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Problems with estimating oceanic heat transport – conceptual remarks for the case of Fram Strait in the Arctic Ocean

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Abstract

While the concept of oceanic heat transport – or rather heat transport divergence – is known since long, it is sometimes applied inaccurately. Often temperature transports are computed across sections with unbalanced volume flow which then depend entirely on the choice of the temperature scale. The consequences of such arbitrariness are demonstrated with a simple calculation exercise for the passages to the Arctic Ocean. To circumvent the arising difficulties for the Fram Strait as an example we propose a stream tube concept to define a net zero volume flow section which can, with coarse assumptions, be used to determine oceanic heat transport by the Atlantic water flow. Weaknesses of this approach and consequences for observational strategies are discussed.

1 Introduction

Oceanic heat transport is a crucial component in the global climate system because it is partly responsible for compensating the global meridional radiation imbalance (e.g. Hall and Bryden, 1982). Atmosphere and oceans are heated by an insolation surplus at low latitudes and large-scale circulation systems carry warm air, moisture and water poleward. In both subsystems the respective return flow to the equator is colder and eventually the excess heat is released to space by long-wave radiation. Extensive exchange between ocean and atmosphere through various processes underlying temporal and spatial variability complicate the heat budgets of each subsystem. In order to evaluate the earth climate the individual contributions to the terms of the heat balance have to be addressed.

In northern high latitudes, oceanic heat transport is small compared as compared to the mid latitudes. Yet in recent years, the role of the ocean is debated in the context of the sea ice reduction. In order to quantify the oceanic impact, also attempts are made to derive the advective oceanic heat import from sub-arctic oceans.

The principle for the calculation of advective heat transport is described in the oceanographic literature since more than thirty years (Montgomery, 1974; Hall and Bryden, 1982). 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; Cuny et al., 2005; Lee et al., 2004). This makes it worthwhile to bring to mind the basic concepts once more and to illustrate the consequences of their violation. Consequently, heat transport through passages with non-zero volume transport can be estimated, if at all, under rigid assumptions only. A respective approach to estimate heat transport of Atlantic Water flow through Fram Strait from observations is presented.

2 Basic concept of oceanic heat transport

"Heat carried by 1 ton of water at 30°C cannot be compared with the heat carried by 30 tons of water at 1°C." – (Montgomery, 1974).

The physical idea behind oceanic advective heat transport is related to the temperature transport divergence. It can be derived from the heat (more precisely: internal energy) conservation equation, which for an ocean water parcel, under the simplification of incompressibility, can be written as

$$\frac{\partial T}{\partial t} + \boldsymbol{v} \cdot \nabla T = -\frac{1}{\rho c_p} \nabla \cdot \boldsymbol{f}, \tag{1}$$

where T is the potential temperature referred to atmospheric pressure, c_p is the specific heat at atmospheric pressure and ρ is the density. $\boldsymbol{v}$ is the three-dimensional velocity vector, and $\boldsymbol{f}$ compiles turbulent and conductive heat flux densities. Since Eq. (1) deals with temperature derivatives only, T can be expressed either in Celsius scale or in absolute scale or it can be referred to any constant reference temperature T_{ref}. Here, c_p and ρ are considered constant, while $\boldsymbol{v}$, $\boldsymbol{f}$ and T may vary with time, t, and

space. Eq. (1) may be integrated over a full depth ocean volume, Vol. Since $\boldsymbol{v}$ can be considered to have zero divergence, Gauss' theorem can be applied so that

$$\oint ds \int_0^{d(s)} v_\perp(s,z,t) \cdot (T(s,z,t) - T_{\text{ref}}) dz = -\iiint_{\text{Vol}} dx\, dy\, dz \frac{\partial (T - T_{\text{ref}})}{\partial t} - \frac{F_{\text{ocsurf}}(t)}{c_p \rho}. \tag{2}$$

$v_\perp$ is the velocity component perpendicular to the open ocean boundary confining the volume. The integral on the left-hand side is taken over the full depth, d, along the entire ocean boundary of which ds is a length element. F_{ocsurf} is the heat flux through the entire interface between ocean and atmosphere; turbulent and diffusive fluxes across the lateral boundaries are considered negligible here.

Equation (2) expresses the simple fact that currents across the boundary of that ocean segment change the heat content by replacing a certain amount of water of a particular temperature by the same amount of water with another temperature. 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 or loss through currents will be balanced by the heat exchange with the atmosphere, F_{ocsurf}. At variable conditions, in addition a change of the heat content of the ocean segment with time is possible.

This concept sounds (and probably is) trivial. Zero velocity divergence implies that heat (or rather enthalpy, Warren, 1999) transports can only be calculated in a system with net zero volume transport across its boundaries (Montgomery, 1974; Hall and Bryden, 1982), i.e. if

$$\oint ds \int_0^{d(s)} v_\perp(y,z,t) \cdot dz = 0. \tag{3}$$

Multiplying Eq. (3) by the reference temperature T_{ref} and adding it to Eq. (2) shows that in this case the temperature scale or the choice of T_{ref} has no influence on the result.

However, computation of heat transport by evaluating observations or even model results are sometimes far from straightforward. This is partly due to the complexity of ocean currents and the resulting necessity to determine both velocity and temperature along the boundary at a sufficiently high spatial resolution.

A second problem often arises from the formulation of the advective heat transport term itself. It is extremely tempting to integrate only over a partial cross-section instead over the closed boundary. This is fine as long as these terms are regarded as steps for computing the entire integral. However, it is sometimes argued that such "temperature transports" 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 is also suggested that certain reference temperatures such as the volume average temperature (Lee et al., 2004; Kim et al., 2004) are well suited to derive heat transport from a "temperature transport". However, an arbitrary reference temperature makes a "temperature transport" across a partial section also arbitrary and provides wrong results for comparisons as is illustrated in the next chapter.

3 Problems arising from misconception of heat transport

The arbitrariness and thus irrelevance of "temperature transports" across partial sections, i.e. sections across which the net volume transport is *not* zero, is obvious from a simple thought model: Let us imagine a basin with a gyre circulation, homogeneous temperature and no heat exchange with the atmosphere. By integrating $c_p \rho v_\perp T$ along any partial section, a "temperature transport" may be computed which obviously depends on the temperature scale (or reference temperature). Doubling the velocity of the flow would double the so-defined "temperature transport" as well. However, despite an increased "temperature transport" (or even "heat transport" if this term is used synonymously) the temperature and thus heat distribution of this ocean remain unchanged. In fact no heat transport takes place at all which is also evident from integration of the

transport across any net-zero-volume-transport section for any flow strength and for any reference temperature.

In a more complex situation the faultiness of "temperature transports" across partial sections is less obvious. The Arctic Ocean, e.g., has four passages to sub-arctic oceans: the Bering Strait, the Canadian Archipelago, the Fram Strait and the Barents Sea Opening (Fig. 1) and none of these straits accommodates a balanced volume flow. Quantification of oceanic heat transport through these straits is therefore a daunting effort. Many published approaches ignore the involved restraints and thus produce wrong estimates (Schauer et al., 2004; Walczowski et al., 2005; Karcher et al., 2003).

A simple exercise may illustrate the arbitrariness of such attempts. Table 1 shows results for the heat balance of a very simplified Arctic Ocean to and from which ocean currents transport water with temperatures given in different temperature scales, Celsius and absolute. The total heat gain of the Arctic Ocean is independent of the temperature scale because volume transports are balanced. The "temperature transports" of individual flow branches, however, differ considerably between the absolute and the Celsius scale and even change their sign. Indeed, with both scales the "temperature transport" associated with the Fram Strait inflow is larger than that with the Barents Sea inflow, however, the "temperature transport" *ratio* between the two branches depends on the reference temperature. While with the Celsius scale the Fram Strait inflow temperature transport is twice as large as that of the Barents Sea inflow it is more than three times as large with the absolute scale. Consequently, arbitrariness of temperature scales leads to arbitrary results when comparing different heat transport "contributions".

The effects are similar when comparing temporal changes (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. Again, consideration of the changing properties of individual flow branches lead to arbitrary results. In the example given in Table 2 the inflow through Fram Strait is increased by 1 Sv (1 Sv=10^6 m^3/s) with respect to the example of Table 1. To keep the volume of the Arctic Ocean constant we assume that the Fram Strait outflow compensates the change and increases as well. The according "temperature transport" increase in the Fram Strait inflow gives additional 12 TW with the Celsius scale and additional 1104 TW with the absolute scale. The true additional heat transport for the Arctic Ocean is, however, 16 TW, independent of the temperature scale. Again, isolated determination also of variability of a "temperature transport" through a partial cross section provides no useful information.

4 A stream tube concept for the West Spitsbergen Current

The above examples underline that principally all Arctic Ocean openings have to be addressed simultaneously to accomplish the requirements of Eqs. (1) and (2) and thus to obtain oceanic heat transport variability. Since the volume flux through, e.g., Fram Strait is not balanced "heat transport through Fram Strait" is an ill-defined term.

The only way to elude this difficulty is if we can use constraints provided through the Arctic Ocean internal circulation. For example, the warm Pacific Water entering through the shallow Bering Strait has been shown to practically cool to freezing temperature before it exits the Arctic Ocean so that the heat transport can be derived from the inflow referred to freezing temperature (Woodgate et al., 2006). This is certainly not true for the Atlantic inflow through Fram Strait. Therefore, only if we can identify compensating in- and outflow sections, i.e. if we can regard the flow in a stream tube, we can derive the heat transport provided through this pair of partial sections.

To define a stream tube, we assume here that all of the Atlantic Water that is carried by the West Spitsbergen Current (WSC) through Fram Strait to the Arctic Ocean also leaves the Arctic Ocean through Fram Strait[1]. In the Arctic Ocean, part of the Fram

[1] Water from the WSC propagating anticyclonically along the shelf edge into the Nansen Basin might flow on the shelf and return to the northern and then western Barents Sea. This is probably 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. A small fraction

Strait Atlantic Water follows the shelf edge and travels around all basins, part of it returns along the mid ocean ridges and some of it returns after only a short loop in the northern Fram Strait (Fig. 2). The travel within the various loops lasts between months and decades. Warm water anomalies that have entered the Arctic Ocean with the WSC in the nineties are reported to spread in the Eurasian Basin (Karcher et al., 2003; Polyakov et al., 2005) and it is unclear which part of the associated additional heat is released to the surface and which part leaves the Arctic Ocean after several months, years or decades. However, assuming that the water finally returns to Fram Strait we can consider the looping tubes as a closed volume.

This should enable us to use observations of velocity and temperature that were obtained with moored instruments in Fram Strait and compute the heat transport provided by the WSC by adding the temperature transports of in- and outflow (Schauer et al., 2008). The mooring array is maintained since 1997 up to now and covers the entire cross section of Fram Strait between the shelf edges (Fig. 3). Velocity and temperature values are measured once per hour, data processing and averaging details are given in Fahrbach et al. (2001). Time series of temperature transport 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 WSC water and which water stems from other openings like the Barents Sea Opening or the Bering Strait.

We assume that owing to continuity, water from any loop of the returning the WSC will flow southward immediately west of the WSC. There is no indication that the Bar-

might however circulate anticyclonically 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. 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). The net flow is between 2 and 3 Sv (Cuny et al., 2005). Any fraction from Fram Strait is probably small.

ents Sea branch that enters the Central Arctic Ocean through the St. Anna Trough crosses any of the WSC-derived loops. Rudels et al. (1994) and Schauer et al. (2002) 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 (Fig. 2). If this pattern continues along the entire Artic Ocean rim and ridges all West Spitsbergen Current-derived outflow through Fram Strait would take place immediately west of the inflow and the Barents Sea water would flow to the west of that.

While we suggest that southward flow in the central part east of the westernmost northward branch originates from the WSC we have to find out to what extent the East Greenland Current is constituted from WSC water and from other sources. We assume that the warmest water stems from the WSC.

To avoid volume flux uncertainties that arise from poorly resolved deep-water fluxes we limit our computations to the upper and intermediate waters and we use a limiting temperature T_{DI} for distinction between intermediate and deep water. The choice of T_{DI} bears some arbitrariness which is discussed below. Here we will consider results from using T_{DI}=1°C.

With the exception of the front around the 1°C-isotherm outcrop in the western Fram Strait the depth of 1°C is below 500 m. This holds particularly for all northward flowing water (Fig. 3). It is very unlikely that water entering below that depth reaches the surface in the Central Arctic Ocean and consequently it must return at the same temperature through the deep Fram Strait (of course it can be mixed with other deep water, e.g. from the Barents Sea Opening, which would be at similar temperatures). However, it has thus no chance to contribute to the surface heat flux.

The transport of upper layer water warmer than 1°C (Fig. 3) is integrated over the entire cross section. The net volume flux can be positive, zero or negative. With zero net volume flux we obtain immediately the heat flux of WSC to the Arctic Ocean by integrating the according temperature fluxes over the respective section. In the case that the net volume flux of water warmer than 1°C was northward, obviously some WSC

water has been cooled even to temperatures below 1°C before returning. In this case we have increased the integration area over southward flowing water by incrementally increasing the limiting temperature, T_{DO}, including water colder than 1°C until the net flux is zero. With increments of 0.1 K we reach deviations from zero volume fluxes of less than ±0.4 Sv.

A net southward volume flux would mean that there is water warmer than 1°C flowing southward that does not originate from the WSC. This is very unlikely according to oceanographic knowledge. Water that entered through the very shallow Bering Strait will have been cooled to near freezing before it ever reaches Fram Strait after crossing the entire Arctic. Barents Sea water looses much of its heat on the shelf so that it is densified and sinks to intermediate depths when entering the Eurasian Basin in the northern Kara Sea. According to observations taken in the early and mid-1990s (Schauer et al., 2002b) all Atlantic water that leaves the eastern Barents Sea is colder than 0°C. This might have been different in years thereafter. However, would Barents Sea water enter the Central Arctic warmer than at 0°C it would be lighter than Fram Strait water and therefore closer to the surface. In this case it is exposed to Arctic surface influences more than WSC water because it travels along the shelf edge and is more likely to upwell than Fram Strait waters are. Furthermore it has the longest pathway. Therefore, we assume cases of net southward volume flux of water warmer than 1°C in Fram Strait to be caused by a respective error in the velocity field due to insufficient spatial resolution that we can not estimate a heat flux in that period.

The result for this approach of computing heat flux to the Arctic through WSC water is given in Fig. 4. For T_{DI}=1°C, the distinction temperature for the outflow, T_{DO}, chosen to obtain monthly net volume flux, varies between −0.7 and 0.7°C except of one month with −1.6°C. The increase of the annual running mean in the first two years is from 26 TW to 36 TW; note that using the same data but a wrong method, Schauer et al. (2004) stated an increase more than twice as high, from 16 to 41 TW. The annual mean heat flux increased from 26 TW in 1998 to 50 TW in 2004. While the temperatures of the WSC water continued to rise to a record high in 2006 (Fig. 5) the associated heat flux decreased again to less than 35 TW (Fig. 4) because very warm water returned in that same year to the Greenland Sea, either because the immediate return flow in the Fram Strait region was strong or because water from earlier warm pulses that had entered the Eurasian Basin now reached the Fram Strait.

5 Discussion of inherent uncertainties

Some aspects of this approach are debatable. The choice of the limitation temperature for inflow, T_{DI}, is arbitrary. If T_{DI} is too high, parts of the WSC are excluded, and the heat flux is underestimated as is obvious from comparing the curves for T_{DI}=1°C and for T_{DI}=1.5°C in Fig. 4. Ideally T_{DI} should be chosen low enough that further decrease does not alter the results significantly. However, lowering T_{DI} below 1°C produces more and more situations where zero net flow can not be reached (e.g. T_{DI}=0.5°C in Fig. 4) which reflect 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 instrumental coverage was lower than in the second half. Therefore we chose T_{DI}=1°C as a compromise. The record average difference between the monthly heat transports for T_{DI}=1°C and for T_{DI}=1.5°C was 3.5 TW while it was only 2.5 TW between T_{DI}=0.5°C and T_{DI}=1.0°C, showing shrinking, although not neglegible sensitivity.

Obviously the spatial resolution of the measurements remains insufficient – despite the instrumental coverage of this mooring array (16 moorings with 70 instruments) being probably one of the highest in existing ocean observatories. The problem becomes also evident from the variation of the net volume fluxes (Fig. 6). The month-to-month changes of the net volume flow through Fram Strait often reach values of several Sverdrup. If these changes are true, they have either to be compensated by changing flow through other passages or they lead to changes in the water level. However, since all other passages have volume flows in the order of Sverdrup, i.e. of the same magnitude of or even smaller than the monthly changes derived for the Fram Strait flow, compensation of large changes can be doubted. Rudels et al. (2008) discussed that

misinterpreting erroneous Fram Strait net transport by flow through other passages would constrain the Arctic fresh water budget in an implausible way. Similar considerations hold for strong variability of the net flow on monthly to annual scales. Both the Canadian Archipelago and the Barents Sea are shallow and only light upper layer water can pass. Flow changes of the order of several Sverdrup over several months would alter the storage of upper layer water masses in a way that would require explanations through changes of the production.

If the net flow changes in Fram Strait are not compensated, the sea level in the Arctic Ocean must change. The amount of water provided by a one-month imbalance of 5 Sv, distributed over the whole Arctic Ocean (approximate area: 10 million km^2), results in a mean water level change of more than 1 m. If the water is not evenly distributed, local sea level changes would be even larger. This should be visible in tidal gauges.

A second weak point of the presented attempt to derive heat transport is the inherent assumption that no mixing between the Fram Strait and Barents Sea branches takes place. This is certainly not true. Both eddies, likely caused by baroclinic instability at the confluence of the two branches (Schauer et al., 2002), as well as interleaving and consecutive double-diffusive fluxes (Rudels et al., 1999) have been shown to exchange water between the branches and to reduce the respective extremes of temperature and salinity. Since Barents Sea water has been cooled to less than 0°C when it encounters the Fram Strait branch in the northern Kara Sea our approach might overestimate the heat transport divergence if it captures cold Barents Sea water in the outflow part of the assumed stream tube instead of warmer Fram Strait Water which is returning further west. A proper solution to the problem would be to extend the stream tube concept to involve the Barents Sea inflow. Data of the flow into the western Barents Sea are available (Ingvaldsen et al., 2004) but the heat loss to the atmosphere inside the Barents Sea is much larger than the heat loss of Fram Strait water in the Central Arctic and thus would dominate the combined signal. Data of Barents Sea water temperature and flow in the St. Anna Trough are not available; calculation with constant values could, however, provide another approach.

6 Consequences for observational strategies

This paper elucidates difficulties inherent in determination of the oceanic heat delivered through advection to the Arctic Ocean. The above considerations also point to considerable consequences and tasks for observational strategies. First, sections (or openings) have to be identified that are connected by stream tubes and that have therefore zero net volume flow such that in fact heat transport divergence can be derived. To compute heat transport variability, it has to be taken into account that changes of volume transport might be compensated through different sections in different time periods. Consequently, simultaneous observations are needed across those sections. Second, because the heat transport term is non-linear, these observations need to be made at sufficient spatial resolution to capture the velocity and temperature structure over the sections.

In the Arctic the lateral variability is of the scale of several kilometres due to the small internal Rossby Radius. Together with the complex topography in Fram Strait this translates directly into the need of a high number of moorings since so far this is the only way for time series of appropriate horizontal resolution. But even with the high instrumental coverage as in the mooring array used here, the resolution is not enough to capture the volume flow with sufficient accuracy. Furthermore, in order to assess what fraction of the heat is released to the surface in the Arctic Ocean vs. what fraction is simply passing and thus available in the Arctic ocean temporarily, 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. Given the various causes of uncertainties for the volume transport and for the definition of a stream tube which result in large errors of the estimated heat transports and their variability the question can be raised if mooring arrays are the appropriate tool to address this problem.

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Table 1. Heat balance of a simplified Arctic Ocean (the numbers are not meant to describe the real Arctic Ocean). In- and outflow (voltransp) of water with temperatures, Temp, through the various openings cause temperature transports that sum up to heat transport to the Arctic Ocean. For the left rows the Celsius scale was used, for the right rows the calculation was done with absolute temperatures.

	Temp (˚C)	Voltransp (Sv)	Temp transport (TW)	Temp (˚C)	Temp transport (TW)
Tref	0			−273	
Bering Strait	2	1	8	275	1100
Can Archip	−0.7	−1	2.8	272.3	−1089.2
Barents Sea in	5	2.5	50	278	2780
Barents Sea out	2	−0.5	−4	275	−550
Fram Strait in	3	9	108	276	9936
Fram Strait out	−1	−11	44	272	−11 968
		Net (Sv)	Heat transport (TW)		Heat transport (TW)
Arctic Ocean total		**0**	**208.8**		**208.8**

Table 2. As in Table 1. The rows "change to Table 1" give the "temperature transport", respectively heat transport difference between the rows in Table 2 and the respective rows in Table 1. Changed parameters are bold.

	Temp (°C)	Voltransp (Sv)	Temp transport (TW)	Change to Table 1 (TW)	Temp (°C)	Temp transport (TW)	Table 1 (TW) Change to
Tref	0				−273		
Bering Strait	2	1	8		275	1100	
Can Archip	−0.7	−1	2.8		272.3	−1089	
Barents Sea in	5	2.5	50		278	2780	
Barents Sea out	2	−0.5	−4		275	−550	
Fram Strait in	3	**10**	120	12	276	11040	1104
Fram Strait out	−1	**−12**	48		272	−13 056	
		Net (Sv)	Heat transport (TW)			Heat transport (TW)	
Arctic Ocean total		**0**	**224.8**	**16**		**224.8**	**16**

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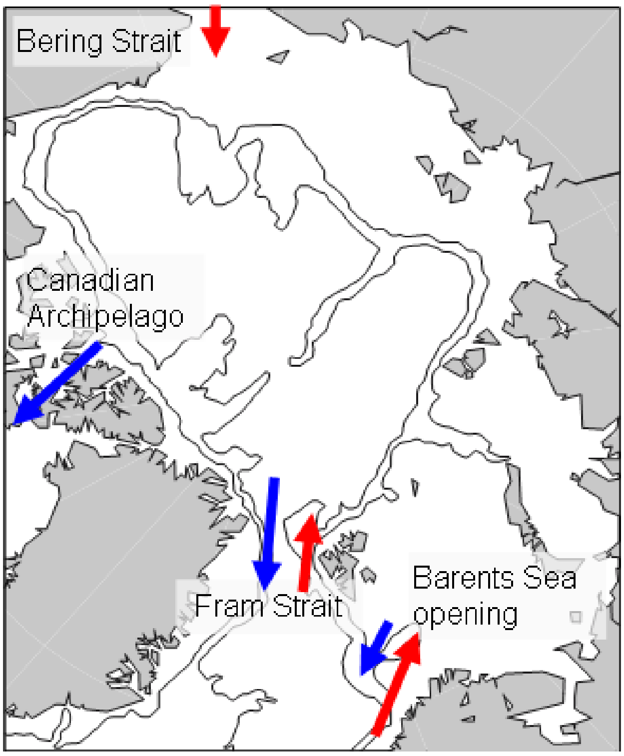

Fig. 1. Map of the Arctic Ocean with a sketch of the flows through passages to the Atlantic and Pacific (see Table 1).

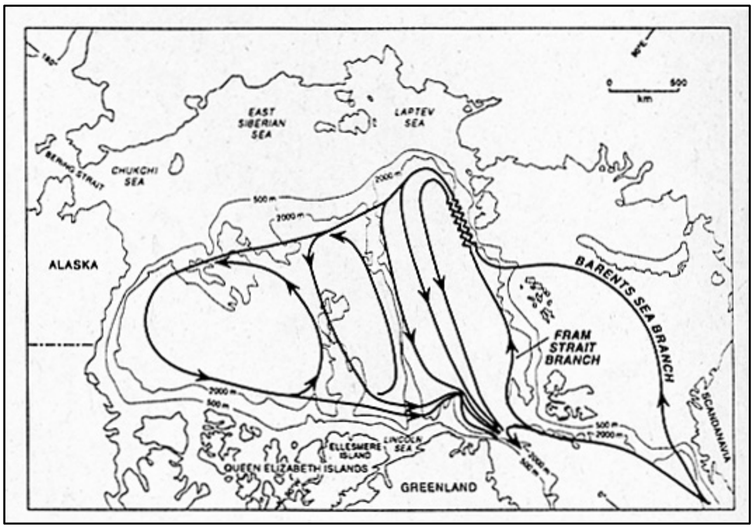

Fig. 2. Circulation scheme of Atlantic-derived water in the Arctic Ocean (Fig. 9 from Rudels et al., 1994).

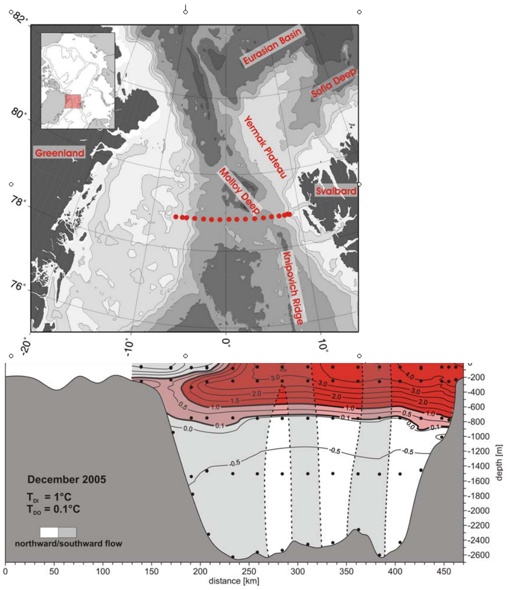

Fig. 3. Upper panel: location of the mooring array in Fram Strait (red dots) maintained from 1997 until 2008, numbers of moorings varying between 12 and 17. Lower panel: average temperature and cross-section flow direction of December 2005. At each dot velocity and temperature was measured. Lines with numbers denote isotherms in ˚C. Dark red marks the range which is warmer than T_{DI}=1˚C (see text for explanation), light red marks the additional range of water warmer than T_{DO} which was 0.1˚C for this month. Gray shading indicates southward flow; no gray shading indicates northward flow. Note that our integration was done over areas in dark red without gray shading for northward flow and over dark red plus light red areas with gray shading for southward flow to give zero net transport.

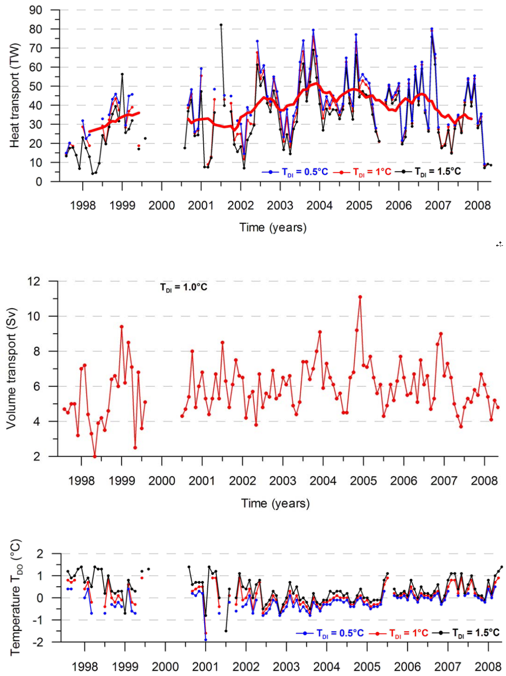

Fig. 4. Time series of heat transport (upper panel) to the Arctic through the WSC water at inflow temperatures higher than T_{DI}=0.5˚C (blue), 1˚C (red) and 1.5˚C (black). The dots connected by thin curves are computed from monthly mean values and the bold curve is the yearly running mean. Middle panel: monthly mean northward volume transport of water warmer than T_{DI}=1˚C. The lower panel gives the respective limiting outflow temperatures T_{DO} (see text for explanation). The gap between summers 1999 and 2000 is due to missing moorings in the Central Fram Strait.

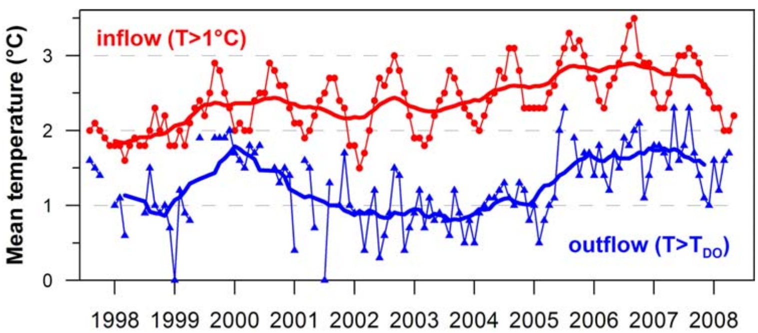

Fig. 5. Time series of the mean temperature of the northward flow (red) of water warmer than 1˚C and of the respective southward flow (blue). Water flowing southward is warmer than the temperature T_{DO} given by the red curve in the lower panel of Fig. 4.

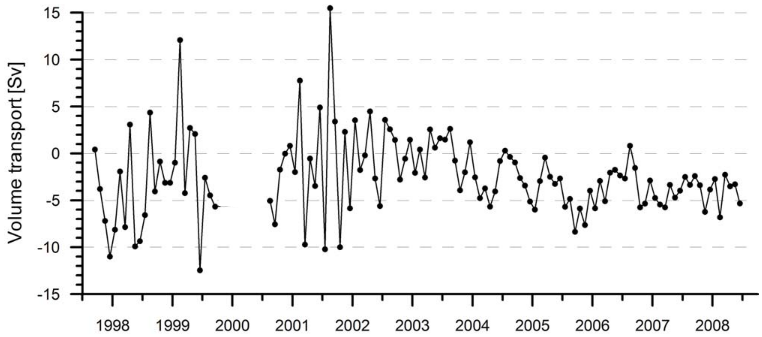

Fig. 6. Monthly averages of total net volume transport through the Fram Strait derived from velocity measurements with moorings.