Chapter 3
Variation of Measured Heat Flow Through the Fram Strait Between 1997 and 2006

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3.1 Introduction

The northernmost extension of the Atlantic-wide overturning circulation consists of the flow of Atlantic Water through the Arctic Ocean. Two passages form the gateways for warm and saline Atlantic Water to the Arctic: the shallow Barents Sea and the Fram Strait which is the only deep connection between the Arctic and the World Ocean. The flows through both passages rejoin in the northern Kara Sea and continue in a boundary current along the Arctic Basin rim and ridges (Aagaard 1989; Rudels et al. 1994). In the Arctic, dramatic water mass conversions take place and the warm and saline Atlantic Water is modified by cooling, freezing and melting as well as by admixture of river run-off to become shallow Polar Water, ice and saline deep water. The return flow of these waters to the south through the Fram Strait and the Canadian Archipelago closes the Atlantic Water loop through the Arctic.

In the past century the Arctic Ocean evidenced close relation to global climate variation. Global surface air, upper North Atlantic Waters and Arctic intermediate waters showed coherently high temperatures in the middle of the last century and also in the past decades (Polyakov et al. 2003; Polyakov et al. 2004; Delworth and Knutson 2000). A likely candidate for this tight oceanic link is the flow through the Fram Strait. Through the Barents/Kara Sea, only the upper layer (200 m) of Atlantic Water can pass – thereby loosing much of its heat to the atmosphere – while the Fram Strait (sill depth 2,600 m) is deep enough to enable the through-flow of Atlantic Water at intermediate levels.

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Two currents carry the warm water from the North Atlantic to the Fram Strait: a western branch which is a baroclinic jet in the Polar Front between the Atlantic Water and the central waters of the Nordic Seas, and an eastern branch, called Norwegian Atlantic Slope Current, which is an almost barotropic current along the Norwegian shelf break (Skagseth et al. 2008). They converge in the Fram Strait to form the West Spitsbergen Current (Walczowski and Piechura 2007) but the difference in both their origin (Hansen and Østerhus 2000) as well as their speed and pathway in the Nordic Seas affect their respective impact on the Arctic Ocean.

The complex topography in the Fram Strait itself leads again to a splitting of the West Spitsbergen Current and to a distribution of the Atlantic Water in at least three branches (Quadfasel et al. 1987a). One branch follows the shelf edge and enters the Arctic Ocean north of Svalbard. This branch crosses the Yermak Plateau which limits its depth to approximately 600 m. A second branch flows northward along the northwestern slope of the Yermak Plateau and the third branch recirculates immediately in the Fram Strait between 78° N and 80° N (Perkin and Lewis 1984; Gascard et al. 1995). Evidently, transports and properties of the different branches determine the input of oceanic heat to the Arctic Ocean. While part of the Atlantic Water flows to the central Arctic and is likely to be responsible for observed changes in heat content there, another part returns in a short loop within the northern Fram Strait. Here it can induce ice melt and thus determines the fractions of fresh water entering the Nordic Seas as ice and as water.

On the western side of the Fram Strait, modified Atlantic Water that originates from the West Spitsbergen Current as well as from the Barents Sea (Rudels et al. 1994) leaves the Arctic Ocean augmented by much of the Arctic fresh water surplus both as ice and in liquid form and occasionally some Pacific Water (Falck et al. 2005). This accumulates to a net southward volume transport through the Fram Strait of approximately 2 Sv (Fahrbach et al. 2001).

In the past few decades, the Atlantic Water flowing into the Arctic was not only warmer than earlier (Quadfasel et al. 1991; Schauer et al. 2004) but the influence of Atlantic Water in the Arctic Ocean also became more widespread: in the 1990s the front separating saline Atlantic-derived upper-ocean water from less saline Pacific-derived waters shifted from the Lomonosov Ridge to the Alpha–Mendelyev Ridge (McLaughlin et al. 1996). These changes, together with a reduced ice cover were attributed to a stronger cyclonic atmospheric circulation over the North Atlantic and the Arctic (Dickson et al. 2000). Morison et al. (2006) described the return of the Atlantic Water distribution and properties to near pre-1990s climatology after the cyclonic atmospheric circulation had relaxed. In the same time sea-ice extent continued to decrease and in the late 1990s another warm pulse of Atlantic Water entered the Arctic Ocean that was seen to propagate around the Eurasian Basin (Polyakov et al. 2005).

While the advection of warm Atlantic Water through the Fram Strait has been known since Nansen (1902) its role in the overall heat budget of the Arctic, as well as the role of its anomalies, are not yet understood. Morison (1991) pointed out that the ocean heat transport to the Arctic is an order of magnitude smaller than that of the atmosphere but that it might still be an important contribution to a delicate balance. Thus, one of the tantalizing current questions is whether there was an oceanic contribution to the decrease of the Arctic sea ice of the past decades. In large parts of the Arctic Ocean the Atlantic layer is shielded from sea ice and atmosphere by a fresh surface layer. In and northeast of the Fram Strait, however, the warm water is close to the surface (Aagaard et al. 1987). It may undergo several freezing/melting cycles during its travel in the boundary current along the Eurasian shelf edge (Rudels et al. 1996) before it meets the fresh surface layer that in the Eurasian Arctic is mainly fed by Siberian river-runoff. A shift in the circulation of the run-off on the shelves may increase the area where heat can be released directly from the Atlantic layer to the ice or atmosphere (Martinson and Steele 2001).

An assessment of the warm Atlantic Water impact to the Arctic Ocean – being either a transient feature or a contribution to the surface heat budget – can be made by relating its inflow to its outflow. First estimates of the oceanic heat budget of the Arctic Ocean suffered from a lack of exact volume flux data (Mosby 1962) and also from the erroneous assumption that the volume flux through the Fram Strait is balanced (Aagaard and Greisman 1975). Later attempts did not always follow the concept of heat flux computation in a stringent way so that an evaluation of earlier Fram Strait heat flux computations is difficult.

A prerequisite for the computation of oceanic heat transport is the knowledge of the volume fluxes. Past estimates of transport through the Fram Strait derived from observations were either based on inverse modeling or on velocity measurements at few locations requiring considerable extrapolations. A method-induced bias seems to result in lower volume fluxes from the inverse method (e.g. Schlichtholz and Houssais 1999) than from direct current measurements (e.g. Hanzlick 1983).

In order to examine the exchange of water through the Fram Strait, to quantify the heat transported with the Atlantic Water to the Arctic, and to better understand the mechanisms involved in its variation an intensive mooring programme was established in 1997. An array consisting of 14–16 moorings, covering the Fram Strait from the eastern to the western shelf edge, allows to resolve the complex flow structure. Since 2000, yearly hydrographic surveys took part between 70° N and 79° N. Here we report the results from these observations that form a unique time series of long-term year-round high resolution flux measurements through a key gateway to the Arctic.

3.2 Data

The results reported here are based on a set of regularly repeated observations carried out in the West Spitsbergen Current between 70° N and 79° N in the past decade (Fig. 3.1). Until 2005 the observations were done in the framework of the European Union projects VEINS (Variability of Exchanges in Northern Seas, 1997–2000) and ASOF-N (Arctic–Subarctic Ocean Fluxes, 2002–2005). Since 2006, the work is carried out as a part of EU-DAMOCLES (Developing Arctic Modelling and Observing Capabilities for Long-term Environment Studies).

An array of moorings measuring currents, temperature and salinity has been maintained along 78° 50′ N to 79° 00′ N since 1997. The instruments were RCM7, RCM8 or DCM11 from Aanderaa Instruments, ADCPs from RDI and 3D-ACM from Falmouth Scientific Inc; all registered velocity and temperature at 2-h intervals.

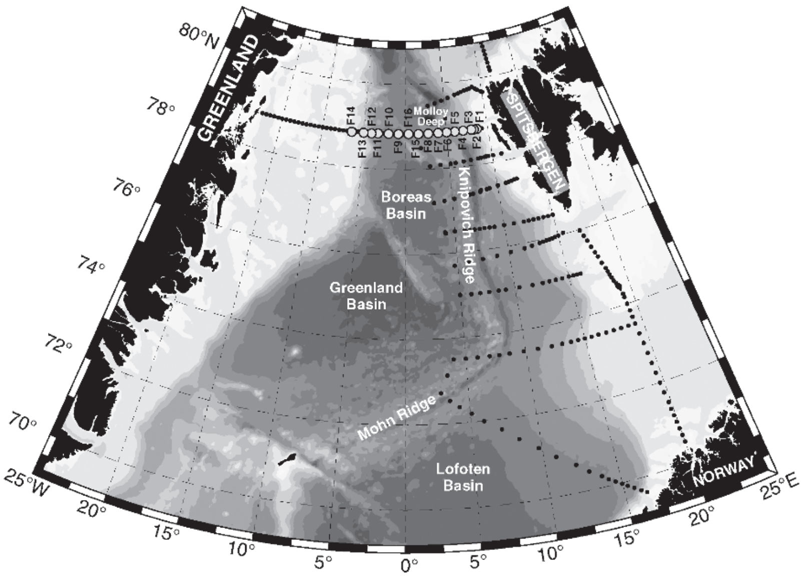

Fig. 3.1 Location of the measurements in the northern Nordic Seas and the Fram Strait between 1997 and 2006. Filled gray circles mark mooring positions during the period September 2002–August 2006; for the respective positions between September 1997 and September 2002, see Fig. 3.2. Black dots show CTD stations taken in August/September. The section overlapping with the moorings along 78° 50′ N was surveyed in the summers 1997–2006. The other sections were taken from 2000 to 2006

The instruments covered the water column from 10 m above the seabed to approximately 50 m below the surface (Fig. 3.2). The measurements extended from 6° 51′ W, the eastern Greenland shelf break, along 79° N to 0° E and continued along 78° 50′ N to 8° 40′ E, the western shelf break off Spitsbergen; since 2002 all moorings were deployed along 78° 50′ N.

The number of moorings and the number of levels equipped with instruments varied over the years (Fig. 3.2). We started with 14 moorings with a relatively narrow horizontal spacing of the moorings over the continental slopes where strong horizontal gradients were expected, and a wider spacing in the interior. It turned out that in this way the return current in the central part of the strait was under-sampled resulting in significant aliasing. Therefore, from 2002 onwards, the number of moorings was increased to 16. In addition, instruments were included at the 750 m level to better identify the lower boundary of the warm Atlantic Water. For more details, see ASOF_N deliverable 6.3. For a description of the data processing we refer to Fahrbach et al. (2001) and Schauer et al. (2004).

The year-round measurements from moored instruments were combined with hydrographic sections, taken along the mooring section during the deployment cruises since 1997 and in addition between 70° N and 79° N since 2000. On all cruises, a Seabird 9/11 CTD system was used. To obtain the horizontal distributions the data were interpolated using the kriging procedure (Walczowski and Piechura 2006). The grids were smoothed with a linear convolution low-pass filter.

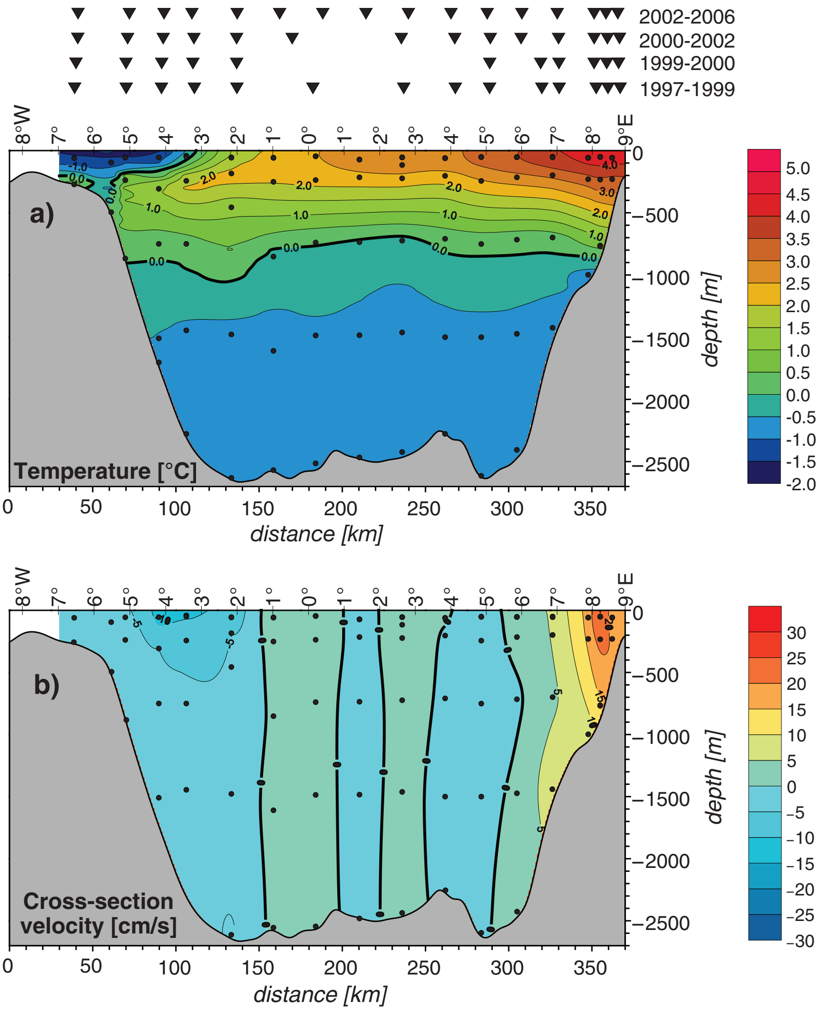

Fig. 3.2 Mean temperature (a) and cross-section velocity (b) distribution for the period September 2002–August 2003 measured from the mooring array. Dots denote the positions of instruments. Triangles on top mark the mooring positions in the different periods

3.3 Flow and Temperature Evolution

3.3.1 Atlantic Water Branches Along the Continental Slope and the Polar Front

Between 72° N and the southern tip of Svalbard, the western branch transporting Atlantic Water along the Arctic front can be derived from the mean baroclinic field while the slope current – due to its barotropic nature – can be identified only from

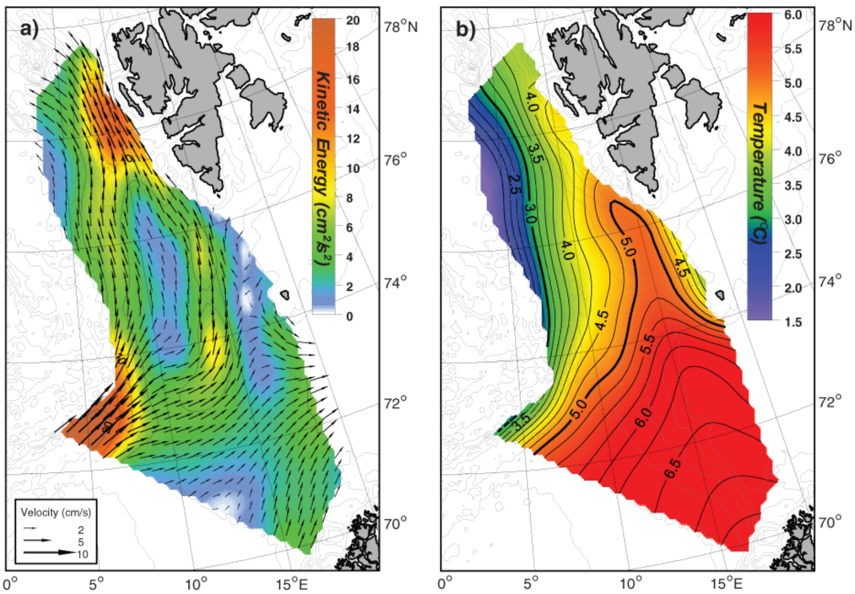

Fig. 3.3 (a) Mean kinetic energy (cm^2/s^2) of baroclinic currents (color scale) and baroclinic currents (arrows) at 100 dbar in the summers 2000–2006. The reference level is 1,000 dbar or the bottom. (b) Distribution of the summer temperature at 100 dbar averaged over the years 2000–2006. The 3 °C and 5 °C isolines are in bold. (Walczowski and Piechura 2007, Fig. 3.2)

the zonal temperature maximum (Fig. 3.3). The western branch is colder than the slope branch due to the difference in temperature of the branches at the entrance to the Nordic Seas (Hansen and Østerhus 2000) and because of cross-frontal mixing with cold water from the Greenland Sea. Part of the warm Atlantic Water from the western branch recirculates along the Greenland Fracture Zone (Quadfasel et al. 1987b) and does not reach the Fram Strait. Between 74° N and 78° N the remaining part of the western branch and the slope branch converge. Their contributions are reflected at the mooring line at 78° 50′ N by the distinct maxima of volume flux density (Fig. 3.4). Due to its position it is mostly the western branch that feeds the immediate recirculation of Atlantic Water while it is mostly the slope branch that crosses the Yermak Plateau to the east.

3.3.2 Flow and Temperature Structure at 78° 50′ N

The mooring data along 78° 50′ N and the hydrographic data clearly show the highly barotropic northward flow of the warm West Spitsbergen Current and the more baroclinic cold East Greenland Current in the western Fram Strait (Fig. 3.5). In the central Fram Strait, the flow is essentially westward, forming one of the recirculation pathways for Atlantic Water. Accordingly, throughout the year the

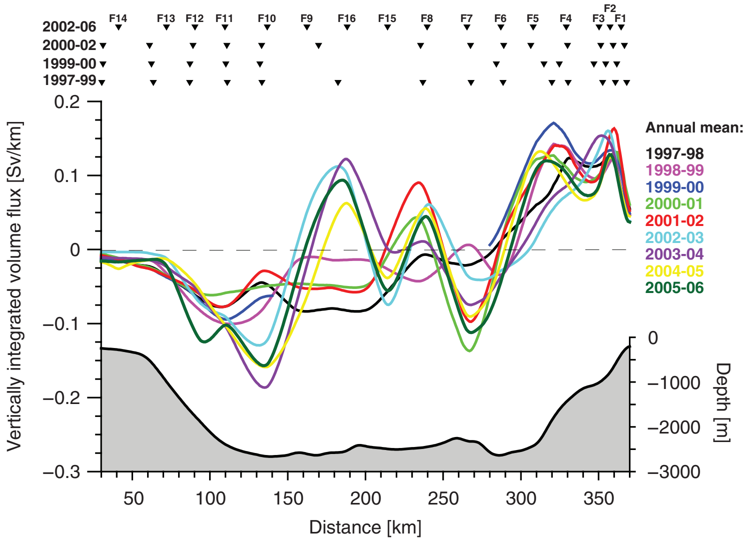

Fig. 3.4 Zonal distribution of annual mean cross-section volume transport per width from 1997 to 2006. The bottom panel shows the bottom depth. Triangles on top mark the mooring positions in the different periods

temperature of the upper layers is highest in the West Spitsbergen Current and decreases towards the west up to the front between the returning West Spitsbergen Current water and the cold Polar Water at about 3° W (Fig. 3.2) – a structure that is known since long from hydrographic summer sections (e.g. Rudels 1987).

The highest velocities were invariably found above the upper slope (water depth <1,500 m) in the West Spitsbergen Current with 9-year mean speed above 20 cm/s. The West Spitsbergen Current also shows the maximum speeds with values above 55 cm/s in the upper 250 m. The East Greenland Current has its core over the base of the continental slope at about 2,500 m where it carries warm modified Atlantic Water southward rather than cold Polar Water (Fig. 3.2). The latter leaves the Arctic Ocean west of this core at somewhat weaker southward velocities.

There is also meridional flow in the central part that is however weaker and more variable than the West Spitsbergen Current and strongly influenced by the complex topography. Immediately west of the West Spitsbergen slope the flow turns southward steered by the southeastern extension of the Molloy Deep. The northward extension of the Knipovich Ridge is likely to be responsible for the northward component at mooring F8. After increasing the lateral resolution in the central Fram Strait in 2002 by adding two moorings also the topographic influence (e.g. that of the Hovgaard Ridge) on the currents further to the west was captured up to the rise of the East Greenland continental slope. While topographic steering is most evident in the near-bottom level (yellow arrows in Fig. 3.5) it also influences the upper-layer meridional component and thus determines the partitioning of Atlantic Water flowing towards the central Arctic Ocean vs the immediate return flow.

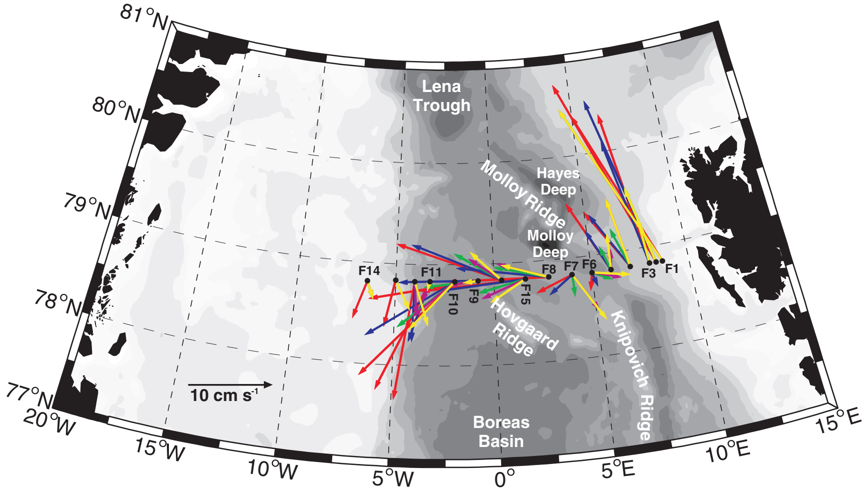

Fig. 3.5 Mean currents obtained from the Fram Strait mooring array. The average was taken over the period 2002–2006 when the mooring positions were kept constant. The color code denotes the nominal depth of the measurement, red: 50 m, blue: 250 m, green: 750 m, magenta: 1,500 m, yellow: near-bottom. With the exception of F15 and F16, the mooring numbering runs from F1 in the east to F14 in the west

Heat can be transported by the mean flow and also by mesoscale features. At irregular intervals southward flow was observed for several weeks at about 8° E (mooring F6) and at the same time northward flow occurred at 7° E (mooring F7), i.e. north of the Knipovich Ridge (not shown, Schauer et al. 2004). South of the mooring line, the current along the Arctic front at times sheds anti-cyclonic baroclinic eddies (Fig. 3.6); these propagate to the north, guided by the topography, and thus explain the intermittent anti-cyclonic features observed in the mooring data.

The strongest flows in the West Spitsbergen Current and in the central Fram Strait occur in winter (not shown, Jónsson et al. 1992; Schauer et al. 2004) which is in accordance with the seasonal spin-up of the cyclonic gyre systems of the Nordic Seas through the wind (Jakobsen et al. 2003). In contrast, the southward volume flow does not show a clear seasonal signal, confirming findings by Jónsson et al. (1992). The upper layer temperatures down to 250 m have a maximum across the entire section in autumn (Fig. 3.7).

3.3.3 Volume Fluxes

The mean net volume flux across the section from 9 years of mooring data is 2 Sv southward, with the standard deviation of 5.9 Sv in the first period with 14 moorings and 2.7 Sv since 2002 when the number of moorings was increased to 16. This residuum is composed of a total of 12 Sv northward flow and 14 Sv southward flow.

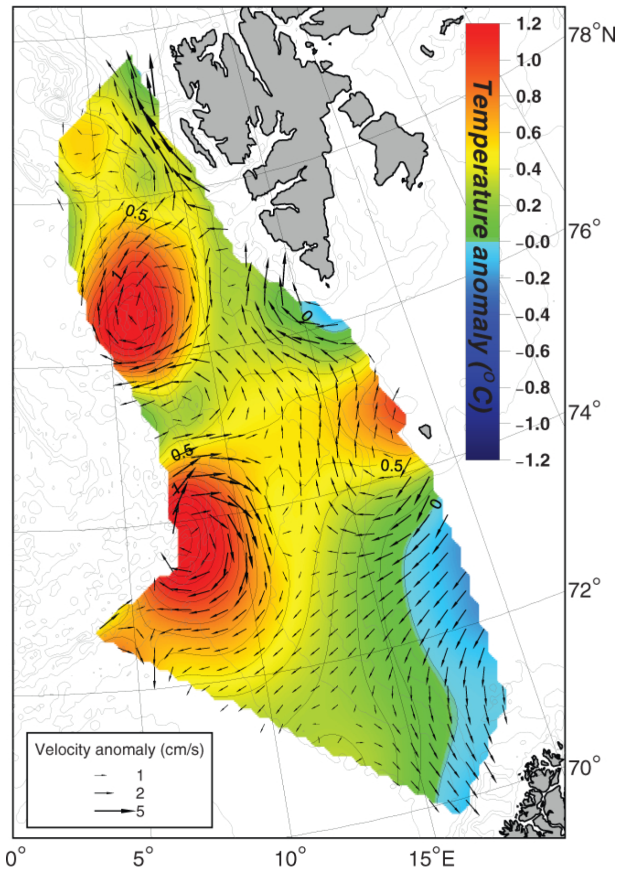

Fig. 3.6 Horizontal distribution of the anomalies of temperature and baroclinic currents in summer 2005 at 100 dbar. The anomalies are with respect to the mean summer values between 2000 and 2006. The baroclinic current is referred to 1,000 dbar or to the bottom

The net southward flow is the compensation for the inflow of Atlantic Water to the Barents Sea opening (Rudels et al. 1994).

While the high velocities of the combined Atlantic Water branches on the West Spitsbergen slope lead to huge volume fluxes in a relatively small area (Fig. 3.4) considerable transports also occur in the current bands in the central part of the strait. Here the mean velocities are low but mostly unidirectional from the surface to the bottom at more than 2,500 m. The weak east–west temperature change in the upper layer (Fig. 3.2) suggests that the banded structure is the projection of meanders of the westward recirculation.

Approximately one third of the northward transport comprises deep water colder than 1 °C that is composed of Greenland and Norwegian Sea Deep Water (Rudels et al. 2008). Part of that water returns within a short loop while the westernmost part of the deep southward flow stems from the interior Arctic Basins.

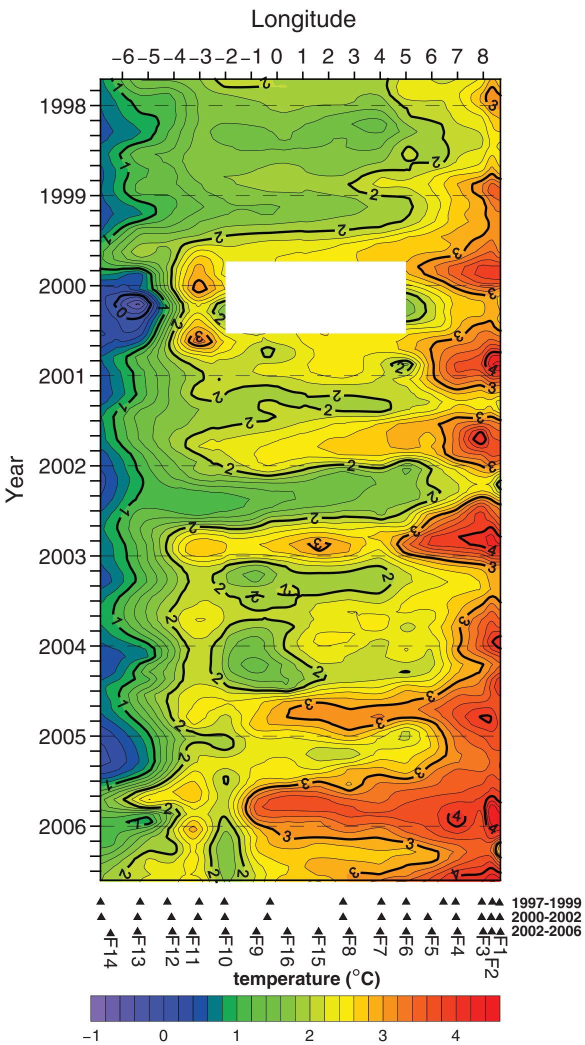

Fig. 3.7 Hovmöller diagram of the monthly mean zonal temperature distribution at 78° 50′ N at 250 m nominal instruments depth from 1997 until 2006

3.3.4 Warming of the Atlantic Water

Both the summer hydrographic data and the year-round mooring data reveal an increase of temperature of the northward flowing Atlantic Water (here water warmer than 1 °C) in the northern Nordic Seas and in the Fram Strait during the decade 1997–2006 (Fig. 3.8). The increase was about 0.5 K between 1998 and 2000 and again about 0.5 K from 2003 to 2006. The significance of this integrated signal is supported by a very coherent course of the time series of individual instruments (not shown). The warming was associated with an increase in salinity and the record maximum values of both properties were observed in summer 2006. With the exception of the first 2 years, the temperature increase was overlaid by a seasonal variation with an amplitude of approximately 0.5 K, but while the summer maxima rose by more than 1 K over 9 years, the winter minima rose much less.

One origin of the warming and the salinity increase are the changes of the sub-polar North-Atlantic with upper ocean temperature and salinity maxima in the Subpolar Gyre and the Faroe–Shetland Channel in 1997/98 and 2003 (Hátún et al. 2005; ICES 2006). On the other hand, changes of the atmospheric cooling of the Atlantic Water during its transfer through the Nordic Seas can mask this signal before it reaches the Fram Strait (Karcher et al. 2008). However, the two temperature maxima occurring both in the Sub-Polar Gyre and in the Fram Strait with a time lag of roughly 2–3 years confirm the fast signal propagation in the boundary current in the Nordic Seas described in (Polyakov et al. 2005).

The hydrographic summer observations at 76° 30′ N (taken here between 2000 and 2006) reveal that warmer water was advected in both the slope current and the frontal current (Fig. 3.9). The average increase of summer temperature at 200 m between 2003 and 2006 was more than 1 K over large parts of the section. This is more than twice as much as the increase in the yearly running mean temperature obtained from the mooring data at 78° 50′ N and also much larger than the increase of the maximum summer temperatures between 2003 and 2006 there. This underlines the difficulty to derive interannual variability from snapshots at a single depth in a region with high seasonality. However, despite being masked by mesoscale features (Fig. 3.6), there is some indication that in 2004 and 2005 the western branch was more warming than the eastern one.

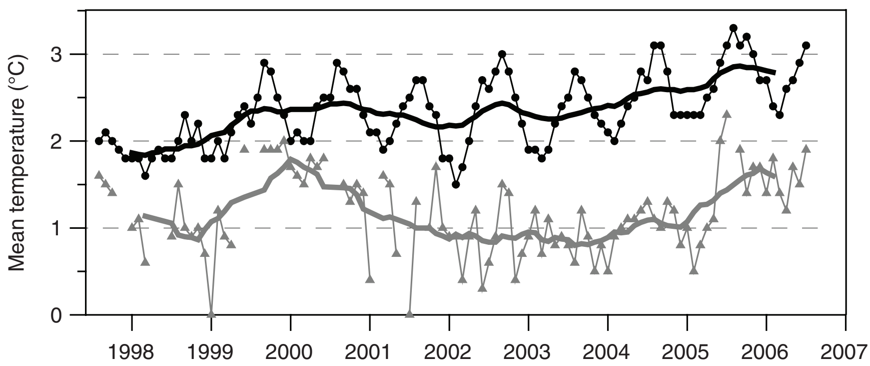

Fig. 3.8 Time series of the cross-section averaged temperature of Atlantic Water derived from mooring data. The black line denotes the temperature of northward flowing water that was warmer than 1 °C. The grey line denotes the temperature of southward flowing Atlantic Water (for explanation see text). Symbols at the thin lines denote monthly mean values, bold lines are 12-month running means

3.4 Heat Transport Through the Fram Strait

3.4.1 *Conceptual Remarks About Estimating Oceanic Heat Transport into the Arctic Ocean*

Since the volume flux through the Fram Strait, just like the flux through all other passages to the Arctic Ocean, is not balanced the heat flux can not be calculated straightforward. The complexity of the flow through the Fram Strait adds to the difficulties finding a reasonable scheme for computing the heat flux. The principle for the calculation of advective heat transport is described in the oceanographic literature since more than 30 years (Montgomery 1974). Nevertheless, the last decade shows a wealth of publications from which a misconception of this principle is evident (among many others: Schauer et al. 2004; Maslowski et al. 2004; Karcher et al. 2003; Lee et al. 2004). This makes it worthwhile bringing to mind the basic concepts once more.

The physical idea behind oceanic advective heat transport is related to temperature flux convergence. Practically this may be referred to a defined ocean volume (or mass) holding a certain amount of heat. Currents across the boundary of that ocean segment can change the heat content by replacing a certain amount of water of a particular temperature by the same amount of water with (usually) another temperature. The difference of the heat content of the replaced volumes is the heat gain or loss of the considered ocean segment. Such an exchange can be achieved by ocean currents of any scale, by basin-wide gyres or overturning cells as well as by small eddies.

At stationary conditions the heat gain/loss through currents has to be balanced by sinks/sources, S, like, e.g. heat exchange with the atmosphere. This heat balance of the ocean segment is resumed in the equation

$$S = \oint ds \int_0^H c_p \cdot \rho \cdot v_\perp \cdot T dz \tag{3.1}$$

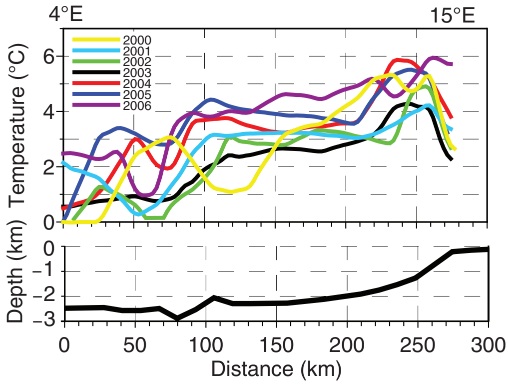

Fig. 3.9 Temperature at 200 dbar along 76° 30′ N between 4° E and 15° E from the summers 2000 to 2006 (Walczowski and Piechura 2007, Fig. 3.5a). The bottom panel shows the bottom depth

with c_p specific heat, ρ density, v_{v1} the velocity component perpendicular to the open ocean boundary confining the segment and T the temperature of the flow. The integral is taken over the full depth, z, from top to the bottom, H, around the entire ocean boundary of which ds is a boundary length element. This concept holds as well for variable conditions in which case also a change of the heat content of the ocean segment, H, with time, t, is possible.

$$\frac{\partial H}{\partial t} + S(t) = \oint ds \int_0^H c_p \cdot \rho \cdot v_{\perp}(t) T(t) dz \tag{3.2}$$

This concept sounds (and probably is) trivial. It implies that heat transports can be calculated in a system with mass conserved only (Montgomery 1974; Hall and Bryden 1982). However, heat transport computations by evaluating observations and even model results are sometimes far from straightforward. This is partly due to the complexity of ocean currents that often does not allow to determine velocity and temperature along the complete boundary at a high enough resolution. A second problem often arises from the formulation of the advective heat flux term itself. It is extremely tempting to disintegrate the integral over a closed boundary in Equations (3.1) and (3.2) and to calculate "temperature fluxes" over partial cross-sections (Lee et al. 2004). This holds as long as these temperature fluxes are regarded as interim terms required to compute the entire integral. However, it is sometimes argued that temperature fluxes can also be used themselves, e.g. for comparing different cross-section parts (Karcher et al. 2003) or to rate temporal changes through a particular partial cross-section (Schauer et al. 2004). It has also been suggested that certain reference temperatures such as the volume average temperature (Lee et al. 2004) are well suited to derive heat transports from temperature fluxes. However, these as well as any other temperature fluxes are entirely arbitrary and attempts to use them instead of heat fluxes produce wrong results (Schauer and Beszczynska-Möller, in preparation).

Meridional heat transport computed from hydrographic data, e.g. in the North Atlantic south of Greenland has large error bars (Ganachaud and Wunsch 2000) but is reasonable since the Atlantic north of any coast-to-coast zonal section is closed apart from a small influx from the Pacific. This inflow (about 0.8 Sv) (Woodgate et al. 2006) might be neglected in comparison to the meridional flow of O(10–10^2 Sv) through the North Atlantic, and the Bering Strait inflow temperature is similar to that of the deep North Atlantic flow.

With respect to Arctic–Subarctic Ocean fluxes, however, determination of oceanic heat transport principally needs to take into account all openings, Bering Sea, Canadian Archipelago, Fram Strait and Barents Sea Opening, in order to accomplish the requirements of Equations (3.1) and (3.2). Without any further constraints arising from the Arctic Ocean internal circulation heat transport through single straits can not be computed because none of the straits confining the Arctic Ocean has a balanced volume (mass) flux. Consequently one has to define carefully what is meant by "heat transport through the Fram Strait" in order not to deal with an ill-defined term.

The problem does not vanish when "only" temporal changes are compared (Montgomery 1974). Heat transport to the Arctic Ocean can change because of varying temperature difference between inflow and outflow and because of varying flow strengths. Here as well, isolated consideration of the changing properties of individual (in)flow branches leads to arbitrary results (Schauer and Beszczynska-Möller, in preparation).

The only way to elude the necessity of addressing all Arctic Ocean openings simultaneously for heat transport computations evolves if we can use constraints provided through the Arctic Ocean internal circulation. For example, for the inflow of warm Pacific Water through the shallow Bering Strait it has been shown that practically all of this water is cooled to freezing temperature before it exits the Arctic Ocean so that the heat flux can be derived from the inflow only (Woodgate et al. 2006). This is certainly not true for the Atlantic inflow through the Fram Strait. Therefore, only if we can identify compensating in- and outflow branches, i.e. if we can regard them as a stream tube, we can derive the heat flux provided through this pair.

3.4.2 An Approach to Compute the Heat Transported by the West Spitsbergen Current to the Arctic Ocean

With regard to the water carried northward in the West Spitsbergen Current we probably can safely assume that the bulk of this water also leaves the Arctic Ocean through the Fram Strait. Water from the West Spitsbergen Current propagating along the shelf edge into the Nansen Basin might flow on the shelf east of Spitsbergen and return to the northern and then western Barents Sea. This probably is only a small fraction of the water within the upper 150 m since much of the water entering the shelf through a canyon returns in a cyclonic loop to the shelf edge (Gawarkiewicz and Plueddemann 1995). A small fraction might however circulate anti-cyclonically around Svalbard. The flow through the 50 m deep Bering Strait is of the order 1 Sv to the north and there are no reports about Fram Strait water travelling southward to the Pacific (Woodgate et al. 2006). The Canadian Archipelago (sill depth 160 m) is the main gateway for the exit of Pacific Water (Steele et al. 2004) and for a fraction of Barents Sea water (Rudels et al. 2004). Any fraction from the Fram Strait is probably small.

The travel times along the various pathways of West Spitsbergen Current water in the Arctic Ocean, around all basins or only in the northern Fram Strait, last between months and decades. Warm water anomalies that have entered the Arctic Ocean with the West Spitsbergen Current in the nineties have reached the eastern Eurasian Basin 4 years later (Karcher et al. 2003; Polyakov et al. 2005) and we do not know yet which part of the associated additional heat is released to the surface and which part will leave the Arctic Ocean after several years or decades. However, assuming that their remnants finally end up in the Fram Strait we can consider the loops as closed volumes.

This should enable us to use the observations of velocity and temperature in the Fram Strait and compute the heat flux provided to the Arctic by the West Spitsbergen Current by adding the temperature fluxes of northward and southward flow.

Time series of temperature flux can be constructed from the interpolated fields of temperature and cross-section component of the velocity (Schauer et al. 2004). Since the southward volume flow is larger than the northward flow the critical point is how to identify which of the southward flow is returning West Spitsbergen Current water and which water stems from other openings like the western Barents Sea or the Bering Strait.

We assume that owing to continuity, water from any loop of the returning West Spitsbergen Current will flow southward immediately west of the West Spitsbergen Current. There is no indication that the Barents Sea branch crosses any of the West Spitsbergen Current-derived loops. Rudels et al. (1994) and Schauer et al. (2002a) showed that the Barents Sea Water displaces the Fram Strait branch off the slope at the confluence of the two branches in the northern Kara Sea and that further downstream the Fram Strait branch flows at the basin side of the two. If this pattern continues along the entire Arctic Ocean rim, all West Spitsbergen Current-derived southward flow in the Fram Strait would take place immediately west of the northward flow and the Barents Sea water would flow west of that.

While we assume based on continuity reasons (no crossing flow branches) that return flow in the central part east of the westernmost northward branch originates from the West Spitsbergen Current we have to distinguish for the East Greenland Current which part is constituted from West Spitsbergen Current water and which part from other sources. We assume that the warmest water stems from the West Spitsbergen Current.

To avoid volume flux uncertainties that arise from the still poorly resolved deep-water fluxes we limit our computations to the northward flow of upper and intermediate waters and we use a limiting temperature, $T_{DI} = 1\,°C$, for distinction between the two. With the exception of the front around its outcrop the depth of the 1 °C isotherm is below 500 m for northward flow in the West Spitsbergen Current (Fig. 3.2). The argument behind this choice is that water below that depth is very unlikely to reach the surface in the central Arctic Ocean and therefore must return at the same temperature through the deep Fram Strait (of course it can be mixed with other deep water, e.g. generated in the Barents Sea, which would be at similar temperatures). However, it has thus no chance to contribute to the surface heat flux.

The flux of upper layer water warmer than 1 °C is integrated over the entire cross section. The net volume flux can be positive, zero or negative. With zero net volume flux the heat flux of West Spitsbergen Current to the Arctic Ocean is immediately obtained by temperature flux integration over the respective cross section. In the case that the volume flux of water warmer than 1 °C was net northward, obviously West Spitsbergen Current water has been cooled to temperatures below 1 °C before returning. In this case we increased the integration area over water flowing southward to include also colder water. The distinction temperature for returning West Spitsbergen Current water, T_{DO}, was incrementally decreased until the resulting net flux was zero (within ±1 Sv).

A net southward volume flux would mean that there is water warmer than 1 °C flowing southward that does not originate from the West Spitsbergen Current. This is very unlikely: Water that entered through the Bering Strait is cooled to near freezing if it reaches the Fram Strait at all after crossing the entire Arctic. Barents Sea water looses much of its heat in the Barents and Kara seas so that it is densified and sinks to intermediate depths when entering the Eurasian Basin. According to observations taken between the 1960s and mid-1990s (Schauer et al. 2002b) all Atlantic Water that leaves the northern Kara Sea is colder than 1 °C. This might have changed in years thereafter. However, if the water would enter the central Arctic warmer than at 0 °C it would be lighter and closer to the surface. In this case it is exposed to Arctic surface influences more than water from the West Spitsbergen Current because it travels along the shelf edge and is more likely to upwell than the latter is. Furthermore it has the longest pathway. Therefore, in the case of net southward volume flux of water warmer than 1 °C we have to assume that it is caused by a large error of our velocity interpolation and that we can not determine a heat flux in that period from our data.

3.5 Resulting Heat Transports

The result of this approach for computing heat flux to the Arctic by West Spitsbergen Current water is given in Fig. 3.10. The maximum error limits associated with the interpolation between data points are considered to be of equal size as those in (Schauer et al. 2004), ±6 TW, since despite the wrong concept used there the uncertainties arising from the limited spatial resolution remain the same.

For T_{DI} = 1 °C, the distinction temperature for the outflow required to obtain zero net volume flux, T_{DO}, varied between −0.7 °C and 0.7 °C except of 1 month when it was −1.6 °C. Similar as the flux averaged temperature that increased from about 2 °C to almost 3 °C (Fig. 3.8) the annual mean volume flux of the Atlantic Water was rising in the last decade from less than 5 Sv to more than 7 Sv in 2004 and 2005 (Fig. 3.10). Due to the way the Atlantic Water is defined here, the volume flux increase is mostly a consequence of the warming. The temperature increased over the upper 800 m and thus the 1 °C isotherm in the West Spitsbergen Current was found 200 m deeper in 2004 than in 1997.

The annually averaged heat transport increased in the first 2 years from 26 to 36 TW which impressively demonstrates the influence of a wrong method as it was used by Schauer et al. (2004) where the increase was stated to be from 16 to 41 TW. After a dip in 2001, the heat flux increased to its decadal maximum of 50 TW in 2004. While the temperatures of the West Spitsbergen Current water continued to rise to a record high in 2006 the associated heat flux decreased again to 40 TW because much warmer water returned in that year to the Greenland Sea than before (Fig. 3.8). The reason for this can be twofold: Warmer water could finally return from one of the longer loops through the central Arctic Ocean that had entered in previous warming periods like in the early 1990s (Quadfasel et al. 1991). The second possibility is that the anomalously warm Atlantic Water advected in 2005 and 2006 recirculates immediately in the Fram Strait which is suggested from the extraordinarily high temperatures in the central Fram Strait (Fig. 3.7). Then the question must be posed what drives the strengthening of the recirculation vs north- and/or eastward

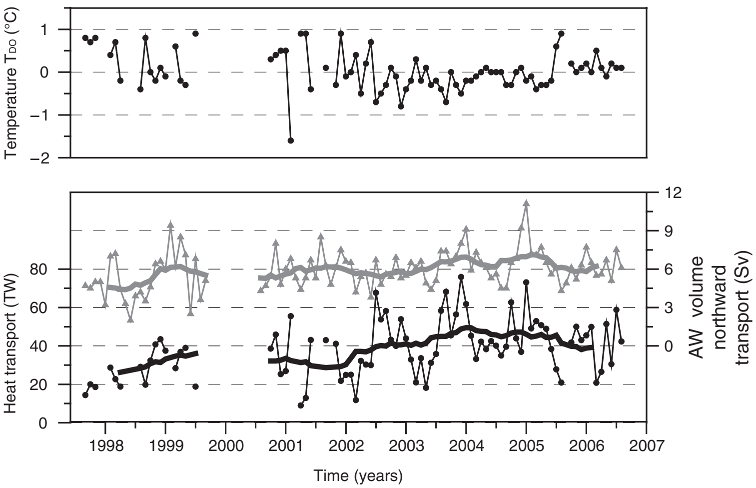

Fig. 3.10 Time series of the volume flux (grey lines) and heat flux (black lines) to the Arctic through Atlantic Water (warmer than 1 °C) in the West Spitsbergen Current. The upper panel gives the outflow distinction temperatures T_{DO} (see text for explanation). Symbols at the thin lines denote monthly mean values, bold lines are 12-month running means. Note that the southward volume flow of Atlantic Water is the same as the northward flow within ±1 Sv. For the uncertainties of the heat flux see discussion in Section 3.6

flow – whether it is a consequence of the change in large-scale atmospheric pattern that returned to a less cyclonic state in recent years or if this is due to a decrease in the pressure gradient across the Fram Strait due to the rising steric height in the Nordic Seas as a consequence of the warming (Jakobsen et al. 2003).

3.6 Critical Discussion of the Limits of the Approach

Besides the volume and heat flux errors inherent in the spatial interpolation, the proposed approach implies several uncertainties.

The choice of the distinction temperature for northward flow, T_{DI}, is somewhat arbitrary. Ideally, T_{DI} should be chosen in a way that the resulting heat flux is not sensitive to small changes. If T_{DI} is too high parts of the West Spitsbergen Current are excluded and the heat flux is underestimated (Schauer and Beszczynska-Möller, in preparation). If T_{DI} is too low many situations arise with non-zero net flow which demonstrate problems with the spatial resolution of the flow. These problems are larger in the first half of the observation period when the mooring number and instrumentation coverage was lower than in the second half.

The most critical point is, however, the disregard of mixing. Diffusion between the Fram Strait and Barents Sea branches during the passage through the Eurasian Basins takes place as double-diffusive layering (Rudels et al. 1999) as well as through mesoscale eddies (Schauer et al. 2002a, b). Also vertical displacement of warm Atlantic Water by entrainment into sinking dense shelf water plumes is a mechanism not explicitly taken into account by the stream tube approach.

Both processes imply that the values as given in Fig. 3.10 are overestimating the heat flux. Entrainment into sinking plumes means that warm West Spitsbergen Current water is returning to the Nordic Seas as deep water which is not considered here. For continuity, the drainage must be replaced by cold deep water upwelling in the central Arctic Ocean. Mixing with Barents Sea water obviously also means that some warm water of the West Spitsbergen Current returns to the Nordic Seas outside of the stream tube.

A gross estimate of the loss of West Spitsbergen Current water (mean temperature 2.5 °C, mean volume transport 6 Sv, Figs. 3.8 and 3.10) to deep waters which have an average temperature of −0.6 °C and −0.7 °C for the northward and southward flow, respectively, yields 0.2 Sv. Assuming thus a contribution of 0.2 Sv of compensating −0.5 °C cold central Arctic deep water included in the return water corresponds to 5% overestimation of the heat flux, i.e. about 2 TW which is within the interpolation induced error limits.

A similar assessment for mixing with the Barents Sea water can hardly be made. According to (Schauer et al. 2002b), 50% of the approximately 2 Sv Barents Sea Water leaving the northern Kara Sea is colder than 0 °C and 50% is warmer. While the cold fraction sinks at the Nansen basin slope deeper than 500 m, the warmer fraction remains in the same depth level as the West Spitsbergen Current water. Assuming the average temperature of the warmer fraction to be 0.5 °C, admixture of this fraction to the West Spitsbergen Current water would explain 10% of the estimated heat flux. If this Barents Sea Water fraction is however cooled to, e.g. −0.5 °C before it is mixed it would effect an overestimation of the heat flux by about one third. Mixture of all Barents Sea water (2 Sv) at −0.5 °C to the West Spitsbergen Current water would imply further reduction by one third and would involve that the heat flux of West Spitsbergen Current water to the Arctic is approximately 10 TW.

These examples show that, for principle reasons, in case of strong mixing the significance of the heat flux variability can hardly be addressed with this approach as long as the variability of the Barents Sea properties at their entrance to and during their passage through the Arctic Ocean are unknown. Would they be known, the stream tube concept for the West Spitsbergen Current could be extended to include the Barents Sea throughflow. Calculation of the heat transports with constant Barents Sea outflow temperature and fluxes in the St. Anna Trough would, however, *a priori* decide upon the variability for which we are searching. In any case, neglecting mixing with Barents Sea Water leads to an overestimation of the heat flux to the Arctic.

3.7 Some Consequences for Observational Strategies

The above considerations point to difficulties inherent to the assessment of the oceanic heat delivered through advection to the Arctic Ocean. They also lead to considerable consequences for observational strategies. First, to compute heat transport variability simultaneous observations are needed at least across those openings that are connected by currents. Second, these observations definitely need to be made at high spatial resolution of the velocity and temperature structure across these openings. In the Fram Strait the lateral variability that is of the scale of tens of kilometres due to the small internal Rossby radius and the complex topography translates directly into the need of a high number of moorings since this is so far the only way for time series of appropriate horizontal resolution. From measurements that spatially integrate properties like temperature or velocity no heat transports can be derived. Furthermore, in order to assess what fraction of the heat is released in the Arctic Ocean vs what fraction is simply passing by time series have to be long enough to cover the maximum travel time of a parcel which in the case of parcels travelling along the entire Arctic continental slope are decades.

Last but not least it should be mentioned that the same considerations, closed volumes or stream tubes, high resolution and long time series, hold also for the assessment of "fresh water fluxes".

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