The Atlantic inflow to the Barents Sea

by

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1. Introduction

The inflow of Atlantic Water (AW) is of crucial importance for the physical and ecological conditions of the Barents Sea. With the inflowing Atlantic current, fish larvae and zooplankton are advected into the Barents Sea from the Norwegian Sea (e.g. Ozhigin and Ushakov, 1985; Skjoldal and Rey, 1989; Skjoldal et al., 1992; Giske et al., 1998). The climatic conditions affect the growth rate and distribution of zooplankton and fish larvae, as well as fish population parameters as growth, recruitment, migration and distribution (e.g. Ottersen and Loeng, 2000; Stenseth et al., 2002). Recent publications have proposed that a warming of the Barents Sea may lead to a redistribution of the commercial important fish species in the Barents Sea (Blindheim et al., 2001). Thus, an understanding of the variability of the Atlantic inflow is highly important.

The Barents Sea also influences the Arctic Ocean. Both by providing a pathway for AW to the Arctic Ocean, but also as a shallow shelf sea producing dense water through cooling and brine release. According to Schauer et al. (2002a), the Barents Sea plays a special role among the other shelf seas considering the ventilation of the Arctic Ocean because: 1) being close to the Norwegian Sea it receives the most saline marine inflow, 2) through its permanent inflow from the Nordic Seas it continuously supplies water to the central Arctic basins (Rudels, 1987; Rudels et al., 1994), 3) being the deepest Arctic shelf sea, and 4) by receiving only little river input. Observations have shown that the Barents Sea provides intermediate water down to 1200 m depth in the Arctic Ocean (Rudels et al., 1994; Schauer et al., 1997), and, together with the Kara Sea, the Barents Sea is the only source area for shelf waters ventilating the Nansen Basin below the halocline (Schauer et al., 1997). Recent investigations suggest that the throughflow of AW also can take part in a process where water distilled at the surface of the Arctic by freezing ends up at mid-depth in the same ocean (Aagaard and Woodgate, 2001). The process starts with an increase in the export of ice from the Arctic Ocean into the Barents Sea. The ice melts due to contact with the AW; whereby the melt water is entrained in the Barents Sea throughflow and subsequently sinks into the Arctic Ocean. Woodgate et al. (2001) ascribed an observed cooling and freshening of the Atlantic layer in the Arctic to this process, and also Schauer et al. (2002b) pointed out that that the observed freshening was due to a larger input of fresh water to the Barents Sea.

Although the densest bottom water from the Barents Sea flows into the Arctic Ocean (Midttun, 1985), the Barents Sea also is a source for dense bottom water to the Norwegian Sea (e.g. Sarynina, 1969; Blindheim, 1989, Quadfasel et al., 1988). Thus, knowledge of the
variability of the Atlantic inflow to the Barents Sea is important for the understanding of the climatic state of a wider region, and for evaluations of climate change.

The purpose of this thesis has been to identify the forcing responsible for the observed variability in the Atlantic inflow and distribution in the Barents Sea. Section 2 gives a brief overview of the current systems transporting the AW into and within the Barents Sea (2.1), and a brief description of the regional atmospheric fields (2.2). It also gives an overview of the recent investigations considering Atlantic flow and climate variability in the Barents Sea (2.3). In section 3 the results of Paper I-IV are combined and summarized with emphasis on describing the velocity field of the western entrance of the Barents Sea (3.1), the Atlantic transports (3.2), the distribution of AW between Norway and Bear Island (3.3) and climate variability and distribution of AW within the Barents Sea (3.4). In section 4 some reflections on possible consequences based on a combination of results from Paper I-IV are outlined (4.1-4.3). Future perspectives are given at the end.

2. The Atlantic inflow to the Barents Sea
2.1 Large-scale features of the Atlantic inflow

South in the Norwegian Sea, the Norwegian Atlantic Current (NAC) appears as a well defined two-branched current with a western and eastern branch (Orvik et al., 2001) (Figure 1). The western branch appears as an unstable frontal jet carrying a total mean of 3.4 Sv (1 Sv = $10^6$ m$^3$s$^{-1}$), while the eastern appears as a topographically trapped, near barotropic current with an annual mean transport of 4.2 Sv (Orvik et al., 2001). The eastern branch shows a systematic annual cycle with a doubling of the inflow during winter (Orvik et al., 2001). According to Poulain et al. (1996) the two branches join to one single swift current north of 68°N, while recent investigations from Orvik and Niiler (2002) indicate that the NAC maintain its two-branch structure into the Fram Strait. At the entrance to the Barents Sea the current splits in one branch towards the Barents Sea and another towards Spitsbergen, and the Barents Sea probably receives its main input from the eastern branch (Hansen and Østerhus, 2000; Orvik and Niiler, 2002).

When entering the Barents Sea the NAC splits further in two main branches (Figure 1). One branch continues eastwards parallel to the coast and changes name to the Murman Current. During some years the influence of AW may be traced as far as the western shelf of Novaya Zemlya, while in others AW can hardly be traced east of 40°E (Dickson et al., 1970). The other main branch turns north along the Hopen Trench and divides into smaller branches.
Figure 1. Temperature in 100 m depth and main circulation (upper plate). Solid lines show flow of Atlantic Water, dotted lines Arctic Water, and dashed line Coastal Water. Topography of the Barents Sea (lower plate), and location of current meter moorings and the repeat hydrographic section between Norway and Bear Island.
One continues eastwards south of the Central Bank, and others continue as intermediate currents going northwards towards the Hopen Trench and eastwards between the Great Bank and the Central Bank and (Loeng, 1991). Eventually some of the AW leaves the Barents Sea as a westward current in the Bear Island Trough. AW may possibly also enter the Barents Sea at the southern slope of the Svalbard Bank (e.g. Li and McClimens, 1998), and smaller amounts of AW enter the northern Barents Sea east of Svalbard (e.g. Mosby, 1938) and between Franz Josef Land and Novaya Zemlya (e.g. Hanzlick and Aagard, 1980). This thesis will concentrate on the AW entering in the western boundary of the Barents Sea.

2.2 Large-scale features of the atmospheric fields

The most prominent and recurrent pattern of both seasonal and long-term atmospheric variability in the North Atlantic Ocean is the North Atlantic Oscillation (NAO). An important factor in the NAO is the Icelandic low, which during winter gives prevailing southwesterlies in the Norwegian Sea (Figure 2). The variability of the NAC in the Norwegian Sea has a strong coupling to the regional wind field. Stronger southwesterlies gives higher transports (Mork and Blindheim, 2000; Orvik et al., 2001) and, although with a delay of 2-3 years, a narrower current (Blindheim et al., 2000). The wind field also has a substantial impact on the relative proportions going towards the Fram Strait and the Barents Sea (Furevik, 1998). During winter the Icelandic low stretches as a trough towards the Barents Sea. In contradiction to the situation in the Norwegian Sea where the Icelandic low gives a uniform wind field across the most of the NAC, the situation in the Barents Sea Opening (BSO) are southwesterly winds in the southern part and northeasterly winds in the northern part (Figure 2b).

2.3 Previous investigations of the Atlantic flow in the Barents Sea

The first broad analysis of hydrography, currents and climate variability of the Barents Sea was made by Helland-Hansen and Nansen (1909). Since then, numerous investigations have been carried out, and an overview of earlier literature can be found in Loeng (1991) and Haugan (1999). Here, only a brief overview of the most recent literature considering the circulation and transformation of AW in the Barents Sea will be given. Furevik (1998) investigated the bifurcation of the NAC west of the BSO with a numerical model, and found that with realistic topography and stratification, 65-70% of the prescribed inflowing water went northwards toward the Fram Strait and 30-35% eastwards towards the Barents Sea. Mesoscale and large-scale variations in the wind forcing had a large impact on the
Figure 2. Mean sea level pressure (a) and wind field (b) during winter (December through March) and summer (June through August) for the period 1997-2001. The data were obtained from the hindcast database of the Norwegian Meteorological Institute.

bifurcation, and could give a substantial increase in the amount going eastwards. Recent investigations of the currents in the inflow area were made by Haugan (1999), who found that the inflow take place in two cores with a return flow between them. His results were supported by Furevik (2001) and O’Dwyer et al. (2001). Li (1995) and Li and Mc Climans (1998) showed that AW also may enter as a retrograde slope jet on the southern slope of the Svalbard Bank, and their results were documented by current measurements during winter presented in Loeng and Sætre (2001), and by model studies (Ådlandsvik and Hansen, 1998). Recirculation of AW in the Bear Island Trough were investigated by Gawarkiewicz and Plueddemann (1994 and 1995), who concluded that the recirculating water is topographically trapped on the slope of the Svalbard Bank. The inflow of AW in the northern Barents Sea was investigated by Pfirman et al. (1994) and Løyning (2001) for the area east of Svalbard,
and Hanzlick and Aagard (1980), Schauer et al. (2002a) and Schauer et al. (2002b) for the area east of Franz Josef Land.

When passing through the Barents Sea the AW is strongly modified by cooling, mixing and freezing during winter, and all the AW entering in the west is modified and leaves the shelf toward the Arctic Ocean mostly with temperatures below 0°C (Schauer et al., 2002b). The circulation and transformation of AW inside the Barents Sea were described by (among others) McClimans and Nilsen (1993), Pfirmann et al. (1994), Parsons et al. (1996), Loeng et al., (1997), Haugan (1999), Loeng and Sætre (2001), Schauer et al. (2002a) and Schauer et al. (2002b), while the flow at the eastern border of the Barents Sea was investigated by Loeng et al. (1993), Schauer et al. (2002a) and Schauer et al. (2002b). Considering the heat fluxes of the Barents Sea, this was investigated by Häkkinen and Cavalieri (1989), Simonsen and Haugan (1996), Haugan (1999) and O’Dwyer et al. (2001). Bottom water formation in the eastern Barents Sea and on the Svalbard Bank has been known for a long time, but more recently this process has also been documented at the Central Bank (Quadfasel et al., 1992), the Great Bank (Løyning, 2001), and in Storfjorden (e.g. Quadfasel et al., 1988; Schauer and Fahrbach, 1999; Skogseth and Haugan, 2003). Bottom water formation in the Barents Sea was also investigated by Harms (1997) using a numerical model.

Several previous studies have given estimates of the volume flux across the BSO, but the only transports based on moored current meters were from a 2-month time series (Blindheim, 1989). The other published estimations of volume fluxes have been based on vessel-mounted Acoustic Doppler Profiler (Haugan, 1999; O’Dwyer et al., 2001), budget considerations (Loeng et al., 1997; Haugan, 1999), numerical modelling (e.g. Ådlandsvik and Loeng, 1991; Parsons, 1995; Harms, 1992), and geostrophic calculations of the baroclinic currents (numerous Russian studies). Climatic variability in the Barents Sea has been studied by numerous authors, the more recent Grotefendt et al. (1998), Dickson et al. (2000), Furevik (2001), Ottersen and Stenseth (2001) and Ottersen et al. (2003).

3. Summary and main results

The Atlantic inflow and distribution in the Barents Sea were investigated using hydrographic measurements and 4-year long records (August 1997-August 2001) of moored current meters at a section between Norway and Bear Island (Figure 1). The current measurements programme started in a climatically cold period (Figure 3), but in late
winter/early spring in 1998 the climate conditions changed to a warm period that lasted throughout the measurement period.

The main effort of this thesis has been to examine the flow in the inflow area, and Papers I-III deals with this approach. In Paper I the variability of the velocity field and transports were investigated based on the first year of the current measurements. Emphasis was made on describing the monthly and seasonal variability, and the results revealed (among others) that substantial changes in the velocity field occur from one month to another. Considering the seasonal variability, no clear signal in the transport was found, but a change in the velocity field between winter and summer appeared. The time series were extended to 4-year long records, and the velocity fields were further investigated in Paper II; first briefly on a few-day time scale, then more thoroughly on a fortnightly time scale. The results revealed that the velocity field adopts patterns that to a large degree are determined by variations in sea level induced by changes in the wind field. The seasonal cycle was further investigated in Paper III. The investigation showed a higher Atlantic inflow during winter than summer and related this to the shear of the local wind field. It was also shown that the first year of the measurement period was likely to be abnormal considering the usually present seasonal cycle in AW transport, but normal considering the seasonal cycle in the velocity field (although the latter was not expressed explicitly). The climatic variability in the Barents Sea related to the Atlantic inflow was investigated in Paper IV. The results indicated that the warming of the 1990s might be due to natural variability rather than anthropogenic effects. It was also shown that the regional wind field has significant influence on the Atlantic inflow to the Barents Sea, while local atmospheric forcing seems to be dominating for the distribution of the AW within the Barents Sea. In the following the results from Paper I-IV are recapitulated with
focus on describing the different aspects of the Atlantic inflow without addressing the actual paper from which the results are taken.

3.1 The velocity field

The velocity field in the Atlantic inflow to the Barents Sea shows substantial and complex variations both on a few-day and on monthly and seasonal time scales. In general the velocity field is dominated by frequent and large fluctuations. The flow is predominately barotropic, and sea level changes induced by the local wind field seem to be a key factor determining the spatial distribution of the velocity field. The main process is that the wind field creates converging and diverging Ekman transports that accumulate water within the section thereby creating barotropic pressure gradients and associated geostrophic currents. This process can generate sea level gradients offshore in the Bear Island Trough through a shear in the wind field, and are enhanced by the topographic constraints with the Norwegian coast in the southern parts and the open areas in the northern parts.

The mean velocity field shows the Atlantic inflow as a wide flow occupying most of the section between 71°30’N and 73°30’N (except in the deeper parts of the Bear Island Trough). The core is located at 72°30’N, but is not stable at this location as there may also be outflow there. Southwesterly winds in the southern part of the section and southeasterly winds further north accumulate surface water in the southern part, and at the same time move the surface waters of the northern part out of the section. The result is a pressure gradient creating a geostrophic velocity field enhancing the wide inflow. When the predominant wind direction is northeasterly, the Ekman transport will produce a higher water level in the north than in the south. The associated pressure gradient drives a relatively strong outflow in the northern part and a weaker inflow in the south. The two flow regimes may be persistent for several weeks, and are related to the relative strength and lateral extension of the Icelandic low and an Arctic high-pressure system. However, the alignment of the local isobars must be considered to describe the details of the flow.

When influenced by local wind or downstream processes, the velocity field can adopt different patterns. Apparently a wind field with a strong cross-sectional component creates moderate sea level gradients offshore in the Bear Island Trough area, which causes the flow to occur in distinct cores with return flow or stagnant water between them.

A seasonal signal in the velocity fields was found. During winter the currents are intensified especially in the middle parts of the section. The strongest seasonal signal is in the upper 150-200 m at the northern boundary of the AW flow (near 73°30’N) showing an
outflow during winter and an inflow during summer (Figure 4). Near 71°30’N, there is almost no seasonal signal at all. The seasonality in the middle part of the section is related to sea level changes within the section induced by a shear in the cross-sectional wind stress. The middle part of the section experiences a stronger shear in the wind field during the winter than in summer, thereby inducing stronger net Ekman transports and stronger water level gradients. The barotropic currents therefore give a relatively strong seasonal signal in these areas. Near 71°30’N on the other hand, the alluring result arises that both wind speed and direction change substantially from winter to summer, but the net Ekman transport does not. The water level gradient, and the associated geostrophic current, therefore has almost no seasonal signal in that area. The reason for the seasonal signal found in the northern parts is not properly examined and therefore not fully understood, although it is likely to be connected to the wind field.

3.2 Atlantic water transports

The mean AW transport to the Barents Sea for the period August 1997-August 2001 was a net inflow of 1.5 Sv. The warm core jet on the slope of the Svalbard Bank (Li and Mc Climans, 1998) is not included in this estimate. During years with abnormal transports, for instance like the first year of the measurements, the annual mean transport may differ substantially from this estimate. In some periods, the flow is reversed, and there is a net Atlantic outflow from the Barents Sea towards the Norwegian Sea. Based on the first year of the measurements, the monthly mean transport fluctuated over a range of almost 10 Sv. This variability was, in some periods, clearly linked to the local atmospheric pressure fields. In

![Figure 4](image_url)

**Figure 4.** Schematics of the distribution of AW (inside the thick line) and direction of the Atlantic flow during winter and summer. Grey areas show eastward flow, and white areas (inside the thick line) show westward flow. The figure is drawn with basis on the mean current during winter and summer for the 4 year period (updated version of Figure 8 in Paper I), in combination with the seasonal distribution of AW based on repeated sections of hydrography (Figure 6 in Paper III).
other periods, the correlation between the local surface atmospheric pressure and volume flux was poor. The first year of the measurements showed that the variability in the inflow was, to some extent, connected to two factors: 1) the temporal variability in the deeper parts at the core at 72°30’N, and 2) the spatial variability in the northern boundary of the inflow. By assuming the spatial structure of the current being invariant over interannual time scales (which was confirmed at least for the 4-year period, section 3 and Figure 4), this indicates that only two current meters are sufficient to cover the section between 71°15’N and 73°45’N for long-time monitoring (i.e. monthly to seasonal). Additional instruments may be necessary if variability on shorter time scales is to be investigated.

Considering the seasonal signal, a higher inflow during winter than summer was found except for the first year of the measurements. The increase during winter is mainly due to stronger currents in middle parts of the section as described for the velocity fields, and due to the presence of inflowing AW in the upper 50 m (Figure 4). The seasonality is counteracted by a weaker Atlantic inflow in the southernmost areas during winter, and a stronger recirculation (Figure 4). The mean AW transport was estimated to 1.7 Sv during winter and 1.3 Sv during summer, but with a pronounced minimum in inflow or even outflow in spring due to an annual event of northerly winds.

3.3 Distribution of AW within the BSO

A seasonal signal in the distribution of AW within the section was found, showing that the Atlantic domain is displaced northwards during winter compared to summer (Figure 4). In the south, this is connected to coastal downwelling during winter. The southwesterly winds dominating during winter (Figure 2b) give an onshore Ekman transport, the water level along the Norwegian coast is raised, and coastal downwelling moves the AW down-slope away from the coastline. During summer the winds are reversed, but weaker, and upwelling is not necessarily initiated. In the northern part of the section, the northward displacement during winter is probably due to stronger recirculation associated with the stronger winds and/or the stronger inflow.

3.4 Climatic variability and the distribution of AW in the Barents Sea

Based on hydrographic measurements in three standard sections in the Barents Sea, horizontal fields of ocean temperature in 100 m depth and surface air temperatures, the climatic variability in the Barents Sea was investigated. The Barents Sea climate fluctuates between warm and cold periods. By comparing decade by decade we found that although the
1990s had high temperatures, both the 1930s and the 1950s were warmer. This indicates that the warming of the 1990s may very well be related to natural variability rather than anthropogenic effects.

The effect of the NAO in the Barents Sea was evaluated, and it was pointed out that during positive NAO winter index (Hurrell, 1995) the joint action of (at least) three oceanic responses might give an increase in the Barents Sea temperatures. The responses are: 1) stronger southwesterly winds enhancing the inflow, 2) more winter storms penetrating the Barents Sea, 3) the narrowing of the NAC (Blindheim et al., 2000), and the associated lower heat loss to the atmosphere, giving higher temperatures of the inflowing AW. It was then shown that the NAO, through regional and local effects, has a significant influence on the Barents Sea on decadal time scales. Strong influence also occurs during extreme NAO events. Still, local atmospheric forcing not captured by the NAO winter index seems to be dominating for the distribution of the water masses within the area, at least on annual time scales. The local pressure field appears to change the relative strength of the two branches of AW going northeast in the Hopen Trench and east in the Murman Current (Figure 1), thereby having a significant effect on the local climate. The local pressure distribution not captured by the NAO winter index also has some influence on the total inflow to the Barents Sea.

4. **Possible consequences of the results**

In the following, some aspects that arose during the progress of the papers will be outlined. These should not be viewed as thorough investigations; rather as reflections that all need further study before any conclusions can be made.

4.1 **Possible connections between the Atlantic inflow and upstream variability**

When investigating the velocities in Paper II, the current across the BSO was decomposed by EOF analysis into different anomalously velocity fields, where the strength of the fields varied in time (Figure 7 in Paper II). The most dominant field (EOF1) showed a wide inflow or outflow centred at $72^\circ 30'N$, and explained 34% of the variability in the velocity field. By reconstructing the velocity field from the EOF1, then calculating the transport, a time series representing the total transport associated with the wide flow pattern is obtained (Figure 5). Compared to the flux of AW from the original data, the time series vary much in phase, and the correlation coefficient is 0.88. This can be interpreted as that although the velocity field adopts many different patterns, about 77% of the variability in the AW transport is related to
temporal changes in the core at 72°30′N. A similar result was found when investigating the first year of the measurements (Paper I).

The anomalous velocity field represented by the EOF1 was found to be forced by the local wind field (through accumulation of water within the BSO). If this is correct, the strong correlation can also be interpreted as that the local wind field forces about 77% of the variability in the AW transport to the Barents Sea, leaving only about 23% of the variability to the remotely forced NAC, bottom water production etc. This is consistent with Furevik (1998) who found that the relative strength of the flow eastwards towards the Barents Sea and northwards towards the Fram Strait to a large degree is determined by the local wind forcing. His results showed that while the effect of topography and stratification gave a transport of 65-70% of the prescribed inflow northwards and 30-35% eastwards, the mean winds during the winter changed the relative proportions to about 30% northwards and 70% eastwards. The presence of polar lows in the BSO gave an even larger amount going into the Barents Sea. Since the winds are highly variable, it means that the AW transport into the Barents Sea is only weakly related to the AW transport upstream in the NAC (assuming that there is enough AW upstream to supply what’s needed, and taking into account that some AW enters the Barents Sea due to topographic steering). The temperature of the AW going into the Barents Sea depends on the temperature of the AW west of the BSO. This temperature depends on the NAC upstream temperature, NAC upstream transport, and the heat loss along the way. This means that although the temperature of the AW going into the Barents Sea depends on the upstream transport, a hypothesis may be proposed saying that the Barents Sea in general will be more affected by upstream variability in temperature than upstream variability in transport. Further investigations are obviously needed to test this hypothesis.

Figure 5. Transport of all water masses estimated from the reconstructed velocity field of EOF1 (dashed line), and transport of AW estimated from the original data. All data have been filtered removing fluctuations with periods less than 14 days prior to calculations. Positive transport is into the Barents Sea (i.e. eastward).
McClimans et al. (1999) proposed that the weekly to monthly variability in the barotropic slope current of the NAC is forced by sea level changes in the northeast Atlantic propagating along the Norwegian coast as barotropic Kelvin waves, and support for this was found in analysis of the Norwegian Sea. The assumption behind the theory was that the local atmospheric variability will be averaged out on weekly to monthly time scales. However, the dominating effect of the local winds found here makes it unlikely that such an assumption is valid for the Atlantic inflow the Barents Sea. The signal will be there, but it will probably be small compared to the effect of the wind.

4.2 Interannual variability of AW in the BSO

The results from Paper III showed that the AW in the BSO is displaced northwards during the winter compared to the summer. The stronger downwelling favourable along-coast winds during the winter were proposed as the main forcing mechanism in the southern part. The reason for the displacement in the northern part was not identified, although a stronger recirculation of AW was proposed. Anyway, it is reasonable to expect this feature (either directly or indirectly) to be connected to the atmospheric fields during winter, and these fields are characterized by the strength and extension of the Icelandic low into the Norwegian Sea (section 2.2). As the interannual variability of the Icelandic low is substantial, there is also a possible interannual variability in the distribution of AW within the BSO. By using the seasonal cycle as basis, a hypothesis is proposed saying that the distribution of AW in the BSO are more northwards when the Icelandic low is strong. The NAO winter index (updated from Hurrell, 1995) is used to capture the strength of the Icelandic low. To quantify the distribution of AW in the BSO, the areas occupied by AW north of 73°45′N and south of 71°15′N were calculated (from the repeat hydrographic sections as in Paper III). The results (Figure 6a-b) indicate a clear relation with a larger amount of AW on the slope of the Svalbard Bank when the NAO winter index is high, i.e. a more northward distribution in the northern part when the Icelandic low is strong\(^1\). However, there is also a larger amount of AW near the Norwegian coast when the NAO winter index is high (Figure 6c), although this signal is much weaker than in the north. According to the hypothesis this should fluctuate in an opposite manner, indicating that the situation observed in the seasonal cycle cannot be extended to interannual time scales in the southern area. Also evident from the figure is that

\[^{1}\text{There is a lag between some of the parameters in Figure 6. This will not be investigated in this preliminary analysis.}\]
The interannual variability in the northern area is more dominant than the seasonal cycle, while the seasonal variability dominates in the southern area.

The above results indicate a positive correlation between the NAO winter index and the area occupied by AW, a result clearly evident when investigating the total area across the BSO occupied by AW (Figure 6d). Earlier investigations have shown a positive correlation between the NAO winter index and the mean AW temperature in the BSO (also evident in Figure 6e). This means that both the temperature and the extent of AW increase with increasing NAO winter index (Figure 6a and d-e), although with different lags. These results oppose the situation found in the Norwegian Sea where a stronger Icelandic low is associated with higher temperatures but a narrower current (Blindheim et al., 2000; Mork and
Blindheim, 2003). The different response in the BSO is connected to the non-uniform wind field across the BSO and the presence of recirculating water in the north. A stronger Icelandic low will in general give stronger southwesterlies across most of the NAC in the Norwegian Sea (Figure 2b), pushing it towards the Norwegian coast thereby giving a narrower current. In the Barents Sea the stronger Icelandic low will give stronger southwesterlies in the south, pushing the NAC towards the Norwegian coast. At the same time there will be stronger northeasterlies in the north, pushing the AW and the Polar Front northwards leaving more room for the AW. In addition, the stronger inflow associated with the stronger winds may increase the recirculation on the southern slope of the Svalbard Bank, thereby pushing the Polar Front up-slope and increasing the lateral extent of AW. This can explain the simultaneous increase in temperature and lateral extent. Other aspects may also have influence, for example that the Polar Front is moved up-slope by processes not directly connected to the AW, thereby leaving more room for AW to spread out in the BSO, or the presence of an inflow of AW on the southern slope of the Svalbard Bank (e.g. Li and McClimans, 1998). However, these aspects serve probably more as amplifying effects than driving mechanisms.

In summary, this preliminary investigation has shown that both the mean temperature and lateral extent of AW in the BSO is positively correlated to the strength of the Icelandic low, although with lags. During years with a strong Icelandic low the temperature rises, and the Polar Front is located further up-slope towards Bear Island than years with a weaker Icelandic low. The front between the Atlantic and Coastal Waters is also located further up-slope towards the Norwegian coast when the Icelandic low is strong, although the variations are weaker.

4.3 Some thoughts on the spring event of northerly winds and wide outflows

Northeasterly winds result in wide outflows from the Barents Sea (Paper II), and an annual event of such winds was identified as driving mechanism for an observed minimum in Atlantic inflow (or even outflow) in spring (Paper III). According to Paper II the northeasterly winds was associated with a centre of high sea level pressure (SLP) located in the Arctic stretching towards the Barents Sea. These results were based on the 4-year period from August 1997-August 2001. To investigate if the presence of a spring event is invariant on interannual time scales, the seasonal cycle in the Arctic climate must be considered. The seasonal cycle in Arctic climate is substantial, and recent publications have shown an oscillatory behaviour in the Arctic SLP on interannual time scales (Thompson and Wallace,
The oscillatory behavior in SLP causes the wind-driven motion in the upper Arctic Ocean to alternate between anticyclonic (ACCR) and cyclonic (CCR) regimes (Proshutinsky and Johnson, 1997). The cyclonic circulation has been dominating since the early 1990s, but shifted to an anticyclonic state in 1998 (Johnson et al., 1999), indicating that the current measurements were taken during an ACCR regime. The difference in the seasonal cycle in the two regimes in the period 1946-1997 was investigated in Polyakov et al. (1999), and they separated the analysis in eight climatic regions (Figure 7).

Figure 7. Colored boxes denote eight climatic regions for analysis of the SLP data shown in Figure 8. Labeled boxes denote six regions that are not of interest in this study. From Polaykov et al. (1999).
Their analysis reveal a spring maximum in SLP in the Central Arctic Basin and the Franz Josef Land region (Figure 8), and in fact in most of the other regions as well. This implies an annual reoccurring spring event in Arctic climate, indicating that the observed spring event in the Barents Sea is representative also on the longer time scale. However, in the Western region (in the Barents Sea) and the Franz Josef Land region, the timing of the spring maximum in SLP is different during years of ACCR and CCR (Figure 8). To investigate this a measure of the northerly wind component is necessary. A rough approximation of the northerly geostrophic wind component for the Spitsbergen area will be the difference between the SLP in the Greenland Sea and the Western region, and the numbers of Table 1 are constructed by manually inspecting Figure 8. The results show that during CCR years the

![Figure 8. Monthly mean (solid lines) and annual mean (dashed lines) SLP averaged over all ACCR (thin lines) and CCR years (thick lines). Note that the SLP scale is different among regions. The SLP data were obtained from NCAR and are from the period 1946-1997. From Polakov et al. (1999).](image-url)
Table 1. SLP difference [hPa] between the Greenland Sea and the Barents Sea in ACCR and CCR years. Positive values means northerly winds.

<table>
<thead>
<tr>
<th></th>
<th>ACCR</th>
<th>CCR</th>
</tr>
</thead>
<tbody>
<tr>
<td>March</td>
<td>-1</td>
<td>-5</td>
</tr>
<tr>
<td>April</td>
<td>4</td>
<td>-2</td>
</tr>
<tr>
<td>May</td>
<td>1</td>
<td>3</td>
</tr>
</tbody>
</table>

annual spring event of northerly winds will take place about a month later and be weaker than during ACCR years, thus the spring minimum in Atlantic inflow (or outflow) may occur about a month later in CCR years compared to ACCR years.

The above results indicate that the timing and the strength of the minimum Atlantic inflow or outflow of AW from the Barents Sea is related to the regime of the Arctic. The regimes persist for 5-7 years (Proshutinsky and Johnson, 1997). When CCR dominates, the Icelandic low is intensified and a tongue of low SLP stretches towards the Barents Sea (Johnson et al., 1999). Moreover, due to excess ice and freshwater transport through the Fram Strait from the Arctic Ocean into the Greenland Sea, the AW flow through the Barents Sea is increased to maintain the water balance of the Arctic Ocean (Polyakov et al., 1999). This suggests a relation between the CCR regime of the Arctic and the warm periods of the Barents Sea (Paper IV), but the correlation is not straightforward (not shown). This may not be surprising as the regimes in the Arctic seems to be determined by the SLP at the North Pole, suggesting a secondary role for the NAO in terms of the two regimes (Johnson et al., 1999), while the NAO (or more precisely: the Icelandic low) is more important for the Barents Sea. The results may be interpreted as that the Atlantic inflow to the Barents Sea is mainly forced by processes in the Norwegian and Barents Seas, while the Arctic influences the inflow in special events only. Some possible impacts of the northerly winds and large outflows will be briefly outlined in the next sections; all of them need more investigation before any conclusions may be drawn.

4.3.1 Possible consequences of the northerly winds

Northerly winds in the Barents Sea will in general result in southward transports of ice and a decreasing surface air temperature. In the ice covered regions the northerly winds may lead to open leads and polynyas thereby giving an increased ice production (for example in Storfjorden, Skogseth and Haugan, 2003). Cooling and brine release are the driving factors in the process of forming dense water in the Arctic (Killworth, 1983; Carmack, 1986 and 1990;
The densest bottom water fills the basins of the eastern Barents Sea (Midttun, 1985), but bottom water with somewhat lower salinity is also formed on the Central Bank, on the Great Bank and on the Svalbard Bank, and some of this leaves the Barents Sea in the Bear Island Trough (e.g. Sarynina, 1969; Blindheim, 1989). The northerly winds may enhance the bottom water formation. Furthermore, the northerly winds may push the entire water masses of the northern and eastern Barents Sea southwards (Paper II). It is tempting to speculate that this would give an extra southward push to the cold waters on the shallow banks formed during winter. This is consistent with the water in the deepest part of the Bear Island Trough showing in general lower temperatures in summer than in winter (not shown), as a 4-5 months transit time seems reasonable.

In cold years with heavy ice formation, ice can drift south of the Polar Front and into the Atlantic water thereby melting (Loeng, 1991). It is possible that a strong spring event of northerly winds will increase the ice production in the ice-covered areas and at the same time being the onset of the melting season in the south. Such a scenery depends on the timing of the spring event; if it happens early (like in ACCR years) the freezing in north may be substantial while the melting in south will be less. The situation may be opposite in CCR years.

### 4.3.2 Possible consequences of the wide outflows

Large outflows from the Barents Sea will obviously have some effect in the Norwegian and Greenland Seas, especially when occurring in the spring when the Barents Sea has the lowest temperatures. How far these water masses are traceable, and how they modify the water masses in the Norwegian Sea and eventually the Fram Strait remains unknown.

Previous studies has shown that the abundance of the most dominant zooplankton in the Barents Sea, *Calanus finmarchicus*, is closely related to the supply of warm Atlantic water flowing from the Norwegian Sea (Skjoldal et al., 1992; Helle and Pennington, 1999; Helle, 2000), with higher abundance when the inflow is strong. *Calanus finmarchicus* spends the winter in the deeper parts of the Norwegian Sea and moves towards surface during late winter. The main transport of this species into the Barents Sea takes place in the spring, and a highly significant correlation between April/May temperatures in the western Barents Sea and the autumn biomass of larger zooplankton has been found (Dalpadado et al., 2003). This means that the timing and strength of the minimum inflow or outflow may have substantial impact on the import of zooplankton, thereby a strong impact on the ecosystem of the Barents Sea.
5. **Future perspectives**

Some aspects of further investigations needed to improve the understanding of the Atlantic inflow to the Barents Sea were outlined in the section 4. Additional possible future topics are:

1. **Distribution of AW inside the Barents Sea.** The available horizontal fields of observed temperature should be excellent for such an investigation, as these may be examined more thoroughly than in Paper IV. In addition, numerical model experiments may be performed, as the observational data are mainly from the summer period.

2. **Estimation of heat fluxes.** The heat loss from the Barents Sea is substantial (Simonsen and Haugan, 1996; Häkkinen and Cavalieri, 1989). The heat flux in the inflowing area can be calculated from the current measurements and surface heat fluxes from the horizontal fields of temperature. This could hopefully reveal valuable information on the heat budget of the Barents Sea.

3. **A better quantification of the Atlantic transport south of 71°15’N (including the Coastal Current that consists of 90% of AW), and north of 73°45’N.** This demands additional moorings in these areas, which will be (at least partly) fulfilled during the ECOBE and ProClim projects funded by the Norwegian Research Council from 2003.

4. **A better understanding of the relation between the water level and the velocity field.** To do so accurate and high-resolution water level is needed. This may be obtained from satellite data or high-resolution (horizontal grid < 5 km) numerical modelling. This is probably crucial to fully explain the situations when the flow occurs in several cores of inflow with return flow or stagnant water between.

5. **Model response to wind.** Numerical experiments with a high-resolution (horizontal grid < 5 km) model where the forcing is the atmospheric field associated with the different flow patterns in Paper II. This means a simulation with a strong Icelandic low stretching towards the Barents Sea, a simulation with a high-pressure system in the Arctic stretching towards Spitsbergen and a simulation with an atmospheric field creating easterly winds in the Barents Sea.
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Variability in the Atlantic inflow to the Barents Sea based on a one-year time series from moored current meters

by

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Variability in the Atlantic inflow to the Barents Sea based on a one-year time series from moored current meters

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Abstract

A 1-yr time series from an array of moorings across the Barents Sea Opening was investigated with emphasis on giving a description of the monthly to seasonal variability in the transport of Atlantic water (AW) into the Barents Sea. The results revealed an inflow with large fluctuations in both time and space, and showed that the circulation implied in the hydrographic conditions, with inflow in south and outflow in north, was not always the case. The flow might occur as a wide Atlantic inflow, as inflow and recirculation in narrow bands side-by-side, or as an outflow covering the occupied section all the way south to 72°N. The variability in the inflow of AW was, to some extent, determined by two factors: (1) the temporal variability in the deeper parts (i.e., below 300 m), and (2) the spatial variability in the northern boundary of the inflow. The volume transport across the section between 71°15′N and 73°45′N was calculated and gave a net inflow of AW of 2 Sv (1 Sv = 10⁶ m³ s⁻¹). In some periods, the inflow was reversed, and there was an outflow from the Barents Sea towards the Norwegian Sea. On a monthly basis, the transport fluctuated over a range of almost 10 Sv. The variability was, in some periods, clearly linked to the local atmospheric pressure fields, although additional forces were needed to account for the total flux. In some periods, the correlation between the local surface atmospheric pressure and volume flux was poor. A mesoscale eddy, present for about one month, was observed. A seasonal signal, consistent with the generally accepted seasonal cycle for the Barents Sea, was not found. However, the current measurements indicated a change in the flow field between summer and winter. During the winter, the frequent passing of atmospheric lows intensified the currents, producing a structure with strong lateral velocity gradients and a distinct, surface intensified, relatively high-velocity core of inflow and recirculation further north. During the summer, the weaker winds caused the inflowing area to be wider, the horizontal shear and the velocities to be lower, and there were two cores of inflow.

Keywords: Barents Sea; Mass transport; Seasonal cycle; Current oscillations; Barotropic motion

1. Introduction

The inflow of Atlantic water (AW) is of vital importance for both the physical and the biological conditions in the Barents Sea. Moreover, the Barents Sea serves as a thoroughfare for AW to the Arctic Ocean, where it affects both the Eurasian Basin (Rudels et al., 1994) and the Canadian Basin (Jones et al., 1995). Thus, an understanding of the variability of the Atlantic inflow to the Barents Sea is of crucial importance for the understanding of the climatic state of a wide region.
The main inflow of AW to the Barents Sea takes place through the Barents Sea Opening (BSO) in the western part (Fig. 1), and hydrographic characteristics of the water masses have been monitored on a standard section between Fugloøya (at the Norwegian coast) and Bear Island for many years by the Institute of Marine Research. The hydrography on the section indicates a rather stable inflow of AW in the southern part of the section, but characteristics of the AW may alter between seasons and years. Some of the AW recirculates in the Bear Island Trough and exits the Barents Sea as a westward current topographically trapped on the slope of the Svalbard Bank flowing parallel but southward of the cold Bear Island Current (Novitskiy, 1961; Pfirman et al., 1994; Gawarkiewicz and Plueddemann, 1995; Parsons et al., 1996; Li and Mcclimans, 1998; Adlandsvik and Hansen (1998); Loeng and Sætre (2001). AW may also recirculate further south in the section as semi-permanent countercurrents (Blindheim, 1989; Haugan, 1999) and in eddies, as the presence of eddies in this area seems to be considerable (Blindheim, 1989). The outflow of Arctic water takes place along the northern slope of the section. The circulation and transformation of AW inside the Barents Sea is described by several authors (e.g., Nansen, 1906; Novitskiy, 1961; Rudels, 1987; Haugan, 1999). Investigations considering the flow field on the southern slope of the Svalbard Bank have been performed by, among others, Gawarkiewicz and Plueddemann, 1995; Parsons et al. (1996); Li and Mcclimans (1998); Adlandsvik and Hansen (1998); Loeng and Sætre (2001).

The variability of the Atlantic inflow may be considerable (Adlandsvik and Loeng, 1991). Variability of time scales from day to year is clearly linked to atmospheric pressure, with the highest flow of water taking place when the atmospheric pressure in the Barents Sea region is low (Loeng et al., 1997). Although the net effect of atmospheric forcing on the transport is modest, fluctuations due to the local wind have been shown to be of the same magnitude as the mean transport (Adlandsvik and Loeng, 1991).

Based on 2 months of current measurements, Blindheim (1989) presented a picture of the circulation in the BSO consistent with hydrographic observations—an steady inflow in the south and outflow in the north. He calculated a mean transport of \(3.1 \text{ Sv} (1\text{ Sv} = 10^6 \text{m}^3\text{s}^{-1})\) into and \(1.2 \text{ Sv}\) out of the Barents Sea. With a 1-yr time series of current measurements at the outflow area between Novaya Zemlya and Franz Josef Land, Loeng et al. (1993) found volume fluxes varying from 0.7 to 3.2 Sv out of the Barents Sea, with an average of 1.9 Sv. Based on available literature considering all inflow/outflow areas, Loeng et al. (1997) made a balanced budget for the Barents Sea throughflow. Their results indicated an average total inflow of approximately \(4 \text{ Sv}\), of which the throughflow of AW contributed to half. Haugan (1999) presented a conceptual model of exchange across the BSO. He gave a mean total inflow of 5 Sv, whereby 2 Sv recirculated in the Bear Island Trough and 3 Sv was Barents Sea throughflow.

A seasonal cycle with higher inflow of AW to the Northern Seas during winter has been commonly accepted (e.g., Blindheim, 1993; Loeng et al., 1997). For the Barents Sea, this was deduced from geostrophic calculations of the baroclinic current carried out by numerous Russian scientists (Uralov, 1960; Timofeev, 1963; Moretsky and Stephanov, 1974; Orlov and Poroshin, 1988; Potanin and Korotov, 1988). As the currents in the Barents Sea are primarily barotropic in nature.
(e.g., Blindheim, 1989; Haugan, 1999), a seasonality in the baroclinic component may not necessarily give a seasonal signal in the total flux. A seasonal signal was also indicated in the current measurements from the outflow area in the northeastern parts of the Barents Sea (Loeng et al., 1993). The seasonality is closely connected to the seasonal variations in the regional atmospheric pressure fields. Calculations made by wind-driven models (Ádlandsvik, 1989; Harms, 1992) show that the southerly winds, which are dominating during the winter, increase the wind-driven part of the Atlantic inflow, while the weaker, more fluctuating easterly winds, which are common in the summer, decrease the wind-driven inflow. However, recent publications indicate that, for the subsurface part of the Atlantic inflow, seasonal variation is harder to find (e.g., Furevik, 2001; Haugan, 1999). This result is also partly supported by Hansen and Østerhus (2000), although their focus was the total Atlantic inflow to the Nordic Seas.

As the variability of the Atlantic inflow through the BSO impacts both the Barents Sea and the Arctic Ocean, an understanding of these fluctuations is essential. To give a good description of the variability, long-term current measurements are needed, but until now such observations have been rare. This paper presents a 1-yr time series from an array of moorings across the BSO. The main focus is directed towards giving a description of monthly to seasonal variability in the transport of AW to the Barents Sea as observed in the current measurements. Short time variability is discussed by Ingvaldsen et al. (1999).

2. Materials and methods

The data material consists of current measurements from 5 moorings with a total of 19 Aanderaa current meters RCM7 (Aanderaa Instruments, 1987) across the BSO in the period August 1997–August 1998. A sixth mooring was deployed on the slope of the Svalbard Bank but was dragged by a fishing vessel after a short time. In addition to current speed and direction, the instruments recorded temperature and conductivity. Details on the moorings are given in Table 1. It should be noted that mooring no. 1 was not deployed until late September 1997; that is, 1 month after the others. Data were recorded every 20 min, but were decimated to 1-h series. To fill in the gaps in the time series (see Table 1), simple linear interpolation of the velocities from the instrument above and/or below was performed. This should be an adequate method since the velocities are predominantly barotropic but may have strong lateral velocity gradients. The data were low-pass filtered using an order 4 Butterworth filter (Roberts and Roberts, 1978).

The volume flux was estimated by dividing the section between 71°15′ N and 73°45′ N into rectangles, with each current meter assigned a rectangle surrounding it. The transport within each rectangle was estimated from the east–west current component (i.e., the current component normal to the section). Since mooring no. 1 was not deployed before the end of September 1997, values from October 1997 (from the same mooring) were used to represent September 1997. The inflowing Coastal Current along the Norwegian coast and the currents on the slope of the Svalbard Bank are not included in the transport calculations.

Hydrographic observations were carried out on the section in August and October 1997 and January, March, April, June and August 1998 using CTD.

3. Results

The hydrographic conditions across the BSO are exemplified by observations from August 1998 (Fig. 2). This was a typical summer situation with a strong pycnocline at about 50 m depth south of the Polar Front (approximately 74° N). AW ($S > 35.0$) occupied large areas of the section, and a splitting of the Atlantic inflow into two cores can be seen. The hydrographic conditions below the pycnocline illustrate the main features throughout the year. During winter and spring there is no distinct upper layer.

Arctic Water ($34.3 < S < 34.8$, $T < 0°C$, Loeng (1991)) was not observed at the temperature and
salinity recordings from the current measurements. No effort has been made considering the transformation of AW and the production of locally formed water masses inside the Barents Sea, consequently all outflow was assumed to be recirculated modified AW or modified Arctic water.

The observed current in the area was predominantly barotropic, with only a minor decrease in velocity towards the bottom, as seen at moorings nos. 3 and 4 (Fig. 3). An exception was mooring no. 5, located in the deepest part of the Bear Island Trough, where the speed generally increased towards bottom, and the deepest instrument showed persistent outflow (e.g., Fig. 4). For moorings nos. 3 and 4, monthly mean velocities up to 18–19 cm s\(^{-1}\) were observed. Mooring no. 3 was located in what is expected to be the main inflow area of AW, and the residual current was most of the time directed into the Barents Sea. There were, however, two periods when the residual current was reversed in the whole water column, the last 2 weeks of January 1998 and all of April 1998. The velocities were stronger in April 1998 and also increasing towards the bottom. An outgoing current was also evident at mooring no. 4 in January and April 1998, and this mooring showed an additional strong outflowing episode in December 1997, which can be traced only at the bottom of mooring no. 3.

The monthly mean flow field across the section showed substantial fluctuations (Fig. 4). The conditions in September–November 1997 reflect the most typical situation anticipated, with inflowing AW in the southern part of the section and outflowing water in the northern part, as also

### Table 1
Details of the current meter moorings

<table>
<thead>
<tr>
<th>Mooring</th>
<th>Location</th>
<th>Depth of instrument (m)</th>
<th>Bottom depth (m)</th>
<th>Observation period</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mooring no. 1</td>
<td>71°31.0'N, 19°46.2'E</td>
<td>50</td>
<td>227</td>
<td>28.09.1997–26.08.1998</td>
</tr>
<tr>
<td></td>
<td></td>
<td>125</td>
<td></td>
<td>28.09.1997–03.05.1998</td>
</tr>
<tr>
<td>Mooring no. 2</td>
<td>71°58.9'N, 19°37.5'E</td>
<td>50</td>
<td>309</td>
<td>20.08.1997–27.08.1998</td>
</tr>
<tr>
<td></td>
<td></td>
<td>125</td>
<td></td>
<td>20.08.1997–27.08.1998</td>
</tr>
<tr>
<td></td>
<td></td>
<td>225</td>
<td></td>
<td>20.08.1997–04.03.1998</td>
</tr>
<tr>
<td></td>
<td></td>
<td>294</td>
<td></td>
<td>20.08.1997–04.03.1998</td>
</tr>
<tr>
<td>Mooring no. 3</td>
<td>72°30.7'N, 19°33.2'E</td>
<td>50</td>
<td>388</td>
<td>20.08.1997–27.08.1998</td>
</tr>
<tr>
<td></td>
<td></td>
<td>125</td>
<td></td>
<td>20.08.1997–27.08.1998</td>
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<tr>
<td></td>
<td></td>
<td>225</td>
<td></td>
<td>20.08.1997–27.08.1998</td>
</tr>
<tr>
<td></td>
<td></td>
<td>373</td>
<td></td>
<td>20.08.1997–27.08.1998</td>
</tr>
<tr>
<td>Mooring no. 4</td>
<td>72°59.7'N, 19°33.0'E</td>
<td>50</td>
<td>419</td>
<td>21.08.1997–28.08.1998</td>
</tr>
<tr>
<td></td>
<td></td>
<td>125</td>
<td></td>
<td>21.08.1997–28.08.1998</td>
</tr>
<tr>
<td>Mooring no. 5</td>
<td>73°29.9'N, 19°19.4'E</td>
<td>50</td>
<td>480</td>
<td>21.08.1997–16.10.1997 and 05.03.1998–28.08.1998</td>
</tr>
<tr>
<td></td>
<td></td>
<td>125</td>
<td></td>
<td>21.08.1997–28.08.1998</td>
</tr>
</tbody>
</table>
shown by Blindheim (1989). The inflowing AW was located in one core situated somewhere around 72°30’N, and the maximum velocity was 10 cm s\(^{-1}\). The average transport estimate for e.g., November gave a net flux of about 2 Sv. In December 1997, the inflow core centred at 72°30’N was weaker, but noteworthy features at this point were the cell structures of in and outflow. The strong outflow event in April 1998 showed up with outflowing currents all the way south to 72°N. The average transport estimates from December and April gave net fluxes out of the Barents Sea of approximately 0.5 and 2.5 Sv respectively. The monthly mean flow field gives a general impression of different flow fields between autumn/early winter and summer. This will be discussed later.

Low-pass filtered time series of the volume flux with cut-off periods of 30 and 90 days (Fig. 5) reveal a flux with strong variability on time scales ranging from one to several months. The monthly mean transport was fluctuating between 5.5 Sv into and out of the Barents Sea, and with a standard deviation of 2.5 Sv. The strongest fluctuations of the inflow occurred in late winter and early spring, with both a maximum and minimum value occurring in this period (Fig. 6). During the summer, the inflow was less variable and relatively strong. The recirculation seemed to be more stable at a value of approximately 1 Sv, but with more interruptions of high outflow episodes preferably during winter and early spring. In contrast to what was expected, the flux calculations did not indicate a clear seasonal cycle with the highest Atlantic inflow during winter.

The vertically integrated currents reveal the shifts in transport across the section (Fig. 7) and show horizontal movements of the inflow cores and the recirculation. The core of Atlantic inflow seemed to have a relatively stable location at 72°30’N but was suppressed by the mentioned outflow episodes. On the other hand, there were episodes where the recirculation was suppressed with inflow through the whole section (e.g., July, 1998). Also evident from Fig. 7 are periods where the flow field is dominated by inflow and outflow in narrow bands side-by-side, and with outflow as far south as 71°30’N.

The mean currents for the summer and winter seasons are shown in Fig. 8. The winter season was represented by the period September to mid-March and summer was represented by mid-March to the end of August. During winter, there was a distinct surface-intensified core of inflowing water at about 72°30’N, and outflowing water north of about 73°N. The transport estimates were ca. 2.5 Sv inflow and approximately 1 Sv recirculation. During summer, the upper 200 m were flowing into the Barents Sea for the entire sampled section. Two cores of inflow were evident, with a surface-intensified core located at 73°30’N overlaying deeper outflowing water, and the main core at 72°N. The maximum velocities were higher during winter. Below ~300 m, the summer and winter mean flows were similar. For the summer, a calculation of the volume fluxes gave an inflow of about 3 Sv and a recirculation of about 1 Sv.

The monthly mean velocity recordings at the 19 instruments were correlated against the total volume flux estimates, and the results are shown in Fig. 9a. As the transport is a weighted sum of
the individual velocities from the instruments, some correlation related to the area represented by each RCM was expected. However, the highest correlation was not found where the weighting areas are largest, suggesting the correlation arises from the structure of the current. The highest correlation coefficients for the whole year were 0.83 and 0.72, corresponding to the instruments at mooring no. 3 bottom and mooring no. 4 at 125 m, respectively. To examine if the correlation changed with the season, the correlation coefficients for winter and summer were computed using the same time spans as for mean currents (Fig. 9b and c). The correlations were generally higher during summer than winter. While the three upper instruments at mooring no. 3 showed rather poor correlation during winter, they all had correlation coefficients above 0.85 during the summer. As the correlation between the individual instruments within a mooring was high, the instruments chosen to represent the total flow must be from different moorings. Both seasons then showed maximum correlation at the same instruments, i.e., mooring 3 at bottom and mooring 4 at 125 m depth (Fig. 9b and c). Choosing these two to represent the total mean flow, a linear relationship between velocity

Fig. 3. Time series of 30 days low-pass filtered current vectors from mooring nos. 3 and 4 in the period August 1997 to August 1998. Positive velocity is flow into the Barents Sea.
Fig. 4. Discrete monthly mean current across the BSO. The hatched areas show currents into the Barents Sea and contours are in cm/s.
at these instruments and total flux was assumed. Multiple regression analysis then gave the following expression for total transport $U_{\text{reg}}$:

$$
U_{\text{reg}} = 0.4476 \times 10^6 \text{ m}^3 \text{ s}^{-1} + 0.2824 \times 10^6 \text{ m}^2 u_{3, \text{bottom}} + 0.1855 \times 10^6 \text{ m}^2 u_{4, 125},
$$

where $u_{3, \text{bottom}}$ and $u_{4, 125}$ are the velocities (in m s$^{-1}$) at the instruments at mooring no. 3 bottom and mooring no. 4 at 125 m, respectively. The results are shown in Fig. 10. The regression line transport varied between almost 5 Sv into and ca. 5.5 Sv out of the Barents Sea, with a standard deviation of 2 Sv. The correlation between

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**Fig. 5.** 30 days low-pass filtered (upper panel) and 90 days low-pass filtered (lower panel) volume flux through the BSO as estimated from the current measurements. Positive values represent flow into the Barents Sea. The asterisks represent discrete values of monthly average transport.

**Fig. 6.** 30 days low-pass filtered volume flux through the BSO separated into inflow (positive values) and outflow (negative values).

**Fig. 7.** Time series of transport through the BSO based on 30 days low-pass filtered and vertically integrated currents (left panel), and the total transport in Sv through the section (right panel).
transport from all instruments and regression line transport was 0.90 for the entire period. A splitting of winter and summer seasons gave a correlation coefficient at 0.86 for the winter season and 0.95 for the summer season.

4. Discussion

The current measurements revealed a rather complicated flow pattern and high variability across the BSO. Hydrographic observations presented by Blindheim and Loeng (1981) and Dickson and Blindheim (1984) have given an impression of a rather broad and stable inflow area of AW over the southern slope of the section, while an outflow of Arctic water takes part along the slope south of Bear Island. This impression was supported by 2 months of current measurements reported by Blindheim (1989), which showed a total inflow of 3.1 Sv and an outflow of 1.2 Sv along the northern slope of the section. Our transport estimates based on monthly average currents are in good accordance with Blindheim (1989), although we included neither the Coastal Current nor the Arctic outflow. However, compared to transport values presented in Parsons (1995), Loeng et al. (1997) and Haugan (1999), our values seem to underestimate the recirculation (about 1 Sv) and the inflow during winter. An underestimation of the recirculation was obviously due to the lack of measurements on the slope of the Svalbard Bank. In fact, following Haugan’s (1999) exchange model for mass across the BSO, the total recirculation is about 2 Sv, whereby half of it takes place in the section we are covering and the other half on the slope.

The fact that we had no observations on the slope of the Svalbard Bank (i.e., our section ends at 73°30’N) might also have caused an underestimation of the inflow. AW may cross the BSO as a retrograde slope jet as described in Li (1995) and Li and Mcclimans (1998), and as documented by current measurements during winter presented in Loeng and Sætre (2001) and by model studies (Adlandsvik and Hansen, 1998). According to Li and Mcclimans (1998), the warm core jet is a remotely forced inflow but with occurring current reversals. The core speeds are reduced and strengthened by tail (southwesterly) and head (northeasterly) winds, respectively. The typical Polar Lows (northeasterly winds) dominating during winter can intensify the current by 30% (Li and Mcclimans, 1998). This northern branch of Atlantic inflow is expected to be more variable than the main branch further south, and the ADCP data presented by Haugan (1999) sometimes showed outflow and sometimes inflow on the slope. Total transport associated with the warm core jet is about 0.5–1 Sv (Haugan, 1999; Li and Mcclimans, 1998). An addition of 0.5–1 Sv to our inflow values will give estimates in better accordance with earlier estimates. There is also a possibility that AW may enter the Barents Sea south of our southernmost mooring as a mixture with Coastal Water. Considering the part of the section between 71°15’N and 73°45’N, our estimates are in good accordance with the earlier estimates, and most of them were based on balance budgets and numerical models, not measurements.
The time series of transport revealed three periods with a net outflow through the BSO (December 1997, January–February 1998 and April 1998). A closer examination of the time series of inflow and outflow (Fig. 6) clarifies some differences between the three outflow episodes. The December 1997 episode had a medium strong outflow and still a relatively strong inflow, while in both of the outflow episodes of January–February and April 1998 the inflow was almost none. This means that the December net outflux seems to be dominated by a strong increase in outflux, while in January–February the net outflux was mainly caused by a decrease in influx (Fig. 6). The situation in April was somewhat special as the amount of inflowing water showed a significant decrease and the amount of outflowing water a significant increase creating a combined effect on the total flux. Obviously the inflowing AW was suppressed southwards by the outflowing water (Fig. 4). The temperature in the outflowing water was in the range 1–5°C (not published).

An examination of the time series of current vectors reveals a characteristic vector pattern at mooring nos. 3 and 4 in December 1997 (Fig. 3) with both moorings showing a closed and converging vector pattern. As demonstrated by Foldvik et al. (1988), this is the characteristic for the registration of a cyclonic eddy. A mesoscale eddy
centred somewhere between 72°30'N and 73°N and a period of about 1 month might produce a vector pattern as observed. Such an eddy is consistent with earlier studies in the area. As noted by Blindheim (1989), the current in the shear zone in the Bear Island Trough was turbulent and unstable, and there seemed to be frequently occurring eddies in this area. This was confirmed by surface drifter data (Loeng et al., 1989; Poulain et al., 1996; Loeng and Sætre, 2001).

The presence of mesoscale eddies complicated the task of obtaining reliable transport estimates through the BSO, especially on shorter time scales. The moorings were deployed with a mutual distance of 50 km, which gave a rather coarse resolution considering the width of the flow cells. These features may not have been properly captured by the present grid of moorings, but, according to Loeng and Sætre (2001), most of the eddies in this area have a limited extension and are also rather limited in time. Recent publications have indicated that the exchange through the section may take place as a two-core inflow with stagnant water or outflow in between (Haugan, 1999; Furevik, 2001). A cell structure was already proposed by Kudlo (1961), who based his conclusions on geostrophic calculations. However, these authors based their conclusions on snapshots since no continuous time series were available. As shown by Ingvaldsen et al. (1999), the variability on shorter time scales is substantial. The array of moorings show a two-core system at times (Ingvaldsen et al., 1999), but there may be almost complete reversals of the current through parts of the section during 1–2 days. As hypothesised in Ingvaldsen et al. (1999), the two-core structure seems to be initiated by the strong, highly fluctuating wind typical for late autumn and winter. However, most of these features are averaged out when looking at monthly and longer time scales. The present transport estimates are, therefore, expected to be reliable for monthly to seasonal time scales.

Ádlandsvik and Loeng (1991) found a close relationship between the fluctuations in regional wind and the fluctuations in the Atlantic inflow, with the lowest inflow of water occurring when the atmospheric pressure in the Barents Sea was high. Fig. 11 shows calculated volume flux through BSO compared to surface air pressure at Bear Island. The correlation was good during spring 1998, and the observed outflow episodes in January–February and April 1998 were obviously closely connected to the meteorological conditions. However, there seemed to be a rather poor correlation between air pressure and total volume flux during autumn 1997 and summer 1998. The atmospheric high observed in January–February and April 1998, or the easterly winds associated with it, is likely to give a decrease in the inflow (Fig. 6). However, as demonstrated by Asplin et al. (1998), these episodes were not well simulated by a numerical model designed to capture the variability due to wind, and considering the increased velocities towards bottom at mooring no. 3 in April (Fig. 3), the outflow could not be a purely wind-driven effect. An additional cause was, therefore, needed as reinforcement. Horizontal pressure gradients caused by a higher water level in the eastern Barents Sea and/or the Arctic Ocean are a good candidate. An enhanced water level might be due to accumulation of water by westerly winds and/or an atmospheric low over the area. In fact, the transport estimates show that the period prior to April 1998 was characterised by maximum inflow to the Barents Sea (Fig. 5), and flux estimates from the Fram Strait from the autumn and winter 1997–1998 show a clear maximum in inflow to the Arctic in March 1998 (pers. com. Eberhard Fahrbach). These enhanced inflows might provide the accumulation of water needed for the creation of horizontal pressure gradients. This is not in conflict with the numerical simulations performed by Asplin et al. (1998), as their model domain did not include the Arctic.

The measurements did not reveal the generally accepted seasonal cycle with higher Atlantic inflow during winter, but rather indicated an inflow of 2.5 Sv during winter and 3 Sv during summer. The estimated recirculation was more or less constant throughout the year. A lack of a clear seasonal signal in the recirculating water is consistent with earlier findings (Parsons, 1995). As described in Haugan (1999), this might be due to the recirculating water being fed from the deeper parts of the AW, which has a weaker seasonal variability. This
is in good accordance with Furevik (2001) who analysed hydrographic data from the BSO in the period 1980–1996 and found that the seasonal variability is most energetic in the upper water masses, while fluctuations in the deeper water masses are dominated by interannual variability.

Most of the increased wintertime transport is likely to occur in the upper layers in response to wind forcing. The upper current meters were deployed in 50 m depth, which was below the wind-driven layer, and the transport estimates used the current at 50 m depth for the whole upper 50 m. The wind-driven current in the upper layer, and its seasonal variation, was, therefore, missing in the estimates. This is likely to be one factor causing the underestimation of the transport and lack of seasonal signal, even though the indirect wind effects, such as deeper currents caused by upwelling/downwelling, were included. However, it cannot be the only explanation as a seasonality in the transport was reported by Loeng et al. (1993), who also had the upper current meters in 50 m depth. An alluring, but, at this point highly speculative, thought is that our low inflow-values during winter may be partly due to the fact that we are missing a possible seasonal signal in the inflow that may take place at the slope of the Svalbard Bank and/or a seasonal signal in the mixture of AW and Coastal Water which may flow into the Barents Sea south of our southernmost mooring (e.g., Haugan, 1999).

Although a seasonality in the transport was not found, a seasonal changing flow field may be expected, as this has been reported in other inflow areas of Atlantic water, e.g., in the Faroe Current and the Faroe Shetland Channel (Hansen and Østerhus, 2000), and the Norwegian Sea (Mork and Blindheim, 2000). Such a result showed up nicely in the estimated contour field from the current measurements (Fig. 8). The frequent passing of atmospheric lows during autumn and winter intensified the currents producing a structure with strong lateral velocity-gradients. Due to the wind, the inflow was surface-intensified and relatively strong. During the summer, the weaker easterly winds caused the inflowing area to be much wider and the velocities lower. The southern bottom-intensified inflow is probably remotely forced by the North Atlantic Current, while the surface-intensified current overlaying the outflowing water is likely to be caused by local winds. At this stage it is important to keep in mind that the observations were only a 1-yr time series. The year was close to normal in the sense that the heat content in the core of the inflowing AW was close to the 20-yr average (Aure et al., 1999), although this do not mean that the absent seasonal cycle has to be representative of the average. However, as the currents in this area are strongly controlled by topography (e.g., Loeng, 1991), there is reason to believe that the structure of the current is invariant over interannual time scales.

The correlations between the velocities at the individual instruments and the total transport were higher during summer than winter. This was expected as the seasonal flow field showed stronger lateral velocity gradients during winter, and the variability was stronger during winter. A peculiar result is that the area of maximum correlation did not coincide with the core of the inflow (Figs. 8 and 9). In fact, during winter the correlation showed a local minimum in the core of inflow. This is partly a result of the correlation, which has been made with the total transport. The mean location of the core was rather stable during winter (Fig. 4), and these instruments could, therefore, not capture the outflow properly, outflow which preferably occurred during winter. It might also be linked to the higher temporal variability during winter, although the connection there is not quite clear. Also, the fact that the maximum correlation did not coincide with the core of the current during summer is not fully understood. However, the regression line analysis (Fig. 10) showed that the total monthly transport for this year could be adequately represented with only two of the current meters; the instruments at mooring no. 3 (72°30’N) close to bottom and at mooring no. 4 (73°N) at 125 m depth. This indicates that variability in the inflow of AW was, to some extent, determined by two factors: (1) the temporal variability in the inflow of AW at mooring no. 3 (72°30’N) close to bottom and at mooring no. 4 (73°N) at 125 m depth. This indicates that variability in the inflow of AW was, to some extent, determined by two factors: (1) the temporal variability in the deeper parts (i.e., below 300 m), as the spatial variability at these depths was much less, and (2) the spatial variability in the northern boundary of the inflow, as this showed substantial fluctuation. The first of
these factors had only a weak seasonal signal, while the second had a very clear seasonal signal.

Due to the strong topographic control of the current, an assumption that the spatial structure of the circulation is invariant over interannual time scales seems reasonable. The results from the regression analysis can then be used to design an optimised measurement programme for studying the longer term (e.g., seasonal or longer) variability of the Atlantic inflow to the Barents Sea. For monitoring long-time fluctuations, only two instruments at the determined locations are needed to cover the section between 71°15’N and 73°45’N. Additional instruments may be necessary to monitor on the southern slope of the Svalbard Bank or if variability of shorter time scales is to be investigated. Unfortunately, no support for this statement can be found by applying the regression functions to other similar data, as no such data sets are available.

5. Concluding remarks

The main purpose of this paper was to give a description of the monthly to seasonal variability in the transport of AW through the BSO. The results show that the circulation implied in the hydrographic conditions does not always occur. The flow might occur as a wide Atlantic inflow, an outflow covering the occupied section all the way south to 72°N, or in inflow and recirculation in nearby cells. There were large fluctuations in the inflow, both in time and space.

The volume transport across the section covered by the moorings was calculated and revealed that in certain periods the net flow was from the Barents Sea towards the Norwegian Sea. On a monthly basis, the transport fluctuated over a range of almost 10 Sv. The variability was, in some periods, clearly linked to the local atmospheric pressure field, although additional forces were needed to account for the total flow. In some periods, the correlation between the local surface atmospheric pressure and volume flux was poor. A mesoscale eddy, present for about one month, was observed.

A seasonal signal, consistent with the generally accepted seasonal cycle for the Barents Sea, was not found. However, the current measurements indicated a change in the flow field between summer and winter. During the winter, the frequent passing of atmospheric lows intensified the currents, producing a structure with strong lateral velocity gradients and a distinct, surface intensified, relatively high-velocity core of inflow. During the summer, the winds were weaker and the inflowing area was wider, the horizontal shear and the velocities were lower, and there were two cores of inflow.

The variability in the inflow of AW was, to some extent, determined by two factors: (1) the temporal variability in the deeper parts (i.e., below 300 m), and (2) the spatial variability in the northern boundary of the inflow. This means that for long-time monitoring of the total transport, two current meters located at 72°30’N near bottom and 73°N at 125 m depth are sufficient to cover the section between 71°15’N and 73°45’N. Additional instruments may be necessary to monitor on the southern slope of the Svalbard Bank or if variability of shorter time scales is to be investigated.

Acknowledgements

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The velocity field of the western entrance to the Barents Sea

by

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The velocity field of the western entrance to the Barents Sea

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Abstract

By the use of 4-year long records from moored current meters between 71°30’N and 73°30’N of the western entrance to the Barents Sea, the velocity field of the Atlantic inflow is examined. The mean velocity field shows the Atlantic inflow as a wide core that occupies most of this section, but the general picture is a velocity field that is dominated by large and frequent fluctuations. The key parameter that to a large degree determines the spatial distribution of the velocity field is sea level changes within the section that are induced by the local wind field. The main process is the Ekman transport through its ability to accumulate water thereby creating strong gradients of barotropic pressure and associated currents. Southwesterly winds along the Norwegian coast will in general create wide inflows while northeasterly winds will result in wide outflows, mainly in the northern parts. These flow regimes may be persistent for up to 2-3 weeks, and are related to the relative strength and lateral extension of the Icelandic low and the Arctic high, although the alignment of the local isobars must be considered in order to be able to describe the details of the flow.

1. Introduction

The main inflow of Atlantic water (AW) to the Barents Sea takes place in the Norwegian Atlantic Current (NAC) through the Barents Sea Opening (BSO) (Figure 1). The hydrography along the section between Norway and Bear Island indicates a rather stable inflow of AW in the southern part and an outflow in the north. Such a circulation scheme was confirmed by two months of current measurement that was presented by Blindheim (1989). Recent publications of vessel mounted Acoustic Doppler Current Profiler (ADCP) (Haugan, 1999) found the inflow to take place in two cores with a return flow between them. The
Figure 1. Map of the Barents Sea. The solid and dashed arrows indicate flow of AW and Arctic water respectively. The solid line indicates the section where the current meter moorings were deployed.

results were supported by the analyses of hydrographic measurements (Furevik, 2001), and by the additional analysis of ADCP data (O’Dwyer et al., 2001). Ingvaldsen et al. (2002) analysed a 1-year time series from moored current meters and found that there were large fluctuations in both time and space of the BSO current pattern. They concluded that the flow might occur as 1) a wide Atlantic inflow, 2) a wide outflow, or 3) simultaneous inflow and outflow in distinct cores. AW may possibly also enter as a retrograde slope jet as described by Li and Mcclimans (1998), and as documented by current measurements during winter and presented in Loeng and Sætre (2001), or by model studies (Ådlandsvik and Hansen, 1998).

The variability of the Atlantic inflow is considerable (Loeng et al., 1997; Haugan, 1999; Ingvaldsen et al., 2002). The variability on time scales from day to year is clearly linked to the atmospheric fields (Loeng et al., 1997), and although the net effect of the atmospheric forcing on the transport is modest, transport fluctuations due to the local wind can have the same magnitude as the mean transport (Ådlandsvik and Loeng, 1991). Variability in the barotropic currents that are generated by sea level gradients is also expected to be important on a short time scales (Mcclimans et al., 1999).

In this paper we use 4-year long records from an array of moored current meters across the southern part of the BSO, in order to link the velocity fields to atmospheric forcing and sea level gradients. Data and analysis techniques are described in section 2 while section 3 gives a brief description of a few days of fluctuations in the velocity field, and examines the forcing behind them. A statistical approach to explaining the spatial distribution within the velocity field is provided in section 4, and the relationship between the flow field and the atmospheric and sea level fields are investigated. A brief summary and conclusions are given at the end.
2. Data and analysis

The data material consists of velocity and temperature measurements from moored Aanderaa current meters RCM7 from August 1997 to August 2001. The moorings were deployed between 71°30’N and 73°30’N, and mainly covered the Atlantic inflow, but not the coastal inflow and the flows on the slope of The Svalbard Bank (Figure 2). The number of moorings deployed, the distance between the moorings, and the number of instruments on each mooring, varied throughout the period (Table 1). Data were recorded every 20 minutes. To fill gaps in the time series due to missing single instruments (Table 1, lower panel), simple linear interpolation of the velocities from the instrument above and/or below was performed. This is an adequate method since the velocities are mostly barotropic (e.g. Haugan, 1999). When moorings 2b and 3b were not deployed, interpolated values from the two surrounding moorings were used. For the period January –August 1999 mooring No. 1 was missing, and values from mooring No. 2 were extrapolated to represent this mooring. In addition, time filtering was performed with an order 4 Butterworth lowpass filter (Roberts and Roberts, 1978).

The presence of mesoscale eddies complicates the task of obtaining reliable results when dealing with current meter moorings, however most of the mesoscale activity in the area had time scales that were less than 14 days (Ingvaldsen et al., 2003). The data material was therefore filtered by removing fluctuations with periods that were less than 14 days, before Empirical Orthogonal Functions (EOF) analysis was performed. Due to the lack in the number of moorings (Table 1, lower panel), the periods January-February 1999 and September 1999-August 2000 were not included in the EOF analysis. As the EOF analysis requires data from all points in time, the separate time series were concatenated before analysis, and the resulting time series were 34 months long. After the analysis, the principal components (the time series of each EOF) were split into their actual time. Due to the

![Figure 2. Salinity in August 1998 and location of all the moorings and instruments. Table 1 gives information of the moorings and instruments deployed at a given time.](image)
concatenation, a thorough analysis of the principal components (as e.g. spectral analysis) is of no significance.

The atmospheric sea level pressure and wind field in 10 m above sea level were taken from the Norwegian Meteorological Institute (Met.no) hindcast archive (updated from Eide et al., 1985).

The sea level height was obtained from numerical simulations that were carried out with NORWECOM (Skogen and Søiland, 1998), which is a 3D, primitive equation, sigma-coordinate coastal ocean model based on the Princeton Ocean Model (Blumberg and Mellor, 1987). The model domain covered the Nordic Seas, the Barents Sea and the Arctic Ocean, and was discretized on a 20 km horizontal polar stereographic grid. The model forcing included initial and boundary conditions from The Norwegian Meteorological Institute-Institute of Marine Research diagnostic climatology (Engedahl, et al., 1998), realistic

<table>
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<tr>
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<th>Mean of Velocity</th>
<th>Kinetic energy</th>
<th>Temperature [°C]</th>
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Table 1. Mooring details and statistics for the current meters (upper part), and time series of functioning time for each current meter (lower part).
meteorological forcing from the NCEP-NCAR reanalysis project, and monthly mean river runoff and tidal forcing. The model has been validated for the Nordic and Barents Sea (Asplin et al., 1998), and the results showed that the produced velocity fields in general are qualitatively good, but the magnitude of currents are underestimated. Based on this, we expect the sea level height pattern to be realistic but with possibly underestimated gradients.

3. Short-time fluctuations in the velocity field

3.1 A basic description

To illustrate the large variability in the velocity fields of the BSO, the period January 21-28, 1998 is investigated (Figure 3). Large fluctuations in the cross-sectional flow occurred, and from one day to another, the current was reversed in large parts of the section. On January 21st, the inflow took place in two relatively strong cores with a small outflow close to the bottom in between them. Two days later, the flow had changed direction towards the Norwegian Sea all the way south to 72°N. The next few days there were at first a wide core of inflow with an area of stagnant water in the middle, then an in and outflow in cores, simultaneously. The maximum velocities were about 20 cm s\(^{-1}\) in both directions. The mean velocity field indicates that the core of Atlantic inflow is located in the area near 72°30′N (Figure 4), but the daily velocity fields reveal that the location of the inflow core is not stable at this location as there may also be an outflow there. This reflects a fluctuating nature of the location of the core, as it moves laterally in and out of the domain of the current meter. The temperature recordings from the mooring at 72°30′N reveal this feature very clearly (Figure 5). It is striking how the instrument near the bottom showed much higher temperature

![Figure 3](image-url)

Figure 3. Vertical sections showing daily mean cross-sectional currents (cm s\(^{-1}\)) at four dates in January 1998. Grey areas show eastward flow (i.e. flow into the Barents Sea).
variability than the instruments higher in the water column. At the bottom, the temperature changed more than 3°C between successive days.

The high variability of the flow can also be seen from the kinetic energy computations (Table 1). The kinetic energy of the mean flow is quite low and an order of magnitude less than the kinetic energy of the flow fluctuations, which in general indicates weak mean flows. Mooring No. 4 and the three upper instruments on mooring No. 5 have the highest magnitude of kinetic energy of the flow fluctuations, but closer to zero kinetic energy of the mean flow. These moorings were deployed in the area of the northern boundary of the AW inflow, and the spatial variability in this boundary is substantial (Ingvaldsen et al., 2002). The instrument in the deepest parts of the Bear Island Trough differs from the remainder by having a significantly higher kinetic energy of the mean flow and a relatively low magnitude of kinetic energy of the flow fluctuations. This instrument clearly captures the deep outflows from the Barents Sea.

### 3.2 Wind and sea level as driving mechanisms

The daily mean wind field and the modelled sea level from the 7-day period in January 1998, reveal high variability in both time and space (Figure 6). On January 21st, there were strong southwesterly winds across the section. Due to the Ekman transport being to the right of the wind direction, this will give an inflow in most of the section, which is consistent with the velocity field observed that day (Figure 3a). The situation on January 23rd was the opposite: strong northerly winds resulting in a westward Ekman transport, and an outflow
Figure 6. Daily mean wind and sea level (cm) for the same days as shown in Figure 3.

were observed across most of the section on that day (Figures 6 and 3b). On January 25th, the winds were weak southwesterlies in the southern part and weak southeasterlies in the northern parts (Figure 6c). The associated Ekman transport would be an inflow in the south and an outflow in north and otherwise low velocities, which are highly consistent with the observed velocity field (Figure 3c). The situation on January 27th, with strong flow in narrow cores seems to be forced by winds with an easterly component (Figure 3 and 6d), although the actual physical explanation is not clear.

In summary, there seems to be consistency between the direct Ekman transport and the observed currents, with an inflow that is created by southerly winds and an outflow by northerly winds. However, due to the barotropic nature of the currents, this does not fully explain the observed velocity fields and will be investigated in the next section.

4. A statistical approach for the explanation of the velocity fields

To investigate the relation between the spatial distribution of the velocity fields and the wind, the cross-sectional velocities were decomposed by EOF analysis. The advantage of EOF analysis is that it provides a compact description of the spatial and temporal variability of data series in terms of orthogonal functions, or statistical “modes”. Usually, most of the variance of a spatially distributed data series is in the first few modes whose pattern may then be linked to possible dynamic mechanisms. The mean of the time series were removed prior
to the EOF analysis and the time series were normalized by the standard deviation. An alternative EOF analysis without extraction of the mean prior to the analysis is performed in Appendix 1. The EOF analysis was not sensitive to the length of the time series, or to the inclusion of moorings 2b and 3b (i.e. whether the analysis were performed five or seven moorings, not shown). The leading EOF accounts for 34% of the variance in the flow field, and is, according to the criterion of North et al. (1982), well separated from the second order EOF explaining 20%. However, as the second and third orders EOF are not well separated, only the leading EOF is presented here. The leading EOF has a uni-pole pattern with a core near 72°30’N (Figure 7). Due to both a positive and negative principal component, the EOF1 also represents situations when the current anomalies are reversed. Thus, the negative state of EOF1 represents a situation with an anomalously wide outflow.

4.1 Physical interpretation of the EOF

EOF analysis is a purely statistical method, and to justify the physical interpretation of the leading EOF, we used a simple and indirect method. First, characteristics in the wind field anomalies and the sea level anomalies1 across the section were identified for the positive and negative state of EOF1. Then inferences on the driving mechanisms were made based on the observed characteristics. To identify the characteristics for the positive state, the times where the principal component exceed a specified value and had a positive gradient (to capture the build-up period) were found. The specified values were chosen subjectively so that only the strongest appearances of the EOF were included (the dotted horizontal line in Figure 7b), and the mean wind and sea level fields for these times were constructed. To examine whether the pressure gradient associated with the sea level could create the velocity fields pictured by the

Figure 7. (a) Leading EOF for cross-sectional velocity, and (b) the associated principal component. Grey areas show eastward flow. The dotted horizontal lines in the principal component show the specified values from which the characteristics for the EOF have been taken. Note that the mean velocity was subtracted prior to the analysis, and the time series were normalized by the standard deviation.

1 As the EOF represents flow anomalies, wind and sea level anomalies are the appropriate parameters to investigate.
EOF, the barotropic geostrophic velocity based on the mean sea level gradient was found. The same procedure was applied to find the characteristics of the negative state of the EOF, at the times when the principal component was below a specified value and further decreasing. As for the velocity time series, the time series of the atmospheric fields and sea level fields were filtered to remove fluctuations of periods that were less than 14 days prior to the calculation of the mean.

The mean (anomalous) wind field during the build-up of the positive state of EOF1 (Figure 8a) is southwesterlies in the southern part of the section and southerlies in the northern part. This will create a direct Ekman transport eastwards on most of the section. More importantly, the southwesterly winds in the southern part of the section will give an Ekman transport towards the Norwegian coast, while the southerly winds in the northern part will give an easterly Ekman transport. This causes a higher water level in the south than in the north, reflected in the mean sea level anomaly at times for dominating positive EOF1 (Figure 8b). The geostrophic currents associated with this sea level reproduce mainly the pattern of the positive state of EOF1 (Figure 8c), although not completely. The mismatch may be related to an underestimation of the sea level magnitude (see section 2), and the fact that the numerical model includes more processes, and not only the isolated effect of the Ekman transport.

The wide outflow as described by the negative state of EOF1 (Figure 9a) is characterised by strong northerly winds east of Spitsbergen, northeasterly winds at the section and southeasterly winds southwest of Spitsbergen. This gives an accumulation of water in the north, consistent with the mean sea level anomaly at times for dominating negative EOF1 (Figure 9b). The geostrophic currents associated with the sea level anomaly resemble the pattern of the negative state of the EOF1 (Figure 9c), although with a mismatch as for the positive EOF1.

In summary, the statistical analysis clearly indicates that the anomalous velocity field across the BSO to a large degree is forced by sea level changes within the section. The sea level changes are created by an accumulation of water induced by the local wind field through the Ekman transport. This is consistent with the results of Appendix 1, where an alternative EOF analysis is performed to identify whether the velocity field can adopt certain preferred structures, and to identify the forcing that is responsible for these structures.
Figure 8. Mean characteristics associated with strong occurrences of the positive EOF1. (a) Horizontal fields of anomalous wind, where the associated Ekman transport also is sketched. (b) Vertical view of modelled sea level anomaly (cm) across the BSO, and (c) calculated geostrophic velocities (cms⁻¹) associated with the sea level in (b). Hatched areas show eastward flow. For description of the selection criteria see the text and Figure 7b.

Figure 9. Mean characteristics associated with strong occurrences of the negative EOF1. (a) Horizontal fields of anomalous wind, where the associated Ekman transport also is sketched. (b) Vertical view of modelled sea level anomaly (cm) across the BSO, and (c) calculated geostrophic velocities (cms⁻¹) associated with the sea level in (b). Hatched areas show eastward flow. For description of the selection criteria see the text and Figure 7b.

4.2 Relations between the statistical analysis and the observed flow

The results from section 4.1 showed that when southerly winds prevail (southwesterlies along the Norwegian coast), the anomalous flow is dominated by a wide core of inflow occupying the entire section. Alternatively, when northerly winds prevail, the anomalous flow is dominated by an outflow that occupies the entire section. To relate this to the observed flow, it must be considered as to how the anomalous flow field (Figure 7a) modifies the mean field (Figure 4). When the positive state of EOF1 is dominating, the Atlantic inflow in general is increased, especially within the core at 72°30’N. This is seen clearly in the reconstructed EOF1 at the strongest positive occurrences (Figure 10a). Alternatively, when the negative state of EOF1 is dominating, the outflow in the northern part is increased and the inflow in the southern part is decreased. The largest decrease is found near 72°30’N. When
the flow of the negative EOF1 is strong enough to reverse the current in this area, as it does for the strongest negative occurrences (Figure 10b), it will represent the observed situations of a wide outflow occupying large parts of the section with an inflow only in the southern parts. An example of this may be seen in Figure 3b. A similar situation existed in April 1998 as described in Ingvaldsen et al. (2002). This is consistent with the principal component being negative in most of April 1998 (Figure 7b). Actually, the principal component of EOF1 is also negative in March-April 1999 and 2001, which is consistent with Ingvaldsen et al. (2003) who found a pronounced minimum in the Atlantic inflow (or could even be the outflow) in spring, due to an annual event of northerly winds.

4.3 Relations between the flow and the regional atmospheric fields

Although a relation between the velocity field and the local wind field is established, the wide inflows and outflows are at times persistent for several weeks. Such a persistency can only be caused by a persistent pattern in the regional wind field. The mean regional atmospheric pressure and wind fields (Figure 11a) for the positive EOF1 shows strong similarities with the mean winter situation with the well-known Icelandic low that stretches as a trough into the Nordic Seas. This suggests that persistent wide inflows are a manifestation of the northward extension of the Icelandic low. The results are supported by the alternative EOF analysis performed in Appendix 1.

The mean regional atmospheric pressure and the wind field for the negative EOF1 (Figure 11b) reveals that the northerly winds that causes the wide outflows, are associated with a strong high-pressure area in the Arctic that stretches towards Spitsbergen. This suggests that persistent wide outflows are forced from the Arctic Ocean (also consistent with Appendix 1). The northerly winds move large amounts of water from the Arctic to the northern and eastern Barents Sea, therefore pushing the entire water masses of the northern and eastern Barents
Sea southwards. The supply of water into the Barents Sea from the Arctic Ocean and a southward movement of the northern and eastern Barents Sea can explain why the large outflows can be present for 2-3 weeks without loosing too much water through the BSO during this period. It is also consistent with the necessity for a response time between wind and sea level for such a large-scale movement. This could explain why the temperature in the BSO does not decrease significantly during periods of large outflows (e.g. April 1998 in Figure 5), as the outflowing water will come from the southern parts of the Barents Sea and not from the northern parts as it would if it was purely wind-driven.

The previous results indicate that the wide inflows and outflows across the BSO are related to the relative strength of the Icelandic low and the Arctic high. However, the correlation coefficient between the principal component of EOF1 and the regional atmospheric pressure field shows that the strongest correlation (-0.38) is found in an area west of the BSO (Figure 11c). This probably reflects that this area is a good index for the alignment of the isobars across the BSO, as the local alignment of the isobars decides the local wind field in the BSO. The alignment of the isobars depends on the relative strength of the Icelandic low and the Arctic high, on their lateral extent (i.e. stretching into the Nordic Seas or towards Spitsbergen), and on local processes that are not captured by these largescale systems. The result should be interpreted as that a strong Icelandic low, stretching into the Nordic Seas, creates the persistent southerly winds necessary for a persistent wide inflow (and vice versa for the outflow), but the details of the velocity field is determined by local effects that are not captured by the regional field.
5. **Summary and conclusions**

By using data from moored current meters between 71°30’N and 73°30’N in a section across the BSO from August 1997-August 2001, the velocity fields in the AW inflow to the Barents Sea have been examined. In general, the velocity field is dominated by frequent and large fluctuations, and nearly complete reversals of the currents through parts of the section may occur within 1-2 days.

The velocity field was decomposed by EOF-analysis and related to the wind field, the sea level height as obtained by numerical modelling, and atmospheric sea level pressure. The key parameter that to a large degree determines the spatial structure of the velocity field is sea level changes that are induced by the local wind field. The main process is the Ekman transport through its ability to accumulate water within the section thereby creating barotropic pressure gradients and associated geostrophic currents. This process is enhanced by the topographic constraints with the Norwegian coast in the southern part.

The mean velocity field shows that the Atlantic inflow takes place as a wide core occupying most of the section between 71°30’N and 73°30’N (except in the deeper parts of the Bear Island Trough). Southwesterly winds along the Norwegian coast modulates this velocity field by accumulating surface water in the southern part, and at the same time moves the surface waters of the northern areas out of the section. The result is a pressure gradient, which creates a geostrophic velocity field that enhances the wide inflow. When the predominant wind direction is northeasterly, the Ekman transport will produce a higher water level in the north than in the south. The associated pressure gradient drives a relatively strong outflow in the northern part, often in combination with a weaker inflow in the south. The flow regimes may be persistent for several weeks, and are related to the relative strength and lateral extension of the Icelandic low and the Arctic high, but local processes that are not captured by the regional fields determines the details of the flow.

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We are grateful to colleagues for valuable comments on the manuscript, and to Peter Haugan for helpful discussions. The results presented have partly been obtained through funding from the European Union MAST III VEINS programme and MAIA project. In addition funding has come from the Research Council of Norway through the projects “Variation in space and time of cod and other gadoids: the effect of climate and density dependence on population dynamics” and NOClim.
Appendix 1. An alternative EOF analysis

As an alternative approach, EOF analysis was performed on the time series without removing the mean prior to analysis, but they were normalized by the standard deviation. EOF analysis that is performed this way still decomposes the data series into orthogonal functions; however, to assign a percentage of explained variance, or to investigate the separation according to North et al. (1982), cannot be achieved. The statistical significance of the results cannot therefore be investigated, which makes the method “uneasy” in a statistical sense. However, it serves as a more intuitive approach to investigate if the velocity field adopt certain preferred structures, and as the results are in accordance with the results of the traditional EOF analysis, we find it instructive to include.

The leading EOF (MEOF1) closely resembles the mean with one wide core of an inflow, and outflow in the deeper parts of the channel (Figure 12a). This is consistent with the principal component being positive most of the time. Even though the EOFs were based on lowpass filtered data with a cut-off period of 14 days, some similarities with the daily current fields in January 1998 can be seen (Figure 3); especially the flow fields observed on January

![Figure 12](image-url)
21st and 25th. The MEOF2 shows a pattern with an inflow south of 72°30’N a stronger outflow in the north when positive (Figure 12a). This means that the wide outflows are separated into its own mode with this method. Due to both positive and negative principal component, this mode also represents situations when the currents are reversed. The velocity field of January 23rd, 1998 mainly resembles the MEOF2 (Figure 3b). Out of interest, we also presents the third order EOF that represents situations when the flow takes place in several distinct, narrow cores of inflow and outflow (Figure 12a). One core of inflow is located in the area between 72°45’N and 73°N, the other somewhere south of 72°N, although the flow might be reversed. The exact positions of the cores are obviously subject to uncertainties due to the insufficient lateral resolution of current meters. The flow reversals that are evident from this mode may be caused by very small changes in the flow field, as the cores moved in and out of the points occupied by the current meters. The multi-core velocity field is in agreement with Haugan (1999) who found two cores of Atlantic inflows sited close to the mentioned positions, with a return flow or stagnant water in between. The flow situation in January 27th, 1998 was an example of the negative state of this mode (Figure 3), while the flow field on January 21st, 1998 resembles the positive state (together with the MEOF1). There are several examples of all three structures in the field observations, both only on the daily values, but also on time scales up to monthly means.

**Physical interpretation of the EOFs**

To seek physical interpretation of the EOFs, the same method as used in section 4.1 was performed, except that ordinary wind and atmospheric pressure were used instead of anomalies. The mean wind and pressure fields during the build-up of the MEOF1 (Figure 13a) closely resemble the situation found for the positive state of EOF1 (Figure 11a). The sea level and geostrophic velocities literally fit better using this method than the traditional. With the principal component being positive most of the time, this suggests that the wide inflow is a manifestation of the remote forcing of the NAC in combination with the Icelandic low. Note that due to the NAC, the water level will always be higher in the southern part. The strength of the Icelandic low will therefore only modulate the situation.

Also for MEOF2, the mean wind and pressure fields during the build-up (Figure 13b) closely resembles the situation found for the negative state of EOF1 (Figure 11b), and also
Figure 13. Mean characteristics associated with strong occurrences of the (a) MEOF1, (b) MEOF2 and (c) MEOF3. Horizontal fields of wind and atmospheric pressure (top). Sketch of the wind and the associated Ekman transport (no 2 from top). Vertical view of sea level (cm) across the BSO (no 3 from top), and the calculated geostrophic velocities (cms⁻¹) associated with the sea level (bottom). Grey areas show eastward flow. For description of the selection criteria see the text and Figure 12b.

Here give the sea level anomaly² and geostrophic velocities a better fit than by using the traditional method. The velocity field as described by the MEOF3 is characterised by strong easterly winds in the Barents Sea (Figure 13c). The strong cross-sectional wind results in moderate along-section sea level gradients, probably due to the shear of the wind field. The sea level anomaly present when this mode is dominating is capable of producing a

² As the velocity field at any given point in time is a sum of all the EOFs, the higher order EOFs should be regarded as anomalous flow. Therefore, we will here use sea level anomaly and not ordinary sea level as for MEOF1.
geostrophic velocity field much like the MEOF3, suggesting that also this velocity pattern is forced by the gradients of sea level height. The results are consistent with the results from section 3 where it was found that the flow in narrow cores seems to be associated with winds with an easterly component.

The analysis has confirmed the results from the traditional EOF analysis in that the velocity field across the BSO is forced mainly by sea level changes induced by the wind. In combination with the fact that the MEOF3 may be both positive and negative, this means that the cores of inflow should not be viewed as locked to the positions given in Figure 12a. Instead, they should be interpreted as having a location that is determined by the wind field, but it is possible that the wind field has preferred horizontal distributions that will create the cores of inflow and outflow at fixed locations. Moreover, due to the interaction with the background flow of the NAC, the distinctness of the cores of inflow and outflow generated by the sea level will depend on its location. Our investigation cannot fully describe these aspects.

The velocity field pictured by the MEOF3 has some similarities with an eddy, suggesting that some of the variability captured by that mode may be related to eddy-activity. However, due to the length of the time series, sporadic eddies will only contribute significantly in the EOF-analysis if they appeared at fixed locations. If this is the situation, they would have had to be generated within the area or transported along a certain depth contour. Such eddies might be generated in the shear zone between the inflowing and the outflowing waters of the Bear Island Trough, but it is unlikely that these can persist long enough as not to be removed by the time filtering. This is supported by Loeng and Sætre (2001) who found that the eddies in the area seem to be rather limited in both time and space.

References
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The seasonal cycle in the Atlantic transport to the Barents Sea

by

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The seasonal cycle in the Atlantic transport to the Barents Sea

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Abstract

The seasonal cycle in the Atlantic inflow to the Barents Sea is investigated by using 4-year long records from moored current meters, additional moorings of shorter duration, and hydrographic measurements. Except from the first year of measurements, a higher transport during winter than summer is observed, and is related to barotropic currents that are forced by sea level changes within the section induced by a shear in the cross-section wind stress. It is possible that variations in the remotely forced Norwegian Atlantic Current also contribute. The seasonal variation within the section is not uniform, and the strongest seasonal signal is found in the middle and northern parts. In the area near 71°30’N the alluring result arises that the wind (both speed and direction) changes substantially from winter to summer, without changes of the net Ekman transport. The water level gradient, and the associated barotropic current, therefore has no seasonal variation in this area. However, there is a seasonal signal in the front between the Atlantic and Coastal Waters near the Norwegian Coast. This is due to coastal downwelling during winter that is forced by strong southwesterly winds. The mean transport of Atlantic Water is estimated to 1.7 Sv (1 Sv=10^6 m^3 s^-1) during winter and 1.3 Sv during summer, but there is a pronounced minimum in Atlantic inflow (or even outflow) in spring due to an annual event of northerly winds.

1. Introduction

The Barents Sea serves as a thoroughfare for Atlantic Water (AW) to the Arctic Ocean, where it affects both the Eurasian Basin (Rudels et al., 1994) and the Canadian Basin (Jones et al., 1995). Recent investigations suggest that this throughflow can take part in a process
where water distilled at the surface of the Arctic by freezing ends up at mid-depth in the same ocean (Aagaard and Woodgate, 2001). The process starts with an increase in the export of ice from the Arctic Ocean and into the Barents Sea. The ice melts due to contact with the AW, whereby the melt water is entrained in the Barents Sea throughflow and subsequently sinks into the Arctic Ocean. Woodgate et al. (2001) ascribed an observed cooling and freshening of the Atlantic layer in the Arctic to this process. Thus, an understanding of the variability of the Atlantic inflow to the Barents Sea, and a quantification of the volume transports, is not only important for the conditions in the Barents Sea itself, but also for the understanding of the climatic state of a wider region.

The main inflow of AW to the Barents Sea takes place in the Norwegian Atlantic Current (NAC) through the Barents Sea Opening (BSO) (Figure 1). Along the Norwegian Coast, the Norwegian Coastal Current (NCC) brings relatively fresh and warm water into the Barents Sea. During winter, the NCC is deep and narrow, while during summer it is wide and shallow (Sætre and Ljøen, 1971). North of the NAC, the Bear Island Current takes relatively colder and fresher water of Arctic origin out of the Barents Sea.

Several previous studies have given estimates of the volume flux across the BSO, but moored current measurements in the area have been rare. Prior to our measurements the only transports based on moored current meters were from a 2-month time series (Blindheim, 1989). All other published estimations of volume fluxes have been based on vessel-mounted Acoustic Doppler Profiler (Haugan, 1999; O’Dwyer et al., 2001), as well as budget considerations (Loeng et al., 1997; Haugan, 1999), numerical modelling (e.g. Parsons, 1995), and geostrophic calculations of the baroclinic currents (numerous Russian studies). A

Figure 1. Map of the Barents Sea and the main current systems. NAC: the Norwegian Atlantic Current, NCC: the Norwegian Coastal Current, and BIC: the Bear Island Current. The solid line shows the section where the current meter moorings were deployed.
seasonal cycle with a higher inflow during winter than summer has been inferred based on the general wind field, on the geostrophic calculations, and on current measurements in the eastern Barents Sea (Loeng et al., 1993).

In this paper we use the 4-year long time series from an array of moored current meters in the main Atlantic inflow in the BSO (Figure 2). We also consider two additional moorings on the slope south of Bear Island, hydrographic sections, 10-m wind field, and sea level height as obtained from a numerical model. The investigation addresses the seasonal variation in the Atlantic inflow, and emphasizes these issues: (1) description and quantification of the exchanges of AW between the Norwegian and the Barents Seas, and (2) identification of the forcing responsible for the variations in the Atlantic inflow. The paper is organized as follows: In section 2 the data material is presented, while section 3 deals with description and quantification of the AW transport. In section 4 the results presented in section 3 are discussed and the forcing responsible for the seasonal variations are investigated. The results are summarized and concluded in section 5. The 4-year time series from the moored current meters is the same as used by Ingvaldsen et al. (2003a), but then with the focus on giving a description of the observed velocity field.

2. Description of the data

The 4-year records consist of velocity and temperature measurements from moored Aanderaa Recording Current Meter 7 (RCM7) deployed between 71°30’N and 73°30’N in the period from August 1997 to August 2001 (Figure 2 and Table 1). The number of moorings deployed, and the number of instruments attached to each mooring, varied over the period. Data were recorded every 20 minutes. To fill gaps in the time series due to missing single instruments (Table 1, lower panel), simple linear interpolation of the velocities from the

Figure 2. Salinity in August 1998 and location of all the moorings and instruments. Black boxes denote instruments included in the 4-year time series, grey boxes instruments from the short time moorings. Information of the moorings and instruments in the 4-year time series are given in Table 1.
instrument above and/or below was performed. This is an adequate method since the velocities are mostly barotropic (e.g. Blindheim, 1989). When moorings No 2b and 3b were not deployed, interpolated values from the 2 surrounding moorings were used. For the period January –August 1999 mooring No 1 was missing, and the values from mooring No 2 were extrapolated to represent this mooring. Time filtering was performed with an order 4 Butterworth lowpass filter (Roberts and Roberts, 1978).

The presence of mesoscale eddies (with time scales of a several days) can seriously alias the result when calculating volume fluxes from moored current meters, but as shown in Appendix 1 the influence of mesoscale activity in this area is relatively small when investigating fortnightly or longer time scales. Based on this, the time series were filtered by removing fluctuations with periods less than 14 days prior to further analysis.

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<td>225</td>
<td>5.0</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>1082</td>
<td>464</td>
<td>7.8</td>
</tr>
</tbody>
</table>

Table 1. Mooring details and statistics for current meters in the 4-year time series (upper panel), and functioning time chart for each current meter (lower panel).
Transports between 71°15’N and 73°45’N were calculated by assigning a rectangle surrounding each current meter. The volume fluxes were calculated from the cross-sectional current component (which is the east-west component as the section is aligned north-south). For the period January-February 1999 there were only a few instruments working (Table 1). Fortunately the bottom RCM7 at mooring No 3 worked for the entire period, as earlier this location was found to be the representative for the total flow through the section (Ingvaldsen et al., 2002). A linear relationship between the velocities from the available instruments and the total flux was assumed, and regression analysis for the period of August 1997-December 1998 was used to express the total transport and to estimate the flux from January-February 1999. The same method was used for the periods with only 3 moorings (September 1999-August 2000, see Table 1). Due to the barotropic structure of the currents, the expression for total volume flux was based on the 2 deepest instruments on the 3 moorings. The regression analysis was performed for the period of September 2000-August 2001. When error estimates are assigned to the volume fluxes, the calculations were performed as described in Appendix 2.

The current meter material is extended with data from 2 moorings that were deployed on the shelf south of Svalbard (Figure 2). One mooring (No 6) was deployed at 73°45’N in the period of August-October 1997 and had RCM7 current meters deployed in 50 and 125 m depths. The other mooring (No 6b) was deployed at 74°00’N in the period of August 1998-March 1999 and had RCM7s deployed in 50 and 106 m depths. The settings and processing of the data were identical to the 4-year series of the current measurements. A small data set sampled after the 4-year period is also included in this paper. For the period of August 2001-March 2002 a mooring with an Aanderaa profiling Doppler Current Meter 12 (DCM12) and a RCM7 was deployed at 73°30’N. The DCM12 was deployed in a depth of 45 m and measured velocity in 5 bins between 5 and 41 m every second hour, while the RCM7 was deployed in a depth of 60 m.

The section between Norway and Bear Island is a standard hydrographic section and is sampled 6 times a year by the Institute of Marine Research. Here, salinity and temperature from the period 1997-2001 is used.

The wind field at 10 m above sea level were taken from the Norwegian Meteorological Institute (Met.no) hindcast archive (updated from Eide et al., 1985), while the sea level height was obtained from numerical simulations that were carried out by the use of NORWECOM (Skogen and Søiland, 1998). NORWECOM is a 3D, primitive equation, sigma-coordinate coastal ocean model based on the Princeton Ocean Model (Blumberg and Mellor, 1987).
model domain covered the Nordic Seas, the Barents Sea and the Arctic Ocean. The model was discretized on a 20 km horizontal polar stereographic grid, and forcing included initial and boundary conditions from The Norwegian Meteorological Institute-Institute of Marine Research diagnostic climatology (Engedahl, et al., 1998), realistic meteorological forcing from the NCEP-NCAR reanalysis project, and monthly mean river runoff and tidal forcing. The model has been validated for the Nordic and Barents Sea (Asplin et al., 1998), and the results showed that the produced velocity fields in general are qualitatively good, but the magnitudes are underestimated. Based on this we expect the sea level height to be realistic but with possible underestimated gradients.

Throughout the paper *winter* will denote the months of December through March, and *summer* June through August.

3. Results

The mean salinity distribution in January and August (Figure 3) reveal that the moorings generally captured the main part of the Atlantic inflow (identified by the 35 psu isoline), although some parts are missing, especially in the south during summer and in the upper 50 m during winter. Motivated by the available data sets (basically the location of the moorings), the section was divided into 4 regions (Figure 3). The seasonal signal of each region is investigated and an attempt to quantify the transport of AW is also made.

3.1 Region 1 - the area occupied by the moorings

The moorings covered the area between 71°15’N and 73°45’N and from 50 m depth to the bottom. The quality of the salinity data from the RCM7s was poor, and an identification of the different water masses in the flow by temperature and salinity could not be made. Thus,
AW was defined by temperature as water above 3°C. Concerning the seasonal cycle, three features of the monthly mean transports are striking (Figure 4). The first is a remarkable transport minimum in spring (March-April). The second is the apparent lack of a clear seasonal cycle with higher inflow during the winter than in the summer. However, when separated into different moorings it is evident that there are several contradicting seasonal cycles within the section (Figure 4b-d). In the northern part of the section there is a seasonal cycle with an outflow of AW during winter and an inflow during summer. The middle part has a weak seasonal cycle with a higher inflow during winter, while the southern part has a very stable inflow without any seasonal signal at all. Due to its high stability, the southernmost mooring has the highest mean inflow of AW, although it in general has the weakest velocities. The third feature is that the first year of the measurements showed high inflow compared to the others (Figure 4a), and substantial differences appear if the seasonal

![Figure 4. Monthly mean transport of AW estimated from (a) all moorings, and (b-d) the contribution from 3 separate moorings. Positive transport is into the Barents Sea (i.e. eastwards).](image)

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1 This should be adequate as temperature is the characteristic parameter separating AW from the outflowing water masses in the north. Salinity is the key for separating the AW from the Coastal water, but as the Coastal water occurs mostly in the upper 50 m in the area of the moorings, it will not seriously affect our estimates.
Table 2. Mean transport from current measurements separated in the first year and the last 3 years.

<table>
<thead>
<tr>
<th>Atlantic Water</th>
<th>Winter</th>
<th>Summer</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mooring area, Sep 97-Aug 98</td>
<td>1.7 ± 4.1</td>
<td>2.1 ± 3.0</td>
</tr>
<tr>
<td>Mooring area, Sep 98-Aug 01</td>
<td>1.5 ± 1.9</td>
<td>1.1 ± 1.8</td>
</tr>
</tbody>
</table>

mean transports are calculated separately for the first and for the last three years (Table 2). The first year had an abnormally high inflow especially during summer compared to the remainder of the time series, and during the same year the climate of the Barents Sea changed from a cold state to a warm state (e.g. Figure 5 in Ingvaldsen et al., 2003b). Based on this we propose that the first year of the measurements are not likely to be representative for the normal (or usually present) annual cycle. The most realistic estimate of the AW transport between 71°15’N and 73°45’N and 50 m to bottom for a normal year is therefore 1.5 Sv (1 Sv=10⁶ m³ s⁻¹) during winter and 1.1 Sv during summer. The annual mean is 1.2 ± 1.1 Sv.

3.2 Region 2 - the upper 50 m above the moorings

The mean salinity distributions during summer and winter show a clear seasonal signal in this area (Figure 3). During winter the AW is present all the way to surface, while during summer the stratification is strong and by the end of summer no AW is present in the upper 50 m. To attempt to estimate the transport of AW we investigate whether the direct wind drift in the upper meters gives a significant contribution. The instantaneous and vertically integrated Ekman transport has a cross-sectional component $U_E$ given by

$$U_E = \frac{1}{\rho_0 f} \tau^y$$

where $\rho_0$ (1026 kgm⁻³) is the density of seawater and $f$ (1.39·10⁻⁴ s⁻¹) is the Coriolis parameter. $\tau^y$, the along-sectional wind stress, was obtained from the along-sectional wind velocity $v$ by

$$\tau^y = \rho_{air} C_d |v| v$$

where $\rho_{air}$ (1.3 kgm⁻³) is the density of air and $C_d$ is the drag coefficient as given by Large and Pond (1981). The wind data are provided on a regular grid, and the grid points lying closest to the section were identified. To determine the integrated transport a rectangle surrounding each point was assigned. Although the Ekman transport on a daily time scale
varied between 1.5 Sv into the Barents Sea and 1.6 Sv out, the variability was high and the resulting mean values were close to zero. The mean values were 0.02 Sv during winter and 0.00 during summer with a long-term mean of 0.01, i.e. negligible considering our accuracy.

Another method for estimating the transports in the upper 50 m is to extrapolate the velocity from the current meter in 50 m depth to the surface. To investigate the reliability of this method, transport estimates from a profiling current meter DCM12 and a RCM7 deployed in 60 m were compared. Volume transports in the upper 50 m were estimated by 5 bins from the DCM12, and by letting the RCM7 represent the top 50 m (Figure 5a). The results show surprisingly small differences, both for amplitude and phase (the correlation

Figure 5. Calculation of volume fluxes in the 4 regions shown in Figure 3. (a) Volume flux between the surface and 50 m estimated by both a profiling current meter DCM12 and by extrapolating a RCM7 deployed in 60 m. (b) The area occupied by AW for each repetition of the standard section in the period 1997-2001 south of 71°15’N, and (c) north of 73°45’N. (d) Volume flux north of 73°45’N estimated from the mooring at 73°45’N in August-October 1997 (dotted line), and from the mooring at 74°00’N in August 1998-February 1999 (solid line).
coefficient is 0.91). These measurements were done mainly during fall and winter, when the stratification is weak and the winds are strong. During summer the strong stratification may prevent such a vertical homogeneity, and it seems as if the fit between the DCM12 and the RCM7 is weaker in early fall when the upper waters are still stratified (Figure 5a). However, the good comparison with the DCM12 justifies that a simple extrapolation of the observations from 50 m depth to the surface is adequate. The transport in the upper 50 m was therefore calculated from the instruments in 50 m depth and the resulting mean total transports were 0.25 Sv both for summer and winter. To quantify how much of the flux is AW, we noted that AW covers an area of about 75% of the box during winter (Figure 3), thus giving an AW inflow of (using January as a representative for winter) 0.18–0.2 Sv. Taking August as a representative for summer the transport of AW is ~0.0 during summer (Figure 3).

### 3.3 Region 3 - the area south of 71°15’N

The mean salinity distribution shows that the Atlantic inflow is pushed down-slope away from the Norwegian coast during the winter compared to the summer (Figure 3). The calculated area occupied by AW (defined by S>35 and T>3°C) reveals this feature clearly with a much smaller amount of AW present in late winter and spring than in summer (Figure 5b).

Considering the transports in this area Blindheim (1989) presented a 2-month time series of current meters between Norway and Bear Island, and a rough estimate of the volume transport south of 71°15’N based on his results give a flux of 0.7 Sv into the Barents Sea. Approximately 0.5 Sv of this was the NCC, leaving 0.2 Sv of Atlantic water inflow. The measurements were from the period of September-October, i.e. likely to be representative for the summer situation due to the stratification. The current measurements of Blindheim (1989) indicated also a homogenous flow between 71°30’N and the NCC. A rather crude estimate of the transport of AW is therefore to multiply the area occupied by AW with a constant velocity. Our southernmost mooring revealed seasonally stable currents (Figure 4d), with a long-term mean velocity of about 3.5 cm s^{-1}. This gives Atlantic transports of 0.08 Sv during winter and 0.13 Sv during summer. If the constant velocity is increased to 5 cm s^{-1}, the transports become ~0.13 Sv during winter and ~0.21 Sv during summer. These estimates are in good agreement with the number obtained by Blindheim (1989). As it is likely that this method underestimates the AW flux, our “best guess” is that the Atlantic transport in this area is ~0.1 Sv during winter and ~0.2 Sv during summer.
3.4 Region 4 - the area north of 73°45’N

The area north of 73°45’N occupied by AW was computed as in the previous section. In contradiction to what was found in the south, these results shows that the Atlantic influence is stronger during the winter and spring than during the summer (Figure 5c). Thus, the Atlantic domain is displaced northwards during winter. As was shown earlier, the mooring at 73°30’N had a clear seasonal cycle with inflow during the summer and outflow during the winter (Figure 4b), indicating that the AW found north of 73°45’N during the winter is recirculating water.

Two moorings have been deployed on the slope south of Bear Island, one at 73°45’N, the other at 74°00’N (Figure 2). Unfortunately, none were deployed during summer. By allowing each of the moorings to represent the total area north of 73°45’N, two time series of (total) transport are obtained (Figure 5d). Although most of the time the flow was out of the Barents Sea at these locations, persistent inflow for almost one month occurred. The mean Atlantic transports were -0.1 Sv for both moorings.

As another approach the area occupied by AW was multiplied by a constant velocity as was done in the previous section. The mean velocity of the 2 upper instruments at mooring no 5 were taken as representative for the velocity in the AW also north of 73°45’N. The velocities were about 2 cms⁻¹ westward during winter and 1.5 cms⁻¹ eastward during summer. The calculated volume fluxes of AW become -0.1 Sv during winter and ~0.0 Sv inflow during summer, i.e. in reasonable agreement with the two moorings on the slope.

4. Discussion

The results from the previous sections are summarized in Figure 6. The results reveal an Atlantic transport of 1.7 Sv during the winter and 1.3 Sv during the summer. This gives a seasonal variation of 25%, although an abnormally strong flow in some years may differ substantially from these values. The annual mean is 1.5 Sv. The volume flux in the area occupied by the moorings is by far the largest contribution to the summarized fluxes, which

![Figure 6. Schematics of the distribution of AW within the BSO and the estimated volume fluxes in the different regions.](image-url)
means that the moorings captured most of the Atlantic inflow. This is probably realistic, although it’s importance may be exaggerated due to the rather crude calculation methods used outside the mooring domain. However, independent of the quantification of the actual transports, the 4-year records from the current meters and the hydrography revealed a seasonal cycle in both the current velocities and the distribution of AW within the section. The main findings are that there is a pronounced minimum in Atlantic inflow or even an outflow in the spring (Figure 4), there is in general higher Atlantic inflow during the winter than the summer (Table 2 and Figure 6), the seasonal variation within the section is not uniform (Figure 4), and the Atlantic domain is displaced northwards during the winter compared to the summer (Figure 5 and 6).

4.1 The spring minimum

The current measurements revealed a pronounced minimum in spring (Figure 4). At times it was even an outflow, being persistent as it shows on the monthly mean transports. From a separate investigation of the velocity fields Ingvaldsen et al. (2003a) found that when the wind in the Barents Sea is northeasterly, the water level rises in the northern part of the BSO thereby forcing a relatively strong outflow in the northern part while there is an inflow in the south. This clearly suggests that the spring minimum is connected to a seasonal shift in the along-sectional wind field. The winds in the section (Figure 7) are part of a pronounced spring phenomenon with a periodic wind change from the general southerly direction to a northerly. A similar shift in the winds are occurring in most of the Barents Sea (not shown), and the large outflows are probably a consequence of large-scale horizontal pressure gradients caused by pushing the entire northern and eastern Barents Sea southwards (Ingvaldsen et al., 2003a).

Additional factors may also contribute to the spring outflows. One of these is the effect of the vertical stratification; as the stratification is almost absent in spring, the effect of the wind will reach further down in the water column. Furthermore, the formation of high-density water during the winter as described by Midttun (1985) may take part of the observed outflow. Although the outflows have too high temperatures to be a draining of this winter water, the ongoing process of production of winter water will generate baroclinic horizontal pressure gradients towards the west that will increase during winter. However, due to the barotropic nature of the currents, the sea level should be expected to be important. In other words, the large spring outflows are caused by an annual spring event of northerly winds, but there are probably several (not fully understood) contributing processes.
Figure 7. Monthly mean (a) along-sectional (north-south) and (b) cross-sectional (east-west) wind speed at locations near the moorings. Southerly and westerly winds are positive (i.e. positive directions are northwards and eastwards).

4.2 Higher inflow during winter than summer

The higher inflow during winter is partly due to the presence of AW in the top 50 m, but is mainly due to higher current velocities below the direct wind driven layer (Figure 6). There are (at least) two aspects that can be responsible for the higher inflow in the deeper part during the winter; the NAC is stronger due to upstream conditions (Orvik et al. (2001) found that the eastern branch of the NAC in the Norwegian Sea has a systematic annual cycle which doubles the inflow during winter), and/or there may be local effects in the BSO that amplifies the current.

As shown by Ingvaldsen et al. (2003a) the main process that drives the currents of the Atlantic inflow are sea level changes within the section that are induced by Ekman transport. The Ekman transport is limited to the upper layers and has a moderate effect considering the total flow (e.g. section 3.2), but has a significant effect through its ability to accumulate water. The pressure gradient associated with the sea level gradient then drives barotropic currents. The Ekman transport affects the sea level in two ways. Near a coastline a uniform alongshore wind will increase or decrease the water level directly on the coast, while away from the boundaries, the curl of a non-uniform wind field will result in local accumulation or removal of surface waters. The along-sectional Ekman transport $V_E$ is given by
The cross-sectional wind stress $\tau^x$ is obtained from the cross-sectional wind velocity $u$ by $\tau^x = \rho_0 f \frac{\partial \eta}{\partial y}$ (the other variables are defined as in section 3.2). The sea level at any given point depends on the amount of water transported into and out of the area. Therefore, the net Ekman transport in a given area was calculated as the difference between the two surrounding points in the north and south. Along the Norwegian coast, there is no Ekman transport and $V_E$ was defined as zero, i.e. the calculated values include both the effect of the shear of the along-sectional wind stress and the presence of the coast. During the winter the wind field is clearly sheared and will decrease the water level along all of the section except near the coastline (Figure 8), while during the summer it is only in the southern part that there is an effect. Note that due to the remotely forced NAC the water level in general is higher towards the Norwegian coast, i.e. the wind field will only modify an existing gradient. The wind field, Ekman transports and the consequences for the sea level are outlined in Figure 9. This reveals that due to the local wind field the water level gradient is much steeper during winter than that of summer, and the associated geostrophic currents are therefore stronger during winter. The cross-sectional geostrophic current $U_{geo}$ is given by

$$U_{geo} = -\frac{g}{f} \frac{\partial \eta}{\partial y}$$

where $\eta$ is the water level, $g (9.82 \text{ m/s}^2)$ is the acceleration due to gravity and $f$ is the Coriolis parameter. The water level due to the local wind field is difficult to quantify as it depends on the along-sectional and the cross-sectional Ekman transports as well as Ekman pumping. However, the numerical model finds the mean total water level during summer and winter, and the associated geostrophic currents reveal the deduced seasonal variability with higher velocities during the winter than in summer (Figure 10). The calculated currents include both the effect of the local wind and the eventual stronger gradient due to a stronger NAC during winter. However, a stronger NAC during the winter should, at least in principle, give a uniform gradient along the section, while the difference between the geostrophic currents during winter and summer reveal non-uniformity that resembles the difference in the water level gradients deduced from the shear of the wind field (Figure 9c and 10, see also section 4.3). This suggests that the higher inflow during the winter is mostly due to the shear of the cross-sectional wind, possibly amplified by a seasonal cycle in the remotely forced NAC. The volume flux may be calculated by simply interpolating the barotropic geostrophic
Figure 8. Net along-sectional Ekman transport at the section between Norway and Bear Island. The calculation was performed so that the net transport is positive when the gain is larger than the lose, i.e. a negative transport is associated with a decrease in sea level.

current (Figure 10) all the way to bottom, giving 4.6 Sv during the winter and 3.1 Sv during the summer. Considering that the strong outflow in the deepest part of the channel is not included, and neither is the baroclinic flow contribution, and this is total transport, not only transport of AW, these numbers are reasonable. The seasonal variation is 33%, which is higher than the 25% found in the beginning of this section. This may be related to an opposing seasonal signal in the deep outflow or in the other contributions that are not included.
4.3 The non-uniform seasonal variability within the section

The shear in the cross-sectional wind induces a non-uniform seasonal cycle within the section, which resembles that in the geostrophic velocities (Figure 9c and 10). The largest winter to summer differences in water level gradient due to the shear of the wind field is found near the Norwegian coast (Figure 9c), and this area also shows the largest winter to summer difference in geostrophic velocity (Figure 10). Near 71°30’N alternatively, there is almost no seasonal signal at all, which is also in accordance with the current measurement (Figure 4). This shows the importance of the shear of the wind field, as both the wind speed and direction in this area changes substantially from winter to summer (Figure 9), but the net Ekman transport does not (Figure 8 and 9). Therefore, the water level has the same gradient during both seasons (Figure 9), and the associated geostrophic current has no seasonal signal (Figure 10). North of 72°00’N the difference between winter and summer increases northward towards 73°30’N both for water level gradient (Figure 9c) and geostrophic velocity (Figure 10). A seasonal signal was also found in the current measurements in this area (Figure 4). The geostrophic velocities do no reproduce the outflow in the northern part of the section (Figure 10), reflecting that this probably not is driven by local sea level gradients.
4.4 Seasonality in the distribution

The hydrographic data showed that the Atlantic domain in general is displaced northwards during the winter compared to that of the summer (Figure 5 and 6). In the southern parts the Atlantic water is found further down-slope during the winter (Figure 3b). This is likely a result of coastal downwelling. The southwesterly winds along the Norwegian coast during the winter forces the surface waters towards the coast (Figure 9), and the water level along the Norwegian coast rise (consistent with the stronger geostrophic currents in Figure 10). By continuity, the water masses of the deeper layer move offshore. This is consistent with the observed extent of the NCC, as being narrow and deep during the winter (illustrated by the 34.7 isoline in Figure 3). During summer the winds are reversed but weaker, the Ekman transport is offshore and the NCC is wide and shallow (Figure 3). The numerical model predicts a reduced water level at the Norwegian coast compared to the winter (Figure 10). A seasonal lateral oscillation of the NCC forced by the Ekman transport of along-coast winds has earlier been reported off southern Norway (Sætre et al., 1998).

The results also showed a seasonal signal in the lateral distribution of AW in the northern parts (Figure 3). The reason for this remains unknown, although it may be linked to stronger re-circulation during the winter due to stronger winds.

5. Summary and conclusions

The seasonal cycle in the Atlantic inflow is examined using 4-year long records from moored current meters between 71°30'N and 73°30'N in the BSO, additional moorings north of 73°30'N, hydrographic measurements, wind, and sea level from a numerical model.

We find a seasonal cycle in both the transports and the distribution of AW within the section. The main findings are:
1. A generally higher Atlantic inflow during the winter than the summer.
   This is connected to sea level changes within the section induced by the shear of the cross-sectional wind field. The local wind field during winter has a shear, and through along-sectional Ekman transport, a relatively steep water level gradient is created. The associated barotropic currents give a higher Atlantic inflow during the winter than the summer. The local wind field is probably capable of producing the observed seasonal variation, although it is possible that the remotely forced NAC also contributes.

2. The seasonal variation within the section is not uniform.
   This is connected to a non-uniform shear of the cross-sectional wind field. The middle and northern part of the section experiences a wind field with stronger shear during the winter than the summer, thereby giving a strong seasonal signal in these areas. Near 71°30’N on the other hand, the alluring result arises that both wind speed and direction change substantially from winter to summer, but the net Ekman transport does not. The water level gradient, and the associated geostrophic current, therefore has no seasonal signal in this area.

3. The Atlantic domain is displaced northwards during the winter compared to the summer.
   At least in the south, this is connected to coastal downwelling during the winter. The southwesterly winds gives an Ekman transport onshore, the water level along the Norwegian coast is raised, and coastal downwelling moves AW down-slope away from the coastline.

4. A pronounced minimum in Atlantic inflow or even an outflow in the spring.
   The situation is forced by a annual spring event of northerly (along-sectional) winds. 
   The transport of AW across the section is quantified and reveals an Atlantic flow of 1.7 Sv during the winter and 1.3 Sv during the summer. This gives a seasonal variation of 25%, although abnormally strong flow in some years may differ substantially from these values. The annual mean transport of AW is found to be 1.5 Sv.

Acknowledgements
We would like to thank Peter Haugan for fruitful discussions and valuable comments on the manuscript. The results presented have partly been obtained through funding from the European Union MAST III VEINS programme and the MAIA project. In addition funding has come from the Research Council of Norway through the projects “Variation in space and time of cod and other gadoids: the effect of climate and density dependence on population dynamics” and NOClim.
Appendix 1. Mesoscale activity

The presence of mesoscale eddies can seriously alias the result when estimating volume fluxes from moored current meters. Although eddies in the area are limited in both time and space (Loeng and Sætre, 2001), eddies with periods of about 1 month have been observed (Ingvaldsen et al., 2002). The presence of eddies should show up as large differences between transport series calculated with different spatial resolution. To investigate the reliability of our transport series, transports for the period of September 2000-August 2001 were calculated based on 7 moorings and based on 5 moorings after replacing mooring No 2b and 3b with the mean of the adjacent moorings (see Figure 2). The resolution in the middle part was 25 and 50 km respectively. Note that prior to the calculations, the velocity data were filtered with a 14-days lowpass filter. The results show surprisingly small differences (Figure 11). Although there are indications that mooring No 2b at times captured a weak return flow or an area of stagnant water which the nearby moorings fail to capture (not shown), a doubling of the resolution does not generally alter the transport values. The largest differences appear during periods of local maximum and minimum transport when the higher resolution sometimes gives higher transports and sometimes lower. The mean of the difference between the two transport series is 0.0, and the error of the mean is ± 0.1 Sv. The transport fluctuations are somewhat sensitive to the increased resolution, but the correlation coefficient is 0.93, i.e. 5 moorings deployed with a mutual distance of 50 km capture 86% of the variability. The relatively good accordance between these two transports series indicates that the influence of mesoscale eddies probably is small when considering fortnightly or longer time scales, i.e. they will not seriously affect the transport series.

![Figure 11. Transport of AW across the section as estimated from the current measurements using 5 and 7 moorings. Positive transport is into the Barents Sea (i.e. eastward).](image-url)
Appendix 2. Error estimates on the mean transports

The transport estimates are subject to errors in the measured velocities, and of the calculation method. The errors in the current measurements are small due to calibration and a large number of individual samples.

The uncertainties in the calculation methods are partly due to the variability in time. The presence of mesoscale eddies can create large errors in the transports, therefore fortnightly means were calculated before the transports were estimated (Appendix 1). To quantify the error associated with fortnightly means due to the eddy noise, the standard error of the mean (S.E.) was calculated. This is given by

\[ S.E. = \frac{S(x,z)}{\sqrt{n}} \]

where \( S(x,z) \) is the empirical standard deviation at a given instrument and \( n \) is the number of degrees of freedom (or number of independent time steps). Individual samples are not independent, and \( n \) was found for each instrument by the autocorrelation function. S.E. was calculated for each instrument, and weighted by the area represented by the instruments to give an error estimate for the velocity means.

Errors also arise due to the extrapolation of each current meter to represent boxes with uniform velocity. Assuming a monotonic structure between the instruments, the error will be small when two adjacent moorings sample almost the same velocity and large when the adjacent moorings have a large velocity difference. Using the same methods as Fahrbach et al. (2001) for current measurements in the Fram Strait, the error was expressed by the mean of the velocity difference between the adjacent points. The error of the velocity difference mean was then found as for the error of the velocity means. It turned out that this error was about the same size as the error of the velocity means.

The largest errors occur if the velocity structure between adjacent moorings is not monotonic, but as shown in Appendix 1 the mean of the difference between the transport series calculated for 5 and for 7 moorings is 0.0, and the error of the mean is ± 0.1 Sv. This is much less than the errors obtained by the velocity difference between the adjacent moorings. The presented errors, which are a combination of the errors obtained from the mean velocities and the box method, may therefore be considered worst-case error estimates for the transports. Note that the error estimates hereafter presented are errors for the mean; they should not be interpreted as an error for the actual transport at a given time.

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Climate variability in the Barents Sea during the 20\textsuperscript{th} century with focus on the 1990s

by

R. Ingvaldsen, H. Loeng, G. Ottersen, and B. Ådlandsvik

ICES Marine Science Symposium, in press.
Climate variability in the Barents Sea during the 20th century with focus on the 1990s

Randi Ingvaldsen, Harald Loeng, Geir Ottersen and Bjørn Ådlandsvik

Abstract

Time series of temperature in three sections representative of Atlantic Water in the Barents Sea reveal that the climate in the region has both long-term and short-term quasi-regular periods. In comparison with other decades during the 20th century, the 1990s were colder than both the 1930s and 1950s. The 1990s started out warm, followed by a short relatively cold period in 1996-1998. During the last years of the decade there was a gradual build-up towards higher temperatures, with very high anomalies during late autumn and early winter.

The NAO has, through regional and local effects, a significant influence on the Barents Sea on decadal time scales and during extreme NAO events. Still, local atmospheric forcing not captured by the NAO index seems to be dominating for the distribution of the water masses within the area. The local pressure field appears to change the relative strength of the two branches going respectively northeast and east, thereby having a significant effect on the local climate. The local pressure distribution not captured by the NAO index also has some influence on the total inflow to the Barents Sea.

Keywords: Barents Sea, climate variability, decadal, North Atlantic Oscillation, 1990s

Rising waters of the Barents Sea during the 20th century with focus on the 1990s

Randi Ingvaldsen, Harald Loeng, Geir Ottersen and Bjørn Ådlandsvik

Abstract

Time series of temperature in three sections representative of Atlantic Water in the Barents Sea reveal that the climate in the region has both long-term and short-term quasi-regular periods. In comparison with other decades during the 20th century, the 1990s were colder than both the 1930s and 1950s. The 1990s started out warm, followed by a short relatively cold period in 1996-1998. During the last years of the decade there was a gradual build-up towards higher temperatures, with very high anomalies during late autumn and early winter.

The NAO has, through regional and local effects, a significant influence on the Barents Sea on decadal time scales and during extreme NAO events. Still, local atmospheric forcing not captured by the NAO index seems to be dominating for the distribution of the water masses within the area. The local pressure field appears to change the relative strength of the two branches going respectively northeast and east, thereby having a significant effect on the local climate. The local pressure distribution not captured by the NAO index also has some influence on the total inflow to the Barents Sea.

Keywords: Barents Sea, climate variability, decadal, North Atlantic Oscillation, 1990s

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1. Introduction

The first broad analysis of the hydrography, currents and climate variability in the Barents Sea was made by Helland-Hansen and Nansen (1909), who suggested that the climate variations in this region probably were of an advective nature. However, not all changes in the Barents Sea are traceable upstream, and advection in the ocean does not explain a major part of the variability in the temperature conditions (Ottersen et al., 2000). Ådlandsvik and Loeng (1991) demonstrated that the inflow of Atlantic Water (AW) to the Barents Sea is determined, by a large degree, by local atmospheric forcing. Their conceptual feedback model stated that cyclonic airflow in the Barents Sea would increase the Atlantic inflow and thereby increase the temperature. The higher temperatures would then maintain a low air pressure. The Barents Sea temperature conditions are further influenced by local winter cooling and ice processes. Formation of water of high density during winter, followed by draining from the sea in bottom currents, may be an important temperature-regulatory factor (Midttun, 1985). The activity in building up dense bottom water may vary from one year to another, followed by variations in the outflow with corresponding changes in the inflow. In the recent years, a relation between the Barents Sea climate and the North Atlantic Oscillation (NAO) has been proposed by Grotefendt et al. (1998), Dickson et al. (2000), Ottersen and Stenseth (2001) and Ottersen et al. (this volume).

The Barents Sea is a pathway for AW to the Arctic Ocean, and the transformation of AW when it passes through the Barents Sea is important for the ventilation of the Arctic Ocean (e.g. Aagaard and Woodgate, 2001). Observations have shown that the Barents Sea provides intermediate water reaching down to a depth of 1200 m in the Arctic Ocean (Rudels et al., 1994; Schauer et al., 1997). Barents Sea climate is therefore also important for large-scale climate developments linked to the Arctic Ocean.

The present study deals with climatic variability in the Barents Sea, emphasizing the Atlantic domain in the 1990s. Temperature data from three hydrographic sections and a geographically distributed data set were analysed. The relation between the temperature variability in the different sections and the horizontal distribution of temperature were investigated. Finally, causes for the observed climatic variability are suggested.

2. Material and methods

Location of the stations and sections from which data was obtained are shown in Fig. 1. This includes time series of temperature in the section Fugløya-Bear Island (FB) in the western entrance of the Barents Sea and the section Vardø-N along 31°13’E. The time series go back to 1953 in the Vardø-N section, while in the section FB the regular observations started in 1964. In the beginning, regular observations were only taken once a year, in late August or early September. Since 1977, regular observations have been carried out more frequently. At the FB section six times a year, and at the Vardø-N section four times each year. The data presented here are average temperatures for 50-200 m in the part of the section where the main Atlantic inflow takes place as identified by salinity (Blindheim and Loeng, 1981). For the section FB this means between 71°30’N - 73°30’N and for the section Vardø-N the area between 72°15’N - 74°15’N was used. Data from the Russian records on the Kola section along 33°30’E are also included. Data from 1921 to the present day are presented as monthly mean values calculated for the depth interval between 0-200 m in the area between 70°30’N - 72°30’N. Historical data were taken from Bochkov (1982) and Tereshchenko (1996), while data from the recent years were provided by PINRO, Murmansk. The meteorological data were supplied by the Norwegian Meteorological Institute. The surface air

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**Figure 1.** The Barents Sea. The solid and dashed arrows indicate flow of AW and Arctic Water respectively. The sections Fugløya-Bear Island (FB), Vardø-N (VN), and Kola (K) are shown, and the area where average temperatures were calculated is indicated. The location of the meteorological stations are symbolised by . Letters symbolize Tromsø Bank (T), the Nordkapp Bank (N), Finger Canyon (F) and the Murman Current (M).
temperatures are from their stations at Torsvåg, Tromsø, Bear Island and Hopen (Fig. 1), while the winter mean sea level pressures (SLP) are from their hindcast archive (Eide et al., 1985).

With basis in data sampled during international fish surveys carried out every year, horizontal fields of temperature from 1970 to 2000 were constructed. Data from the area south of 75°N were sampled by 2 Norwegian and 2 USSR research vessels between 20 August and 10 September each year. Data from the area north of 75°N were sampled by 2-3 Norwegian and USSR vessels in the period 10 September to 10 October. Gaps in the vertical were filled by simple linear interpolation separately for each station. For each separate vertical level a 2D algorithm (Taylor, 1976), combining Laplace and cubic spline interpolation, was applied. The grid distance was 20x20 km. No extrapolation was performed, nor smoothing except for the implicit effect of the interpolation.

The long-term means for the sections were calculated for the period 1961-1990, following the international meteorological mean period, and all anomalies were calculated based on this mean period. For the horizontal temperature fields the long term mean was prepared for the shorter period 1970-2000 due to the availability of data.

3. Results
The Barents Sea climate alternates between warm and cold periods, which have both long-term and short-term quasi-regular fluctuations (Fig. 2). The temperature fluctuation seems similar in all the sections. During the 1950s and 1960s the variability was high and fluctuations with periods about 3-5 years were dominating. Prior to and after these decades, longer periods were more pronounced. Comparisons of the August temperature anomalies decade by decade reveal the same for the oceanographic sections and meteorological stations: the 1990s were warm, but the 1930s and 1950s were even warmer (Fig. 3).

In order to investigate how representative the sections are for the rest of the Barents Sea, the time series of temperature in each point in the horizontal field were correlated with mean temperature of the section time series averaged over the same months (Fig. 4). In general all sections show relatively high correlation with values of 0.6 or higher throughout the Atlantic domain. An exception is in a narrow north-south band near 22°E, where the maps indicate two patches of poor correlation (Fig. 4). The northern patch seems to be located in the area of Finger Canyon, where the local topography modifies the circulation and the frontal position (Parsons et al., 1996). The southern patch is confined to the 400 m deep trough between Tromsø Bank and Nordkapp Bank (Fig. 1) where there is a more or less permanent eddy (Loeng et al., 1989). All maps also indicate that the temperature in the

![Figure 2. Time series of mean August temperature for the hydrographic sections.](image)

![Figure 3. Decadal temperature anomalies for (a) the hydrographic sections, and (b) the air temperatures at the meteorological stations.](image)
Temperature anomalies at 100 m show that the 1990s started out warm both in the Atlantic and most of the Arctic domain (Fig. 6). Thereafter there was a gradual temperature decrease in the western and increase in the eastern parts in 1991 and 1992. The year of 1993 was significantly colder than the previous years, followed by a short warming period towards 1995. The temperature drop in 1996 was sudden and simultaneous in most of the Barents Sea (Fig. 5 and 6). A recovery towards higher temperatures was evident in the western parts in 1998, apparently progressing eastwards towards the central parts of the Barents Sea in 1999.

4. Discussion

The correlation maps between the sections and horizontal fields of temperature (Fig. 4) demonstrated that all the three sections are fairly representative of the Atlantic domain in the Barents Sea. However, due to their more northerly position and thereby distance from the Norwegian Coast, the mean values from FB and Vardø-N are most representative for the northern branch of inflow to the Barents Sea, i.e. for the AW going north into Hopen Trench (Fig. 4). The Kola section on the other hand is located in the Murman Current, and is more representative for the portion of AW continuing east towards Novaya Zemlya.

4.1 Decadal variability and the effect of the NAO in the Barents Sea

In the 1990s the temperatures in the Barents Sea were well above the long term mean (Fig. 3). However, both the 1930s and the 1950s were warmer, indicating that the warming of the last decade may very well be related to natural variability rather than anthropogenic effects. However, since the 1960s there has been a general increase in both oceanic and atmospheric temperatures in the Barents Sea (Fig. 3). During this period the NAO winter index (Hurrell, 1995) changed from its most negative phase in the 1960s to its most positive phase in the late 1980s/early 1990s. The Nordic Seas/Arctic response of this shift is well documented (e.g. Dickson et al., 2000), although not fully understood. The shift has been accomplished by, among several other changes,
Figure 6. Horizontal fields of August-October mean temperature anomalies in 100 meters depth in the 1990s.
1) a significant increase in winter cyclone activity for the region north of 60°N as a whole, with locally significant increases over the central Arctic Ocean, the Barents Sea and the Kara Sea (Serreze et al., 1997; Rogers, 1997), 2) a decrease in late winter ice-extent in the Arctic and eastern Nordic Seas (Johannessen et al., 1999; Vinje, 2001), 3) an increase in the annual volume flux of ice from the Fram Strait (Dickson et al., 2000; Vinje, 2001), 4) an increase in the amount of Atlantic inflow to the Arctic Ocean (Grotefendt et al., 1998; Dickson et al., 2000).

A positive NAO index will result in at least three (obviously connected) oceanic responses in the Barents Sea, reinforcing each other and causing both higher volume flux and higher temperature of the inflowing water. The first response is connected to the direct effect of the increasingly anomalous southerly winds during high NAO (see Fig. 7a). Secondly the increase in winter storms penetrating the Barents Sea during positive NAO (Serreze et al., 1997; Rogers, 1997) will, according to the conceptual feedback model of Ådlandsvik and Loeng (1991), give higher Atlantic inflow to the Barents Sea. The third aspect is connected to the branching of the Norwegian Atlantic Current (NAC) before entering the Barents Sea. Blindheim et al., (2000) found that a high NAO index corresponds to a narrowing of the NAC towards the Norwegian Coast. The narrowing will result in a reduction in heat loss, i.e. higher temperatures in the inflowing water also in the Barents Sea (Furevik, 2001). In addition, we might speculate in that a narrower current forced towards the Norwegian Coast may result in that larger portion going into the Barents Sea, although this has not been documented.

For a negative NAO winter index, there may on the other hand be opposing forces. Associated with a negative NAO winter index is a secondary low-pressure centre over the Barents Sea (e.g. Fig. 5 in Serreze et al., 1997; Fig. 2 in Dickson et al., 2000). While the weaker winds associated with a low NAO will decrease the inflow, there is a possibility for the weak low to enhance it if located over the Barents Sea, thereby reducing the effect of the low NAO. This may be one of the reasons why the correlation between NAO and Barents Sea

![Figure 7. Mean SLP for December-March. (a) Typical situation with high NAO winter index 1990, (b) 1993, (c) 1996, and (d) 1999. Year is decided by the time of the January.](image-url)
temperature are best during positive NAO phases as observed by Dickson et al. (2000) and Ottersen et al. (this volume).

4.2 Variability within the 1990s
The high NAO in the late 1980s/early 1990s resulted in high temperatures in the entire Barents Sea. The temperature decreased throughout the Barents Sea in 1993 (Fig. 5). Also in the Norwegian Sea the temperatures decreased after the first years of the 1990s (e.g. Fig. 5 in Blindheim et al., 2000 and the cold anomaly in south-western parts in 1992 in Fig. 6), indicating that the water masses entering the Barents Sea in 1993 were colder than the long-term mean. In the Atlantic domain the cooling was most pronounced along the Norwegian-Russian coast and in the eastern parts (Fig. 5 and 6). This cooling was more likely an effect of the local atmospheric fields than the NAO, as the NAO index was only slightly lower this year compared to 1992 and 1994 (e.g. ICES 2000). To seek explanations to the observed variations, the mean SLP for the previous winter were examined. The motivation for this was based on the following: during winter the vertical stratification in the Atlantic domain is very weak due to vertical convection, and the water column may be homogenous down to 2-300 m. Although local cooling and freezing is very important in the northern and eastern parts, winter temperatures in the areas occupied by Atlantic water masses vary in parallel with the variations in the inflowing AW even at the surface (Midttun, 1990). The stratification of the upper layers starts in April-May, and during the summer the warming of the water column reaches down to 50-60 m due to turbulent mixing. The AW is therefore effectively isolated from the surface during summer, and the local conditions that determine the surface temperatures (Midttun, 1990) have small influence on the AW. This, in combination with a higher inflow in winter than summer, is probably the reason for why the temperature level for the rest of the year has been found to be “set” by the hydrographic winter, only adjustments to this level taking place later (Ottersen et al., 2000). Izhevskii (1964) found the temperature level for the coming year to be “set” as early as in December. The winter mean SLP shown in Fig. 7 b reveals that the major difference between 1990 and 1993 was the presence of an elliptical low-pressure centre over the Barents Sea and northeastern Norwegian Sea. The east-west extent of the low might have steered a larger portion of the inflowing waters into the Murman Current. This can explain the tongue of cold waters extending from the Norwegian Sea into the Barents Sea along the Norwegian-Russian coast (Fig. 6). The cyclonic centre probably also resulted in more north-easterly winds and lower air temperatures northwest of Novaya Zemlya. This would in turn enhance heat loss to the atmosphere and increase the ice production. In northern parts the ice could drift into the Atlantic Water causing a further cooling due to melting. The temperature decrease in eastern and northern parts was therefore probably related to local cooling rather than changing in relative strength between the two main branches.

1995 was an interesting year due to extremely high summer temperature anomalies in the eastern and northern parts of the Barents Sea (Fig. 6). The hydrographic sections show that there was a rapid temperature increase in the southern Barents Sea in late 1994 and early 1995 (Fig. 5 and ANON, 2001). Advection of this relatively warm water to the eastern Barents Sea may explain some of the positive temperature anomaly. According to Vinje (2001), there also was a minimum extent of winter sea ice this year, and the Kara Sea was partly ice free for the first time since 1864. He ascribed the minimum to the concerted action from atmospheric and oceanic reducing effects, i.e. reduced northerly winds and higher heat content in the ocean.

The dramatic drop in the NAO winter index in 1996 (e.g. ICES, 2000) resulted in a strong decrease in ocean temperatures in the entire Barents Sea except in some areas in the northern and eastern parts (Fig. 6). The drop was simultaneous in the whole southern area (Fig. 5). The mean SLP for this year (Fig. 7 c) reveals that the secondary low-pressure centre usually found in the Barents Sea during low NAO index, was located east in the Fram Strait. The necessary conditions for a counteracting effect on a low NAO as suggested earlier were therefore not present in 1996.

During the last 3 years of the 1990s there was a gradual build-up towards higher temperatures (Fig. 5 and 6). By the end of the decade the western parts were even warmer than in the beginning of the decade (Fig. 5), while east of 40°E temperatures still were well below average (Fig. 6). The clear shift in the temperature anomalies in 1999 from warm to cold along 40°E indicates that the warming in the branch going north into the Hopen Trench was much higher than the warming in the Murman Current (Fig. 6). The winter mean SLP for the preceding winter (i.e. 1999) shows much stronger pressure gradients compared to the high index year of 1990 (Fig. 7 a and d). The difference in SLP between the eastern and western parts of the Barents Sea was 14 hPa in 1999 compared to 10 hPa in 1990. Consequently, there might have been an atmospheric blocking causing the inflowing water to go northwards into the Hopen Trench rather than eastwards enhancing the Murman Current. The temperature anomalies for the sections (Fig. 5) reveal that very high temperatures were observed in FB in January 1999. By March the same year the high anomalies were observed in Vardo-N, but they never reached Kola. The atmospheric blocking probably both decreased the
oceanic winter inflow and pushed the incoming warmer water northwards, thereby confining it to a smaller area. It is also a possibility that the change in wind pattern caused a change in ice distribution which may have contributed to the lack of heating in the eastern parts, but this is probably less likely as the pressure field should not give less cooling in the western parts nor more cooling in the east. The warming in the late 1990s can therefore be attributed to a different distribution of AW within the Barents Sea, as well as higher temperatures of the inflowing water due to the recovery of the winter NAO.

The high temperature anomalies observed during the early winters of 1999 and 2000 is worth noting (Fig. 5). This is ascribed to mild autumns and late onset of winter cooling, as there were relatively high air temperatures during late the autumn/early winter months in these two years. High sea temperature anomalies during late autumn and early winter were also observed along the Norwegian coast in the same period (ANON 2001), indicating a regional phenomenon, which we at this point can not link to any specific forcing.

5. Summary and conclusions
The investigation may be summarised as follows:
1) The 1990s was the warmest decade since the 1950s. As both the 1930s and the 1950s were warmer than the last decade, the warming of the 1990s may very well be related to natural variability rather than anthropogenic effects.
2) The 1990s started out warm both in the Atlantic and most of the Arctic domain. The low NAO winter index in 1996 resulted in a sudden and simultaneous drop in temperature in the entire Barents Sea except in small areas in the northern parts. During the last years of the decade the temperatures have gradually increased in the western parts, but were still well below average east of 40°E.
3) The temperature variability in the three standard sections Fugloya-Bear Island, Vardo-N and Kola in the southern parts of the Barents Sea all give a fairly good representation of the climate fluctuations in the Atlantic domain.
4) During positive NAO winter index the joint action of (at least) three oceanic responses may give an increase in the Barents Sea temperatures. These responses are: 1) more southerly winds enhancing the inflow, 2) more winter storms penetrating the Barents Sea, 3) the narrowing of the NAC (Blindheim et al., 2000), and the associated lower heat loss to the atmosphere, giving higher temperatures in the inflowing AW.
5) The NAO has, through regional and local effects, a significant influence on the Barents Sea on decadal time scales and during extreme NAO events. Still, local atmospheric forcing not captured by the NAO index seems to be dominating for the distribution of the water masses within the area. The local pressure field appears to change the relative strength of the two branches going respectively northeast and east, thereby having a significant effect on the local climate. The local pressure distribution not captured by the NAO index also has some influence on the total inflow to the Barents Sea.

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