv, and if we keep the estimate of 1.2 Sv of intermediate water added by the deep Barents Sea inflow at −0.5 °C, only 0.5 Sv of surface water at freezing temperature needs to be heated to 0.7 °C. The amount of heat required to warm the net outflow volume then becomes 10 TW. The formation of 0.7 Sv. of halocline water still needs 14 TW and the heat storage rate remains 2 TW. This leaves 3 TW to be lost to the atmosphere, which corresponds to a surface heat transfer of 0.6 W m^{-2}, much less than the 2 W m^{-2} often quoted for the oceanic heat loss to the atmosphere (Maykut and Untersteiner 1971; Maykut 1986). The resulting mass, heat and freshwater transports are summarized in Fig. 13.6.

We may also note that by adding the net outflow of low-density surface water to the halocline water, formed by the entering Atlantic water, the export of low

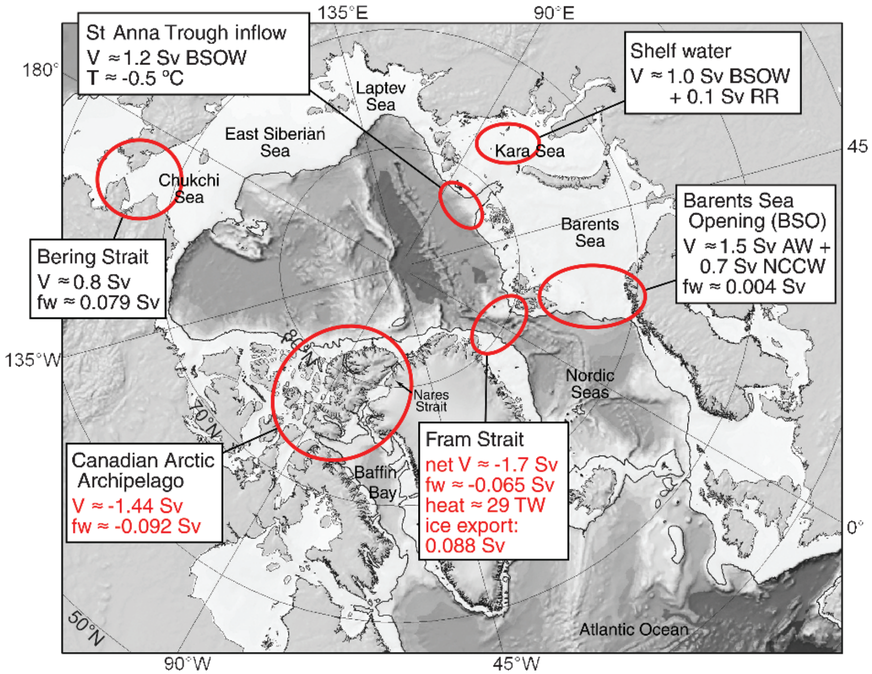

Fig. 13.6 Volume and freshwater balances for the Arctic Ocean. The outflows through the Canadian Arctic Archipelago and the Fram Strait ice export are taken from Dickson et al. (2007), while the net outflow, the heat transport and the export of liquid freshwater through Fram Strait are based on the discussions in the present work

salinity upper layer water becomes 1.2 Sv. This is 0.2 Sv less than the low salinity outflow through the Canadian Arctic Archipelago (Table 13.2). It is somewhat low, but about the same as was calculated in Dickson et al. (2007), and the number is not unreasonable. The estimate of the rate of halocline water formation is a guess and it might be smaller or larger. The surface water fraction provided by the Barents Sea could also be larger than the 1 Sv used here. However, it is not possible to extract more information from these data without becoming excessively speculative and it is time to stop.

13.9 Summary

Transports of volume, heat and freshwater through Fram Strait have been determined from geostrophic velocities computed on sixteen hydrographic sections taken in the strait between 1980 and 2005. To find the unknown reference velocities the deep water exchanges have been determined, which have the least kinetic energy while fulfilling prescribed volume transport and salt transport constraints in the deeper layers. The obtained northward and southward transports are smaller than those estimated from the current meter array, while the net southward transport is larger.

The heat and freshwater fluxes through the strait are calculated relative to the mean outflow temperature and the mean inflow salinity on each section. This choice of reference values removes the northward transport of freshwater and the southward transport of heat through the sections, but it leads to varying reference temperatures and reference salinities. In this study only the mean reference salinity and mean reference temperature over the observation period have been used.

The computed liquid freshwater export, combined with existing estimates of other freshwater sources and sinks in the Arctic Ocean (e.g. Serreze et al. 2006; Dickson et al. 2007), shows that the freshwater transport in Fram Strait almost fulfils the freshwater balance and thus appears realistic. However, there is an imbalance in the volume fluxes, and the net volume export through Fram Strait is found to be too large. As a remedy we hypothesize that the inflow through Fram Strait is underestimated by the geostrophic calculations, and an inflow through Fram Strait, with the mean inflow characteristics, is added to establish volume balance in the Arctic Ocean.

Since the northward and southward transports in Fram Strait do not balance, a unique heat transport through the strait cannot be found. However, the net outflow volume can be examined separately. This simplifies the interpretation of the heat transport, because most of the water that enters through the other passages is less dense surface water that is cooled to freezing point in the Arctic Ocean. The only exception is the large fraction of the Barents Sea inflow, which is dense enough to supply the intermediate layer. This volume has been set to 1.2 Sv at −0.5 °C. These considerations then allow for a discussion of the fate of the heat entering the Arctic Ocean through Fram Strait. As long as all inflows and outflows are not successfully monitored, such approach should provide some insight on the importance of Fram Strait for the Arctic Ocean heat budget.

Much of the barotropic transports that dominate the deep water exchange may be associated with barotropic eddies, implying that the deep water exchange between the Arctic Ocean and the Nordic Seas is smaller than the direct current observations indicate. The geostrophic transports, since they depend upon the density differences between the northward and southward flowing waters, can be seen as mirroring the effects of the water mass transformation processes active in the Arctic Ocean and in the Nordic Seas. This then describes the transport of the water in Θ–S space and thus partly represents the oceanic transport having impact on climate. The fact that additional constraints are needed to obtain a realistic volume balance for the exchanges between the Arctic Ocean and the world ocean shows that the transports through Farm Strait are not just caused by the density changes, but are also forced by large-scale wind fields and sea level slopes. The variational approach applied here, which minimizes the kinetic energy of the exchanges, will remove, or at least diminish, this "external" forcing and thus require additional constraints or information on the freshwater and/or volume transports to become realistic.

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