Chapter 1
The Inflow of Atlantic Water, Heat, and Salt to the Nordic Seas Across the Greenland–Scotland Ridge

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1.1 Introduction

The flow of warm, saline water from the Atlantic Ocean (the *Atlantic inflow* or just *inflow*) across the Greenland–Scotland Ridge into the Nordic Seas and the Arctic Ocean (collectively termed the Arctic Mediterranean) is of major importance, both for the regional climate and for the global thermohaline circulation. Through its heat transport, it keeps large areas north of the Ridge much warmer, than they would otherwise have been, and free of ice (Seager et al. 2002). At the same time, the Atlantic inflow carries salt northwards, which helps maintaining high densities in the upper layers; a precondition for thermohaline ventilation.

The Atlantic inflow is carried by three separate branches, which here are termed: the *Iceland branch* (the North Icelandic Irminger Current), the *Faroe branch* (the Faroe Current), and the *Shetland branch* (Fig. 1.1). These are all characterized by being warmer and more saline than the waters that they meet after crossing the Ridge, although both temperature and salinity decrease as we go from the Shetland branch, through the Faroe branch, to the Iceland branch. All these branches therefore carry, not only water, but also heat and salt across the Ridge.

Systematic investigations on the Atlantic inflow started already at the start of the 20th century with the Shetland branch, which long was treated as by far the dominant inflow branch. These investigations were mainly carried out by Scottish researchers and included measurements of temperature and salinity on two standard sections in the Faroe–Shetland Channel (Turrell 1995). Later, similar investigations were

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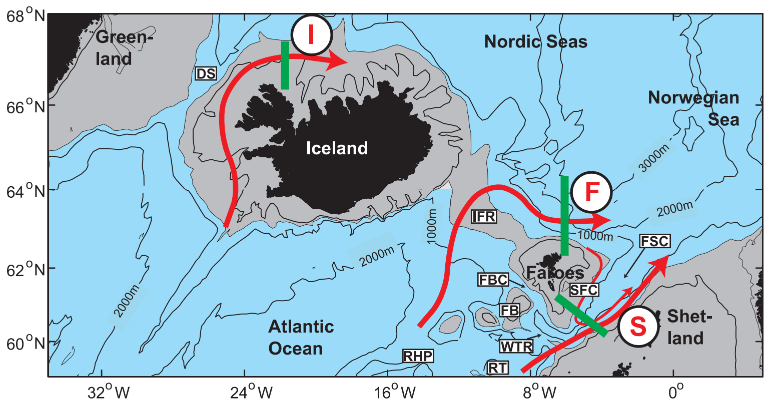

Fig. 1.1 Bottom topography between Greenland and Shetland. Shaded areas are shallower than 500 m. Thick red arrows indicate the three inflow branches: the Iceland branch (I), the Faroe branch (F), and the Shetland branch (S). A thinner red arrow indicates the "Southern Faroe Current (SFC)" and its re-circulation in the Faroe–Shetland Channel (FSC). Thick green lines show the locations of standard sections along which hydrographic and current data have been obtained. Indicated locations are: the Denmark Strait (DS), the Iceland–Faroe Ridge (IFR), the Faroe Bank Channel (FBC), the Faroe Bank (FB), the Faroe–Shetland Channel (FSC), the Wyville–Thomson Ridge (WTR), the Rockall Trough (RT), and the Rockall–Hatton Plateau (RHP)

initiated on the Iceland branch and on the Faroe branch. Sporadic attempts were made to measure currents from research vessels early in the 20th century, but systematic long-term measurements with moored current meters were only initiated in 1985 when Icelandic researchers started monitoring the currents in the Iceland branch (Kristmannsson 1998). For the other two branches, systematic current measurements were initiated with the Nordic WOCE project in the mid-1990s. Building on this, a system has been established, which monitors all the branches of the Atlantic inflow with regular CTD cruises and quasi-permanent current meter moorings. The system is maintained by research vessels from the marine research institutes in Iceland, the Faroes, and Scotland and has received support from the European research programmes through the projects VEINS (Variability of Exchanges In the Northern Seas) and MAIA (Monitoring the Atlantic Inflow toward the Arctic).

This system was further maintained and refined in the MOEN (Meridional Overturning Exchange with the Nordic Seas) project, which was supported by the European FP5, and was a component of ASOF. In the framework of this project, measurements of temperature, salinity, and currents were continued through the ASOF period. ASOF-MOEN also included a numerical modelling component, which studied the exchanges across the Greenland–Scotland Ridge, using an ocean model driven by atmospheric fluxes from reanalysis fields.

The aim of this chapter is to synthesize the information on the Atlantic inflow across the Greenland–Scotland Ridge, based mainly on the results gained by the ASOF-MOEN project and its predecessors, but including other relevant sources, as well. No attempt will be made to repeat the more detailed reviews that have included the Atlantic inflow (Johannesen 1986; Hopkins 1991; Hansen and Østerhus

2000) and neither will we attempt to make a systematic distinction between ASOF and non-ASOF produced results.

1.2 The General Setting

1.2.1 Topographic Constraints

The Greenland–Scotland Ridge separates the Arctic Mediterranean from the Atlantic Ocean and acts as a constraint on all the exchanges across it, the Atlantic inflow as well as the East Greenland Current, and the overflows. On a section (Wilkenskjeld and Quadfasel 2005) following the crest of the Ridge (Fig. 1.2), the warm and saline Atlantic water is seen to be most prominent in the south-eastern parts, where it dominates the section, above the cold and less saline overflow water flowing over the Ridge in many places. In the surface, the Atlantic water extends west of Iceland (Fig. 1.2). The Ridge reaches above the sea surface in Iceland and the Faroes, which split it into three gaps, and this determines the branching structure (Fig. 1.3).

The gap between Greenland and Iceland, the Denmark Strait, is wide and reaches a depth of 640 m. The Atlantic inflow through this gap has to share the cross-sectional area with both the East Greenland Current and the Denmark Strait overflow, and is confined to the easternmost part of the strait.

Between Iceland and the Faroes, the Atlantic water has to flow across the Iceland–Faroe Ridge, which has typical sill depths from 300–480 m along its crest. Atlantic water crosses this ridge over its whole width, in many places passing above the cold overflow water that intermittently crosses the Ridge in the opposite direction.

The Atlantic water that passes between the Faroes and Shetland, can do so along several different routes. The warmest and most saline component flows over the slope as the "Slope Current" (Swallow et al. 1977; Ellett et al. 1979), or "Shelf Edge Current" (New et al. 2001), which has its origin to the south of the Rockall Trough. In addition to this, water of more oceanic origin can pass through the

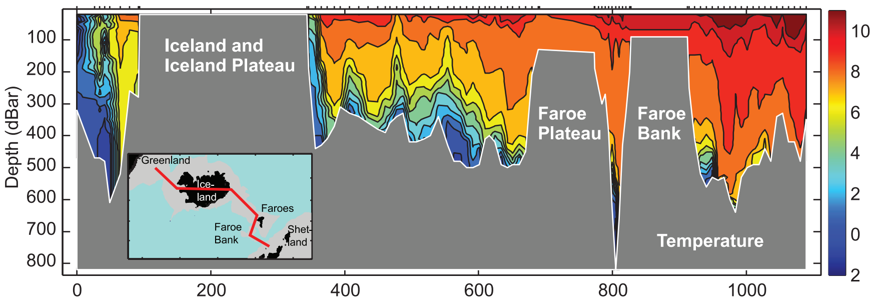

Fig. 1.2 A section following the crest of the Greenland–Scotland Ridge (red line on inset map) showing the temperature in degree Celsius during a cruise in summer 2001

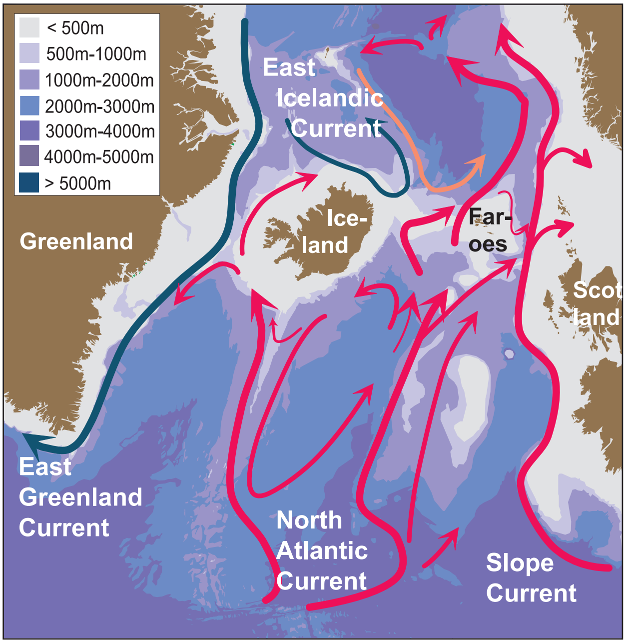

Fig. 1.3 Main flow patterns of warm (red arrows) and cold (blue arrows) currents in the upper layers of the Northeastern North Atlantic. Background colours indicate bottom depth

Rockall Trough, over the Rockall–Hatton Plateau, and even through the Faroe Bank Channel to reach the Faroe–Shetland Channel, although the persistence of some of these pathways is unknown. As these waters pass south of the Faroes, they meet a counter-flow of Atlantic water over the south-eastern Faroe slope. This flow, termed the "Southern Faroe Current" by (Hátún 2004), derives from the Faroe branch. Most of it recirculates in the Faroe–Shetland Channel and joins the other Atlantic water masses in the Shetland branch (Hansen and Østerhus 2000).

1.2.2 The Origin of the Atlantic Inflow Water

In much of the classical literature (see, e.g. review by Hansen and Østerhus 2000), the Atlantic water crossing the Ridge was seen to derive either from an oceanic or from a more continental source (Fig. 1.3). The oceanic source fed the Iceland branch, the Faroe branch, and part of the Shetland branch, whereas the continental source fed the Slope Current and thereby the Shetland branch of the Atlantic inflow. In the Faroe–Shetland Channel, especially, waters from these two sources were treated as different water masses: the "North Atlantic Water (NAW)", carried by the Continental Slope Current, and the "Modified North Atlantic Water (MNAW)", deriving from the oceanic source.

An extreme version of this view, was the proposal by Reid (1979), who suggested a direct import of Mediterranean Water to the Nordic Seas. This suggestion never gained much support and recent observational (McCartney and Mauritzen 2001) and modelling (New et al. 2001) studies have rejected it convincingly.

Distinguishing between an oceanic and a continental source does, however, ignore the continuous exchange between the waters of the Continental Slope Current and the adjacent off-shore waters and time-series show a high degree of coherence between the different Atlantic inflow branches (Section 1.4.2), whether over the continental slope or farther offshore. An alternative view, therefore, does not distinguish between oceanic and continental origin, but rather considers all the Atlantic inflow branches to be fed from two source water masses: the warm and saline ENAW and the colder and less saline WNAW.

The ENAW (Eastern North Atlantic Water) (Harvey 1982; Pollard et al. 1996) gains it properties in the region south of the Rockall Trough, called the "Inter-gyre region" (Ellett et al. 1986; Read 2001; Holliday 2003). This name might indicate a mixed contribution from the two gyres but, certainly, the ENAW has much less input from the Subpolar Gyre than the other source water mass, the WNAW (Western North Atlantic Water), which is carried towards the inflow areas by the North Atlantic Current. The North Atlantic Current is generally considered to originate in the Subtropical Gyre, but it is bounded by the Subpolar Gyre on its northern flank and water from that gyre is admixed into the flow. When it reaches the eastern North Atlantic, it has received sufficient amounts of Sub-Arctic Intermediate Water (SAIW), so that the WNAW is colder and fresher than the ENAW.

1.2.3 The Downstream Fate of the Atlantic Inflow Water

After passing the Greenland–Scotland Ridge, the different branches of Atlantic water progress into the Nordic Seas and from there, parts of the water continue into the Arctic Ocean. The details of the paths and associated water mass changes on route have been reviewed by various authors (Johannesen 1986; Hopkins 1991; Mauritzen 1996; Hansen and Østerhus 2000; Blindheim and Østerhus 2005). The main point to note is that the three different branches affect different regions in the Arctic Mediterranean. The Iceland branch has direct effects only on the southern parts of the Iceland Sea (Swift and Aagaard 1981; Jónsson 1992). The Faroe branch apparently feeds the recirculating water in the southern Norwegian Sea (Fig. 1.3) and thus probably delivers much of its heat and salt to these areas. A part of the Faroe branch also joins with the Shetland branch, which must be considered the main contributor to the North Sea and probably also the Barents Sea.

1.3 Monitoring System

Our knowledge of the Atlantic inflow has been accumulated from a long history of observations, mainly on the hydrography. Here, we focus on the observational system that has been established to monitor the three Atlantic inflow branches and was used in the ASOF-MOEN project.

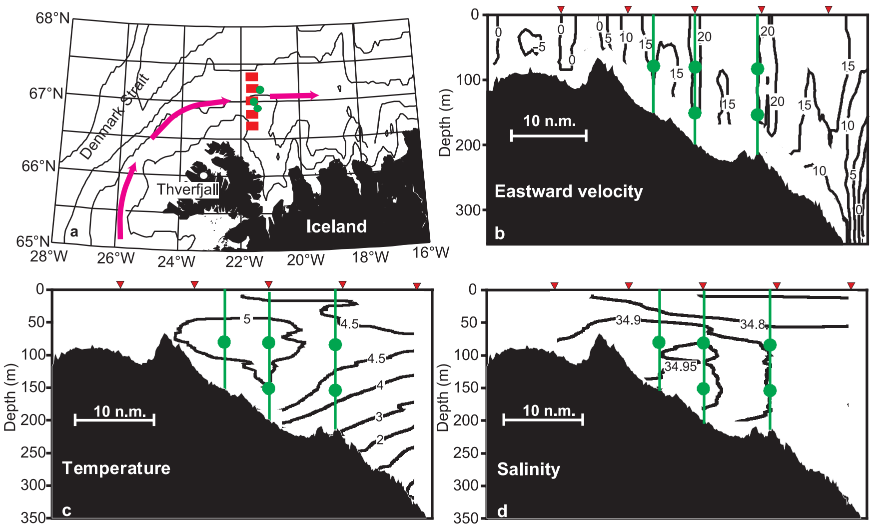

Fig. 1.4 Monitoring system and properties of the Iceland branch. (a) CTD standard stations are indicated by red rectangles. Current meter mooring sites are indicated by green circles. Magenta arrows indicate Atlantic water pathways towards and through the section. (b) Average eastward velocity (cm s^{-1}) based on a total of 20 sections of vessel mounted ADCP data from November 2001–2004, and August 2005 with four sections taken each time. CTD standard stations (red triangles) and current meter moorings (green lines with green circles indicating Aanderaa current meters) are shown. (c, d) Average distributions of temperature in degree Celsius (c) and salinity (d) on the section, based on CTD observations at standard stations (red triangles) in the period 1999–2001

The systematic observations of the Iceland branch have been focused on the Hornbanki section (green line labelled "I" on Fig. 1.1). On this section (Fig. 1.4), CTD profiles have been obtained by the Marine Research Institute in Iceland on several standard stations up to four times a year since 1994 and, during the same period, the inflow of Atlantic water has been monitored by moored current meters. From September 1999, the measurements were extended to three moorings carrying a total of five current meters (Fig. 1.4).

The Faroe branch has been monitored on a section extending northwards from the Faroes along the 6°05′ W meridian (green line labelled "F" on Fig. 1.1). On this section (Fig. 1.5), CTD profiles have been acquired by the Faroese Fisheries Laboratory on several standard stations, at least four times a year since 1988. From the mid-1990s, ADCPs have been moored on the section almost continuously. The number and locations of ADCP moorings have varied somewhat, but since summer 1997, there have always been at least three and sometimes five ADCPs on the section, except for annual servicing gaps.

The observations of the Shetland branch were carried out on a section crossing the channel south of the Faroes (green line labelled "S" on Fig. 1.1). At least four, and before summer 2000, five ADCP moorings have been maintained along the section since November 1994 (Fig. 1.6). These observations have been complemented with ADCP data acquired from oil platforms. Both the Faroese Fisheries Laboratory and the Marine Laboratory in Aberdeen do regular CTD cruises along

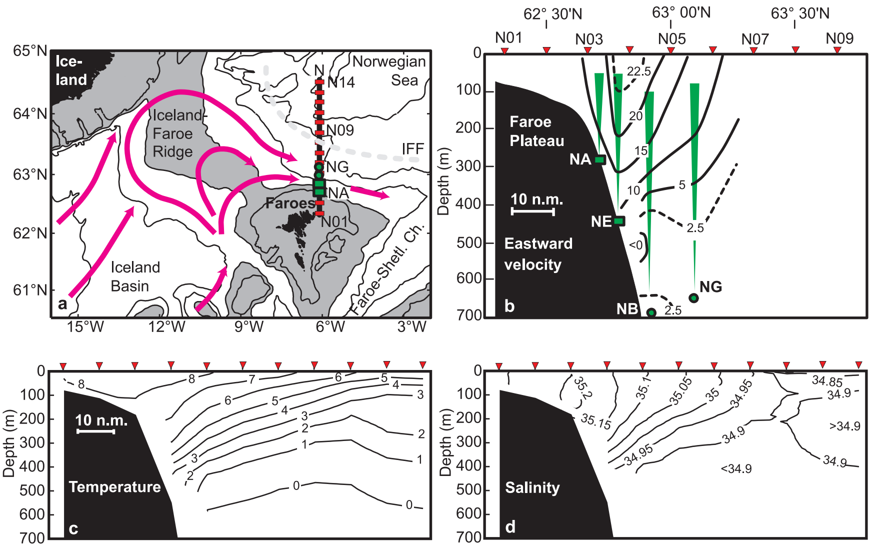

Fig. 1.5 Monitoring system and properties of the Faroe branch. (a) CTD standard stations are indicated by red rectangles, labeled N01–N14. ADCP mooring sites are indicated by green circles (traditional moorings) or rectangles (trawl-proof frames) labeled NA to NG. Shaded areas are shallower than 500 m. The dotted yellow curve indicates the general location of the Iceland–Faroes Front (IFF) and magenta arrows indicate Atlantic water pathways towards and through the section. (b) Average eastward velocity (cm s^{-1}) 1997–2001. The innermost CTD standard stations (red triangles) are indicated as well as the ADCP mooring sites (green circles or rectangles with green cones indicating sound beams). (c, d) Average distributions of temperature in degree Celsius (c) and salinity (d) on the inner part of the section, based on CTD observations at standard stations (red triangles) in the period 1987–2001

this section and altogether four to eight CTD sections have been obtained annually since the mid-1990s.

The region between Iceland and Shetland is heavily fished and traditional current meter moorings have a short survival time in this area. This was the reason for using upward-looking ADCPs instead of more traditional instrumentation. At deep sites, the ADCPs are moored in the top of traditional moorings with the ADCP sufficiently deep to escape trawls. On the slope north of the Faroes, two of the ADCPs are deployed directly on the bottom within frames that protect the ADCPs and other instrumentation from fishing gear (Fig. 1.7).

1.4 Observed Properties

1.4.1 Typical Structure and Properties of the Inflow Branches

The Iceland branch is highly variable but it is of great importance to the regional marine climate and hence the ecosystem in North Icelandic waters (Jónsson and Valdimarsson 2005). There is usually a core of Atlantic water identified by high

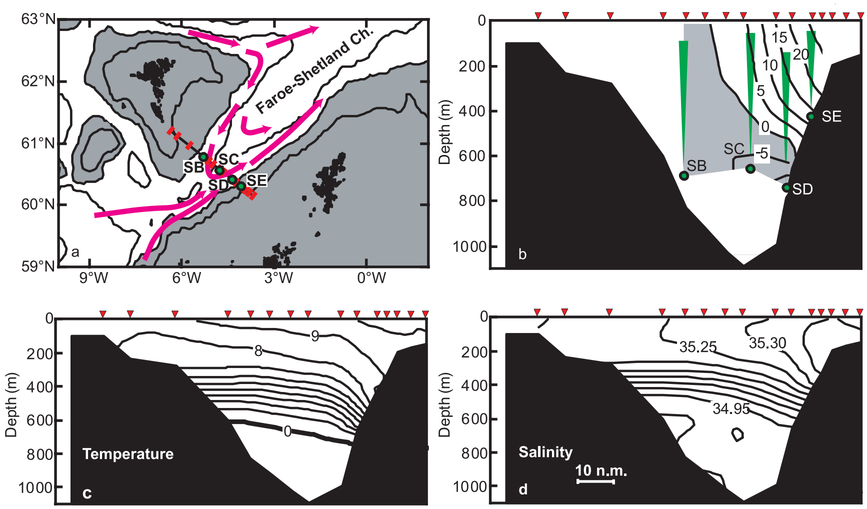

Fig. 1.6 Monitoring system and properties of the Shetland branch. (a) CTD standard stations are indicated by red rectangles. ADCP mooring sites are indicated by green circles labeled SB, SC, SD, and SE. Shaded areas are shallower than 500 m and magenta arrows indicate Atlantic water pathways towards and through the section. (b) Average along-channel velocity (cm s^{-1}) as measured by the ADCP moorings in the period 1994–2005. Shaded area indicates reverse (SW-going) flow. CTD standard stations (red triangles) and ADCP mooring sites (green circles with green cones indicating sound beams) are indicated. (c, d) Average distributions of temperature in degree Celsius (c) and salinity (d) on the section, based on CTD observations at standard stations (red triangles) in the period 1994–2005

Fig. 1.7 Trawl-proof frame containing ADCP, double acoustic releases, ARGOS beacon, and buoyancy, on top of concrete anchor, is being made ready for deployment onboard R/V Magnus Heinason

salinity and temperature, but its location and extent are variable. In Fig. 1.4c, d, the core can be identified over the area covered by the current meters. In this branch, the Atlantic water does not seem to reach deeper than 200 m (Jónsson and Briem 2003). North of the Atlantic water core, the region may variably be dominated by Arctic water masses from the Iceland Sea or Polar water masses from the East Greenland Current.

The Faroe branch carries the Atlantic water that has crossed the Iceland–Faroe Ridge. Northeast of the Ridge, this water meets the much colder and less saline waters of the East Icelandic Current and gets confined into a fairly narrow current, which flows eastwards over the northern slope of the Faroe Plateau. Relatively high temperature and salinity (Fig. 1.5c, d) characterize the Atlantic water, which usually is concentrated on a wedge-shaped area that is bounded by the Iceland–Faroe Front, which hits the Faroe slope at depths 400–500 m, similar to the sill depth of the Iceland–Faroe Ridge. From below, the Atlantic layer is bounded by two water masses (Hansen and Østerhus 2000): the Norwegian Sea Arctic Intermediate Water (NSAIW), which occupies the top of the deep water in the Norwegian Sea, and the Modified East Icelandic Water (MEIW), which derives from the East Icelandic Current and usually is characterized by a salinity minimum.

The Shetland branch carries Atlantic water that has entered the Faroe Shetland Channel from the west in addition to water recirculated from the Faroe branch. A section crossing the channel (Fig. 1.6c, d) has Atlantic water, characterized by high temperature and salinity across the whole channel in the upper layers. Temperature and salinity do, however, increase from the Faroe to the Shetland side of the channel with the highest temperatures and salinities in the core of the Slope Current. Below, the Atlantic layer is bounded by the deep NSAIW and by varying amounts of MEIW from the East Icelandic Current.

1.4.2 Long-Term Variations of Temperature and Salinity

The long-term hydrographic observations allow the generation of long time-series of the properties of the Atlantic inflow. The longest time-series are from the Faroe–Shetland Channel (Fig. 1.8) and show variations on many timescales. Both temperature and salinity peaked around the middle of the 20th century. This feature has been shown to be a characteristic of the northern latitudes (Bengtson et al. 2004). A number of anomalies have been noted, especially in the salinity (Belkin et al. 1998), with the “Great” or “Mid-seventies” anomaly (Dickson et al. 1988) being the most pronounced.

The last decades of the 20th and the beginning of the 21st century show increasing trends in both temperature and salinity, which may perhaps be linked to global change, but which also exhibit large variations on a decadal timescale. Hátún et al. (2005) have shown that these variations to a large extent can be explained by variations in the intensity and extent of the Subpolar Gyre circulation (Häkkinen and

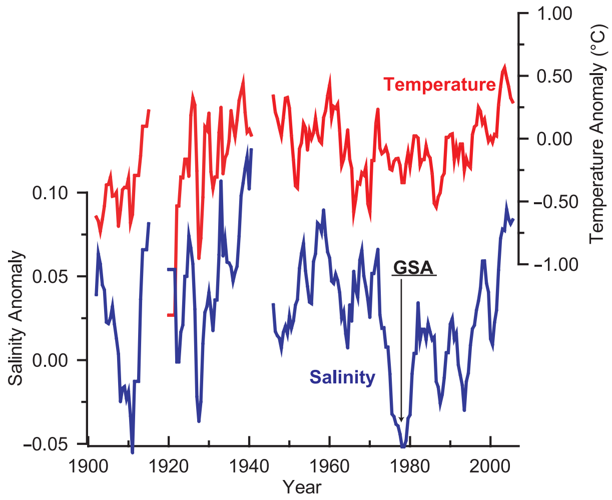

Fig. 1.8 Anomalies of temperature (red) and salinity (blue) over the Scottish shelf. Derived from the temperature and salinity of the water displaying maximum salinity within the Slope Current flowing polewards along the Scottish continental shelf. Anomalies presented are 2-year running means after the average (1961–1990) seasonal cycle has been removed. The "Great Salinity Anomaly (GSA)" is indicated

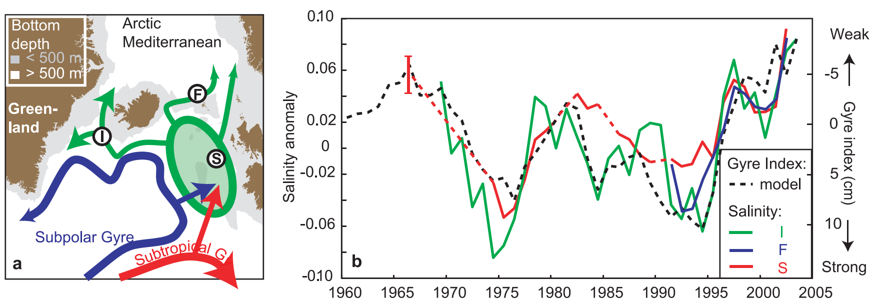

Fig. 1.9 (a) According to Hátún et al. (2005), all of the inflow branches (green arrows) are fed partly from the Subtropical and partly from the Subpolar Gyre in relative amounts that depend on the intensity and extent of the Subpolar Gyre circulation, expressed by the gyre index (Häkkinen and Rhines 2004). (b) The gyre index (dotted black line) from a model is plotted together with observed salinity in the three inflow branches

Rhines 2004). This intensity, the gyre index, is found to correlate well with inflow properties of all the branches (Fig. 1.9) and, using a numerical model, Hátún et al. (2005) could explain this in terms of the source water masses (Section 1.2.2). When the gyre index is high, the Subpolar Gyre extends far towards the east and relatively large amounts of WNAW are transported towards the inflow areas, whereas the warmer and saltier ENAW tends to dominate more, when the gyre index is low. The properties of the Atlantic inflow, thus, seem to be governed by the intensity of the Subpolar

Gyre circulation, which again is believed to depend mainly on the buoyancy flux and convection in the Labrador–Irminger Seas (Häkkinen and Rhines 2004).

1.5 Observed Fluxes

1.5.1 Methods for Flux Estimation

Volume flux through a section is, in principle, simple to calculate as the integral of the normal velocity component over the section. All of the Atlantic inflow branches do, however, flow together with other water masses. For none of the branches, is it possible to define a section that covers all the Atlantic water on the section and no other waters.

To calculate the volume flux of Atlantic water through a section crossing one of the inflow branches, it is therefore necessary to know, not only the normal velocity, but also the fraction of Atlantic water on the section. The methodology is illustrated in Fig. 1.10 and the volume flux of Atlantic water is computed as:

$$V_A(t) = \sum_k \sum_j A_{k,j} \cdot u_{k,j}(t) \cdot \beta_{k,j}(t) \tag{1.1}$$

where the sum is over all the boxes that the section is subdivided into and $\beta_{k,j}(t)$ is the fraction of Atlantic water in box (k,j) at time t. First of all this requires, of course, a definition of Atlantic water. In the literature on the Nordic Seas and Arctic Ocean, the concept of Atlantic water is often defined by its salinity, e.g. as water more saline than 35. Here, we define the flux of Atlantic water as the flux of water crossing the Greenland–Scotland Ridge into the Nordic Seas.

With comprehensive velocity measurements that provide $u_{k,j}(t)$, the problem is reduced to the determination of $\beta_{k,j}(t)$. This is not a trivial problem, but, in principle it can be solved if not too many different source water masses are involved, and if the characteristics (T, S) of these as well as the waters on the section are known. Differences in data availability and conditions have led to different procedures for the different branches. For the two easternmost branches, the Atlantic water fraction in each sub-area is determined from temperature and salinity measurements by using a three-point mixing model (Hansen et al. 2003; Hughes et al. 2006). Fluxes of volume of the Atlantic water component through each sub-area are then computed and summed. For the Iceland branch, the Atlantic water fraction is determined from temperature observations of the pure Atlantic water and polar water upstream and the temperature observed at the current meters (Jónsson and Valdimarsson 2005).

In addition to volume (mass) flux carried by the various branches, heat and salt fluxes are highly relevant. These fluxes are not, however, meaningful, unless the temperature and salinity of the water returning to the Atlantic are known. Instead of producing heat- and salt fluxes, we therefore compute average values of temperature and salinity of the different inflow branches, where the average is weighted

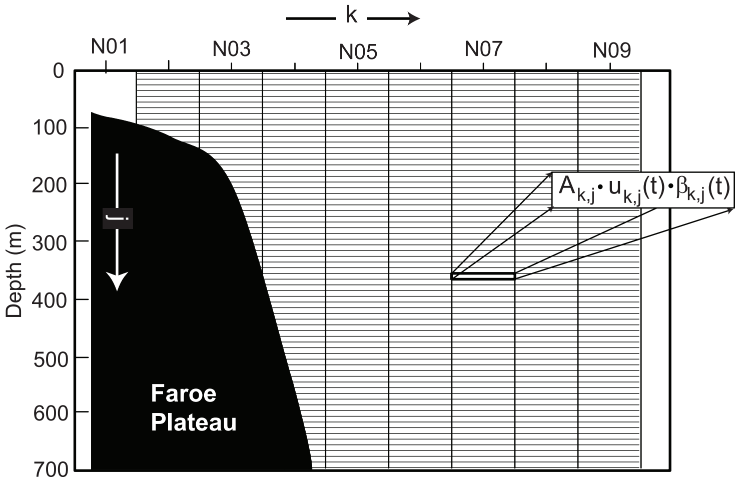

Fig. 1.10 The method for calculating Atlantic water flux exemplified for the section crossing the Faroe branch. The section is subdivided into boxes, which are labelled by indices *k* and *j*. One of the boxes is shown in a magnified scale indicating the parameters that must be assigned to each box: The area: $A_{k,j}$, the eastward velocity: $u_{k,j}(t)$, and the fraction of Atlantic water: $\beta_{k,j}(t)$. Each vertical column of boxes is centered around a standard CTD station (labelled N01–N09)

with respect to volume flux. Since the Atlantic water always is found together with water of other origins, this has to be done with care, as discussed by Hansen et al. (2003). If similar values can be produced for all the other exchange flows, meaningful heat- and salt-budgets can be produced.

1.5.2 Flux Estimates for the Individual Inflow Branches

Flux estimates for the individual inflow branches have been given in a number of publications. For the Iceland branch, Jónsson and Valdimarsson (2005) have determined the Atlantic water fraction within the inflow area of the Hornbanki section as a function of time and computed fluxes for the 1994–2000 period. The average volume flux of Atlantic water was found to be 0.8 Sv. No seasonal variation was found in current velocities (Jónsson and Briem 2003), but the Atlantic water fraction varied seasonally, which gave rise to a seasonal amplitude of 0.2 Sv for the volume flux of Atlantic water with a maximum in September. Monthly averaged volume flux ranged between 0 and 1.3 Sv.

For the Faroe branch, Hansen et al. (2003) have analysed the observations from the June 1997 to June 2001 period. On average, the Faroe branch transported a volume flux of 3.5 ± 0.5 Sv of Atlantic water. Monthly averaged volume flux ranged between 2.2 and 5.8 Sv, but with only a small seasonal variation. Daily averages ranged between 0.3 and 7.8 Sv, with not a single flow reversal during the 4-year period.

For the Shetland branch, Hughes et al. (2006) have analysed the fluxes through the Channel. On average, the Atlantic water flux was estimated at 3.9 Sv for the period September 1994–May 2005. The flux was found to have a seasonal variation with an amplitude of 0.8 Sv, which is 21% of the average, and maximum in November. Monthly averaged Atlantic water flux ranged between 0.8 and 7.5 Sv.

1.5.3 The Total Atlantic Inflow 1999–2001

For the 3-year period from 1 January 1999 to 31 December 2001, Østerhus et al. (2005) computed volume fluxes (Fig. 1.11) and average temperature and salinity values for each of the branches and combined them to produce overall values for the total Atlantic inflow (Table 1.1). The average values for the volume fluxes of the various branches differ slightly from previously published values (Østerhus et al. 2001; Hansen et al. 2003; Turrell et al. 2003; Jónsson and Valdimarsson 2005) but the deviations are small and may be due to the different averaging periods.

Østerhus et al. (2005) estimated an uncertainty of about 1 Sv for the average total volume flux of Atlantic water. Within this uncertainty, their estimate of the total volume flux (8.5 Sv) is consistent with the preliminary estimate reported by Hansen and Østerhus (2000) and also remarkably close to the classical value published by Worthington (1970).

For the 1999–2001 period with concurrent measurements, the Iceland branch was found to carry 10% of the Atlantic inflow volume flux, with the other two branches carrying 45% each. Monthly averaged volume fluxes for each branch and for the total inflow during this period are shown in Fig. 1.11. Although they are of similar intensity on the average, Fig. 1.11 indicates larger variations in the Shetland

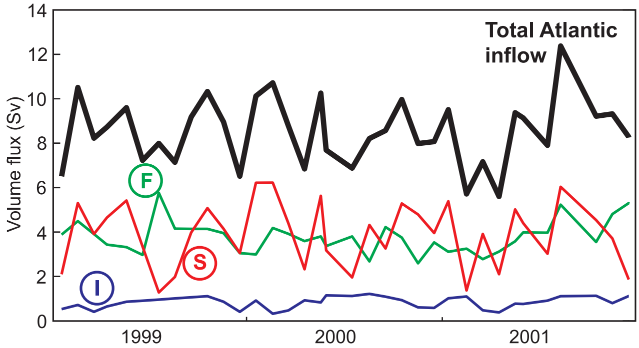

Fig. 1.11 Monthly averaged volume flux of Atlantic water in each of the three branches (coloured lines labelled as the Iceland branch (I), the Faroe branch (F), and the Shetland branch (S)) and in the total Atlantic inflow (black line) for the 1999–2001 period

Table 1.1 Observed characteristics of each of the three Atlantic inflow branches and of the total inflow for the period January 1999–December 2001

	Average			Seasonal var. of vol. flux		
	Vol. flux	Temp.	Sal.	Ampl.	Max.	Signif.
Inflow branch	Sv	°C		Sv	Month	
Iceland branch	0.8	6.0	≤35.00	0.2	Sept.	<0.01
Faroe branch	3.8	8.2	35.23	0.3	Oct.	n.s.
Shetland branch	3.8	9.5	35.32	0.2	Mar.	n.s.
Total Atl. Infl.	**8.5**	**8.5**	**35.25**	**0.4**	**Oct.**	**n.s.**

branch than in the Faroe branch. This might be due to differences in precision of the estimates. Certainly, the Shetland branch is more difficult to monitor accurately due to the recirculation in the Faroe–Shetland Channel and the intensity of mesoscale activity (Sherwin et al. 2006). Results from the ASOF-MOEN numerical modelling activities do, however, show a similar difference between the two branches (Section 1.6).

For the 1999–2001 period, Østerhus et al. (2005) found evidence for a seasonal signal in the Iceland branch with maximum volume flux in September, but the other two branches, as well as the total inflow, showed no statistically significant seasonal variation of the volume flux (Table 1.1). They concluded that a possible seasonal variation of the total Atlantic inflow did not exceed the observational uncertainty, estimated at 1 Sv, in amplitude during this period.

This might seem to conflict with reports of considerably larger seasonal variations in the Norwegian Atlantic Current on the Svinøy section, downstream from the Ridge (Orvik et al. 2001). They only had long-term direct current measurements from the inner branch of this flow, however, and the outer branch has been reported to vary in counter-phase to the inner branch (Mork and Blindheim 2000). The relatively weak seasonal variation of the inflow over the Ridge is therefore consistent with the conclusion of Jakobsen et al. (2003) that the winter intensification of the flow at selected locations like the Svinøy section is primarily linked to spin-up of the local basin gyres.

1.6 Numerical Modelling of the Atlantic Inflow

A number of ocean modelling studies have addressed the Atlantic inflow based on different ocean general circulation models of varying resolution and experimental design (e.g. Karcher et al. 2003; Nilsen et al. 2003; Zhang et al. 2004; Drange et al. 2005). No attempt will be made here to review these studies. Instead, this section presents results from the modelling effort within the ASOF-MOEN project, carried out at the Danish Meteorological Institute. The results are based on an ensemble hindcast simulation for the period 1948–2005 using a global coupled ocean/sea-ice ocean model of relatively coarse resolution (MPI-OM, Marsland et al. 2003), constrained by atmospheric reanalysis data (NCEP/NCAR, Kistler et al. 2001) and observed Arctic river discharges (http://grdc.bafg.de).

The model experiment and results are described in Olsen and Schmith (2007) with a focus on the climatology of the exchanges between the Nordic Seas and the North Atlantic as defined at a set of key sections characterizing the system. The ensemble approach applied in the model experiment is designed to eliminate the role of internal modes of variability and initial ocean conditions on the simulated ocean climate variability and, thus, to isolate the forced response by known atmospheric changes (e.g. the NAO).

Despite the global domain and coarse average resolution, the displacement of the North Pole onto Greenland in the model grid by making use of the curvilinear coordinates results in relatively high resolution in the Nordic Seas. Therefore all three branches of Atlantic surface inflow to the Nordic Seas (Fig. 1.1) can be identified from the simulated upper ocean velocity and tracer fields (Olsen and Schmith 2007): the Iceland branch north of Iceland, the Faroe branch between the Faroes and Iceland, in the model found close to the Icelandic shelf break turning east upon passage of the Ridge, and finally, the Shetland branch, modeled as a broad inflow extending off the Scottish Slope.

At the defined sections, water mass properties are used to distinguish between transports in individual branches of flow. Model mean exchanges for the period 1948–2005 are shown to compare favourably with existing observational estimates for several flow branches in the area, including the exchanges across the Greenland–Scotland Ridge (Olsen and Schmith 2007; see also Chapter 19 for a comparison between model results and observations). For the Atlantic inflow, this is also illustrated in Fig. 1.12 and Table 1.2 for the recent period 1999–2001 with concurrent, high quality observations of all inflow branches (Fig. 1.11).

It is seen that in total, model inflow is about 0.5 Sv higher than observed, which is linked to excess model transport in the Shetland branch compared to observations. The discrepancy is somewhat lower when comparing long-term mean model results with observations (Table 1.2). According to the model results, the total

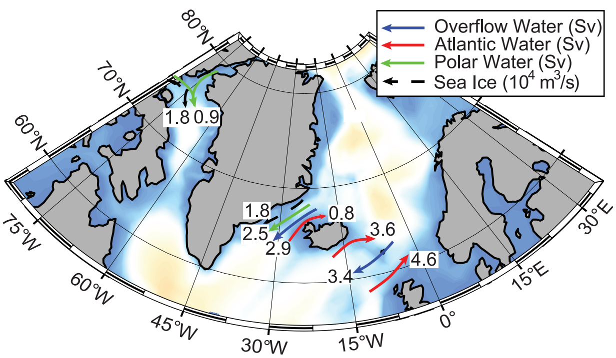

Fig. 1.12 Ensemble- and time-averaged exchanges between the Arctic Mediterranean and North Atlantic for the period 1999–2001 according to the ASOF-MOEN model

Table 1.2 Average values and seasonality (amplitude and time of maximum) of volume fluxes (in Sv) of individual branches and the total Atlantic inflow derived from the ASOF-MOEN model, compared to observed values (based on Østerhus et al. 2005). Observations are only for the 1999–2001 period. Model results are shown partly for this period, partly for the whole (1948–2005) period (All). The correlation coefficients (Corr.) between fluxes from the model and from observations are based on monthly averages and "n.s." indicates "not significant"

Parameter	Average flux			Corr.	Seasonality			
Period	1999–2001		All	1999–2001	1999–2001 (obs)		All (mod)	
Method	Mod.	Obs.	Mod.	Mod.–Obs.	Ampl.	Max.	Ampl.	Max.
Iceland branch	0.8	0.8	0.7	0.74	0.2	Sept.	0.3	Sept.
Faroe branch	3.6	3.8	3.8	−0.38 ns	0.3	Oct.	0.6	Mar.
Shetland branch	4.6	3.8	4.2	0.28 ns	0.2	Mar.	1.2	Dec.
Total Atl. Infl.	9.0	8.5	8.7	0.22 ns	0.4	Oct.	1.2	Dec.

Atlantic inflow did not vary much throughout the 1948–2005 period with only a slight trend, but there was a strengthening of the Shetland branch and a weakening of the Faroe branch (Fig. 1.13b).

The indications in the ASOF-MOEN model results of a nearly constant total Atlantic inflow tend to agree with the findings of Nilsen et al. (2003) from a different model, though in that study, neither of the branches of inflow show robust tendencies, in contrast to the present results. Such stable inflow is, however, at odds with the modeled increase reported by Zhang et al. (2004). Also, the negative correlation between the Shetland branch and the Faroe branch, found by Nilsen et al. (2003), is not supported by the ASOF-MOEN model results.

When comparing model and observations on shorter timescales (Fig. 1.13a), the correspondence is not as good. For the Iceland branch, monthly averaged fluxes in the model and the observations were fairly well correlated, but for the other branches, the correlation coefficients were not significant (Table 1.2). The same conclusion is reached when comparing seasonality in the model and the observations (Table 1.2). Except for the Iceland branch, the model gives higher seasonal amplitudes than the observations and the phases also differ. To some extent, this may be due to different analysis periods. For the 1999–2001 period (Fig. 1.13d), the model does indicate a smaller seasonal amplitude than for the full period (Fig. 1.13c), especially for the Faroe branch. Even for this period, the model still indicates a larger seasonal amplitude than the observations but, when the uncertainties are taken into account, there is no real discrepancy between model and observations.

Summarizing, the ASOF-MOEN model results and the ASOF-MOEN observations show a high degree of correspondence as regards long-term average volume fluxes in the individual branches and the total Atlantic inflow. They also agree on a relatively small seasonal amplitude (<15% of the average flux). They both show fairly similar values for the magnitude of flux variability in the individual branches and total inflow (Fig. 1.13a). It is especially noteworthy, that both observations and model indicate larger monthly variability in the Shetland branch than in the Faroe branch (Fig. 1.13a). When correlating simultaneous monthly averages and seasonal

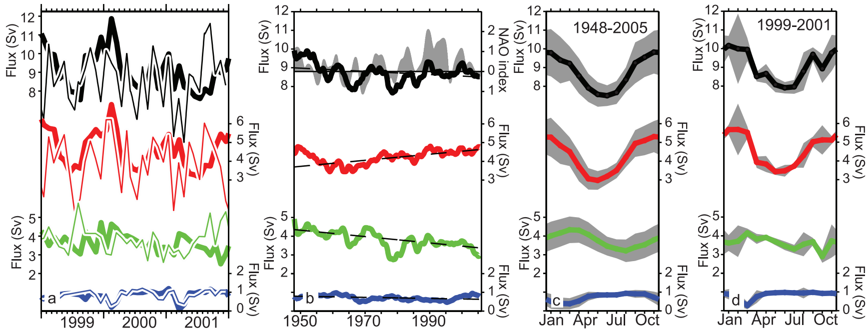

Fig. 1.13 (a) Modeled (thick) and observed (thin) monthly averaged volume flux of Atlantic water in each of the inflow branches and the total Atlantic inflow for the 1999–2001 period. (b) De-seasoned and low-pass filtered (cut-off frequency of 1/24 months^{-1}) modeled volume fluxes and the interpolated, low-pass filtered winter NAO index (gray bars, Jones et al. 1997). The linear trend of each branch is indicated by a thin dashed line. (c, d) Modeled seasonality around the time mean volume flux of each of the inflow branches and the total with the inter-annual spread (2σ) in grey for the periods: 1948–2005 (c), and 1991–2001 (d). The colour coding in all the panels is: Iceland branch (blue), Faroe branch (green), Shetland branch (red), and total Atlantic inflow (black)

variation, however, only the Iceland branch shows significant correlation between the model and the observations. This discrepancy may be due to model deficiencies, or observational inaccuracies, or both, but more work is needed to clarify this.

1.7 Effects of the Atlantic Inflow on the Arctic Mediterranean

It is not the aim of this chapter to give a complete account of the effects of the Atlantic inflow on the Arctic Mediterranean. This topic will be dealt with in other chapters of this book in much more detail, but the Atlantic inflow has tremendous impacts on the area that it enters and no description of it can approach completeness without an overview of the main effects. In the following sections, brief overviews are given for the effects of the Atlantic inflow on the mass (volume), heat, and salt budgets.

1.7.1 Mass Budget

If the estimate by Østerhus et al. (2005) for the 1999–2001 period is used as a basis, the volume flux of the total Atlantic inflow is 8.5 Sv, on the average. In addition to this, 0.8 Sv are reported to enter the Arctic Mediterranean through the Bering Strait (Coachman and Aagaard 1988; Roach et al. 1995) and 0.2 Sv as freshwater (Aagaard and Carmack 1989). Thus, the Atlantic inflow accounts for about 90% of all the water entering the Arctic Mediterranean (Fig. 1.14).

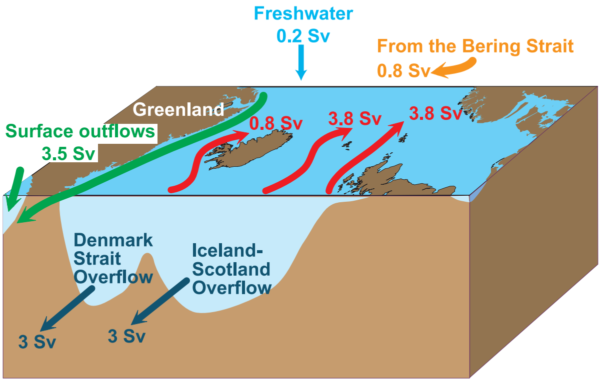

Fig. 1.14 Mass (volume) budget of the Arctic Mediterranean. The value for the volume flux of the surface outflow has been chosen to acquire balance

All of this water has to return to the Atlantic and it does so through several current branches that can be grouped into two main flow systems: the "surface outflow" and the "overflow". The surface outflow includes the East Greenland Current and the flow through the Canadian Archipelago, whereas the overflow includes the deep flow of cold dense water across the Greenland–Scotland Ridge through the Denmark Strait and across the Ridge in different areas east of Iceland. Due to the difficulties of measuring fluxes in shallow ice-covered areas, reliable flux estimates for the total surface outflow have been hard to acquire, but there seems to be a general consensus (Hansen and Østerhus 2000) that the total overflow is around 6 Sv, equally split between the Denmark Strait and the eastern overflow branches (Fig. 1.14).

This is important for understanding the Arctic Mediterranean, but it also has important consequences for the Atlantic inflow, as such. The Bering Strait inflow and the freshwater input are both relatively buoyant and it is not considered likely that they contribute to the overflow (Rudels 1989). This implies that all the overflow water derives from the Atlantic inflow but it also implies that a large fraction of the Atlantic inflow returns as overflow, rather than surface outflow. From Fig. 1.15, this fraction is 71%. This value is, of course, sensitive to uncertainties in the flux estimates, but it is unlikely to be less than 50%. Most of the Atlantic inflow therefore returns as overflow, which has implications for the driving force (Section 1.8).

1.7.2 Heat Budget

The transport of heat to an area by an ocean current can only be determined if the temperatures of all the outflows, as well as the inflows, are known. It is therefore meaningless to consider the heat transport of the Atlantic inflow *per se*. The outflows do, however, have typical temperatures around 0 °C and, with an uncertainty of about 10%, we can therefore estimate the heat import of the Atlantic inflow to the Arctic Mediterranean by using that value for the outflow temperature.

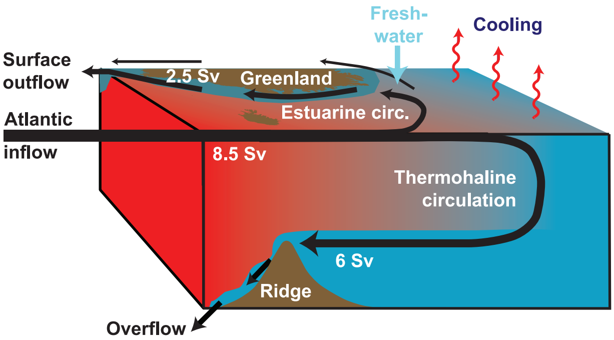

Fig. 1.15 Two circulation systems return the inflowing Atlantic water to the Atlantic Ocean after cooling and freshening. The volume flux of the surface outflow was determined as the difference between the measured Atlantic inflow (this chapter) and the measured overflow (e.g. Hansen and Østerhus 2000). Note that the Bering Strait inflow and its path to the Atlantic is not included in the figure

From this and the average temperatures of the various branches (Table 1.1), the total Atlantic inflow is found to import 310 TW (1 TW = 10^{12} W) of heat. The Shetland branch is the warmest inflow branch and probably contributes most to this heat import, but the different inflow branches may not necessarily contribute equally to the different outflow branches. Hence, the temperature decrease and heat loss of each branch is not well defined without a much more detailed description.

1.7.3 Salt Budget

As for the heat budget, a detailed account of the salt budget requires knowledge of the outflows as well as the inflows. It is, however, possible to make a rough estimate of the Atlantic inflow contribution to the salt budget of the Arctic Mediterranean by a simple calculation. Assuming that 8.5 Sv of Atlantic inflow with salinity 35.25 (Table 1.1) mixes with 0.8 Sv Bering Strait water with salinity 32.5 (Coachman and Aagaard 1988) and with 0.2 Sv of freshwater (Aagaard and Carmack 1989) from runoff and precipitation (P–E), the total outflows must have a volume flux of 9.5 Sv and an average salinity of 34.28.

This implies that the salinity of the Atlantic inflow, on the average, is reduced by ~1, before the water returns to the Atlantic, which may be used to illustrate the (oft-neglected) effect of Atlantic inflow variations on the freshwater balance of the Arctic Mediterranean. Typical variations of the inflow salinity are on the order of 0.1 (Fig. 1.8). A salinity increase of this magnitude would therefore require a 10% increase in freshwater flux in order to maintain a constant average salinity of the Arctic Mediterranean. Similarly, the model results indicate (Fig. 1.13b) that, on decadal timescales, the volume flux of the total Atlantic inflow has varied by about 20% of the average. With constant salinity, this would require a 20% variation in the freshwater flux to maintain balance. These numbers illustrate that both salinity and volume flux variations of the Atlantic inflow need to be taken into account when considering the freshwater (salt) budget of the Arctic Mediterranean.

1.8 Driving Force

All flows in nature require driving forces to accelerate them and maintain them against the retarding effect of friction. This is especially the case for the Atlantic inflow, which exhibits high velocities in the Ridge area (Figs. 1.4–1.6), compared to upstream. These forces may well be affected by future climate change, in which case the inflow may be expected to change. It is therefore important to consider, what forces can drive the Atlantic inflow. The discussion in this section addresses that question, but only as regards the flow across the Ridge, not the circulations in the upstream or downstream basins.

All of the inflow branches are upper layer, surface-intensified, flows, which are fairly uni-directional with depth (Figs. 1.4–1.6). The equations of motion, therefore, include only two external forces that can drive the flow: A surface stress, generated by wind, and a pressure gradient, generated by a sloping sea-surface. By definition, a driving force has to do positive work on the flow and, hence, only the along-flow components of the wind stress or sea-level slope can drive the flow. In the following sections, these two forces are discussed separately and their relative contributions to driving the flow are discussed, although the non-linearity of the system precludes a complete distinction between them.

Both observations (Fig. 1.11) and models (Fig. 1.13) indicate that the total Atlantic inflow is fairly stable with a variable component, superimposed on a constant flow, which seems to contribute considerably more than the variable component, even on timescales as short as a month. The forcing mechanism of the variable component may be studied by correlating flow variations to possible driving forces, but the forcing mechanism of the constant component is more difficult to identify. It is therefore essential to note that the two components may not necessarily have the same forcing.

1.8.1 Wind Forcing

Most upper layer flows in the World Ocean are generally considered to be driven by wind stress and it is natural to assume the same for the Atlantic inflow. This assumption is supported by the fact that the average wind direction in the main inflow region between Iceland and Scotland has a positive component along the inflow path.

The NAO index is commonly used as an indicator of the wind in this region and it may be correlated to the volume flux of the Atlantic inflow. Long time series of the inflow are only available from models and the simulated volume fluxes from the ASOF-MOEN modelling effort can be compared to the NAO index since the ensemble experiment was explicitly designed to disentangle a robust imprint of the variable forcing by the atmospheric reanalysis (Olsen and Schmith 2007). By visual comparison, the total Atlantic inflow shows some similarity to the NAO index (Fig. 1.13b) and the zero-lag correlation coefficient is positive between NAO and the total inflow as well as the Shetland and the Faroe branch, whereas the correlation is negative between NAO and the Iceland branch (Fig. 1.16a). The correlation coefficients are small, however, and not significant statistically, when the autocorrelations of the time-series are taken into account.

To yield further insight into the possible role of the NAO, the correlation analysis is performed for 30-year running segments throughout the hindcast (Fig. 1.16b). This analysis is motivated by the documented shift in the spatial pattern of the NAO in the 1970s, which influenced the marine climate of the Nordic Seas (e.g. Visbeck et al. 2003; Furevik and Nilsen 2005). The results illustrate a near constant imprint of the NAO on the Shetland branch since 1948 with values around 0.3–0.5 though slightly increasing in the latter part. In contrast, a clear shift is seen in the Iceland branch and the Faroe branch from nearly uncorrelated with the index in the early part of the hindcast to being significantly correlated in the recent decades, though with opposite sign; the Faroe branch reaching a positive correlation of 0.72 from 1975 to 2005.

Rather than NAO, it would be preferable to correlate the wind itself with the Atlantic inflow, but what wind parameter? over what region? and with what timelag? (Orvik and Skagseth 2003) addressed that problem by correlating their volume flux measurements off the Norwegian coast with the zonally averaged

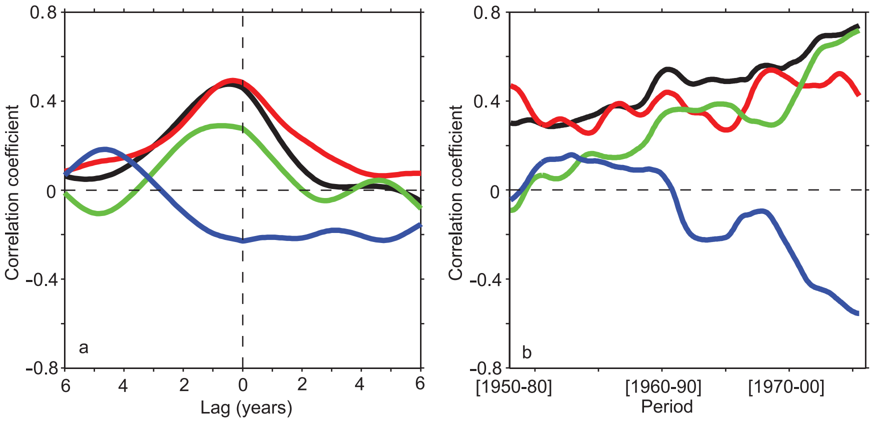

Fig. 1.16 (a) Lagged correlation between the winter NAO index and each individual inflow branch as well as the total Atlantic inflow from the ASOF-MOEN model. (b) Zero-lag correlations of the same parameters for 30-year running segments. Prior to the analysis, the time-series have been de-seasoned and low-pass filtered and the linear trend removed (see Fig. 1.13b). The colour coding in both panels is: Iceland branch (blue), Faroe branch (green), Shetland branch (red), and total Atlantic inflow (black)

North Atlantic wind stress curl at various latitudes and with various lags. They found a maximum correlation coefficient of 0.88 for 55° N and 15 months lag between wind and volume flux. Such a procedure of correlating a variable against several other variables and picking out the maximal correlation does, however, reduce the (already small) number of degrees of freedom and hence the statistical significance of their result. (Sandø and Furevik submitted) were able to partly reproduce these results in an isopycnic coordinate ocean model for the period (1995–2002) considered by (Orvik and Skagseth 2003) but the correlation vanished for the pentad prior to this period.

The directly observed volume fluxes across the Ridge, reported here (Fig. 1.11), are rather short for a comparison to the wind, but for the Iceland branch (Astthorsson et al. submitted) have related the volume flux of Atlantic Water to the wind at Thverfjall in northwest Iceland (Fig. 1.17), indicating that northerly winds reduce the flow of Atlantic water whereas southerly winds increase the flow. This is in accordance with the strong correlation between the spring temperature at Siglunes and the pressure gradient across the Denmark Strait found by Blindheim and Malmberg (2005). This was also suggested by Ólafsson (1999) who reported a significant relationship between hydrographic conditions in spring at the Siglunes transect north of Iceland and the frequency of local northerly/southerly wind directions while he found no correlation with the NAO index.

Thus, there is considerable evidence that variations in the wind stress induce variations in the Atlantic inflow, both as regards the total and individual branches, most clearly seen in the Iceland branch. As noted, however, the variable component of the Atlantic inflow is small compared to the average, whereas the wind stress varies considerably, as illustrated by the seasonal variation. Thus, for the total Atlantic volume flux, the ratio of the seasonal amplitude to its average value is 5% according to the observations

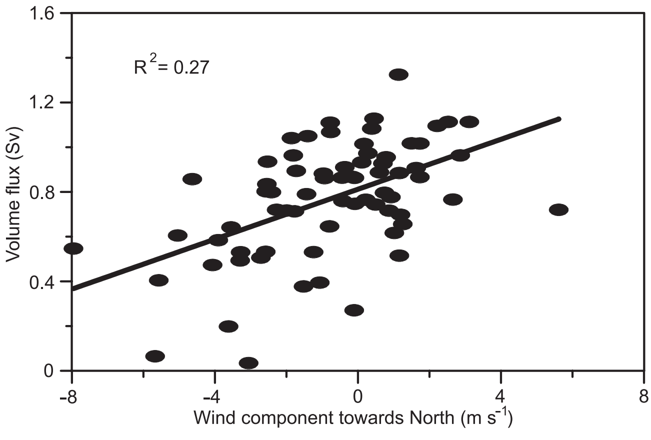

Fig. 1.17 Relationship ($p < 0.01$) between monthly flux of Atlantic water through Denmark Strait and the monthly north–south component of the wind at Thverfjall (Fig. 1.4a), northwest Iceland, for the period 1994–2001. The squared correlation coefficient is indicated

and 14% according to the model. For the wind stress curl averaged over the Nordic Seas, in contrast, this ratio is close to 100% (Jakobsen et al. 2003).

The relative stability of the Atlantic inflow remains also on much smaller timescales than the seasonal. (Hansen et al. 2003) calculated daily Atlantic water volume flux values in the Faroe branch from summer 1997 to summer 2001 and found not a single flow reversal (westward flux) among the 1,348 daily flux estimates. This can be contrasted to the inflow to the Barents Sea, which is much more variable and generally considered to be generated by the wind (Ingvaldsen et al. 2002).

It therefore seems doubtful that wind stress can be the main driving force for the dominant stable component of the Atlantic inflow. This is especially the case for the two main inflow branches but, even for the Iceland branch, wind seems mainly to increase or reduce the volume flux from a basic flow, which is there with no wind (Fig. 1.17) in analogy to the inflow through the Bering Strait (Coachman and Aagaard 1988).

1.8.2 Sea-Level Forcing

In the equations of motion, a water parcel close to the surface is acted on by a force that is proportional to the slope of the surface. Any process that generates a persistent sea-level slope across the Ridge can therefore drive an inflow and two processes within the Arctic Mediterranean can do this (Fig. 1.15). One is the estuarine mechanism (Stigebrandt 2000), which generates the surface outflow. The other is thermohaline ventilation, which generates the overflow. In an alternative terminology, these two processes have been termed positive and negative thermohaline circulation, respectively (e.g. Hopkins 2001).

The outflows, generated by these processes, must be balanced by inflows and the balance has to be maintained on fairly short timescales. Imagine an outflow of ~10 Sv without any inflow. The average sea-level of the Arctic Mediterranean would then sink by ~5 cm a day. This would rapidly establish a sea-level slope across the Ridge. To estimate, how large a sea-level drop is required to drive the observed Atlantic inflow, assume zero initial speed and inviscid flow. This leads to the Bernoulli equation:

$$V^2 = g \cdot \Delta h \tag{1.2}$$

which links the inflow speed V to the sea-level drop Δh across the Ridge. From the observations (Figs. 1.4–1.6), the typical inflow speeds do not exceed 30 cm s^{-1}, which implies a sea-level drop of less than 5 mm. This value is found by ignoring friction, but still, it is so small compared to typical sea-level variations, that this mechanism might seem irrelevant. It remains an inescapable fact, however, that, as long as there is a continuous outflow, this mechanism will turn into effect, if no other mechanism forces an inflow and that it can drive the observed Atlantic inflow with a sea-level drop that is below our observational accuracy.

On the other hand, it is clear, that this mechanism only turns into effect, if no other force maintains an inflow that balances the outflow. The question therefore is, whether there is any evidence for or against this mechanism as an important driver for the Atlantic inflow. An obvious argument against it, is the large variability of the sea-level in the inflow region. From altimetry, the standard deviation of the sea-level is an order of magnitude larger than the 5 mm that are required to drive the inflow.

To investigate this in more detail, a point was chosen downstream of the Ridge. Its location was selected so that it should feel both of the main inflow branches and it was located over the continental slope to keep it relatively unaffected by meandering and eddying. Sea-level height at this point (point A in Fig. 1.18) was then correlated to sea-level height over a wide region (Fig. 1.18a). As could be expected, low correlations were found for the central basins and the Faroe–Shetland Channel, where internal circulation and eddies may dominate, but equation (1.2) is only required to apply when following streamlines and all the upstream inflow region due west of the Ridge was highly correlated to point A. A linear regression analysis, similarly, gave regression coefficients close to 1 (Fig. 1.18b) in this region. This analysis indicates that the typical sea-level drop, as the Atlantic inflow crosses the Ridge, is not as variable as might be expected from a first glance at the altimetry, and it supports the application of equation (1.2).

The next question is, whether sea-level forcing can reproduce established key features of the Atlantic inflow. The discussion above verifies that a sea-level drop of 5 mm across the Ridge should be sufficient to drive the observed total volume flux, but how stable is it? By itself, the pressure gradient generated by a sea-level drop of 5 mm would not seem to be very stable, because an excess inflow of1 Sv would eliminate the sea-level drop in a day. The stability of sea-level forcing, therefore, rests on the stability of the outflows that generate the sea-level drop across the Ridge. As regards the surface outflows, there is little observational evidence on this and, since they are near-surface, variations in wind stress are likely to affect them considerably. Most of the outflow is, however, in the form of

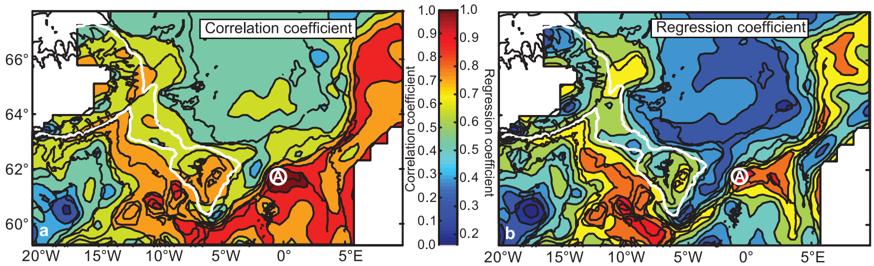

Fig. 1.18 (a) Correlation coefficient between sea-level height at the point A and at other points in the area. (b) Linear regression coefficient (slope) between sea-level height at the point A (y-coordinate) and at other points in the area (x-coordinate). The 500 m depth contour around Iceland and Faroes is enhanced in white to illustrate the Ridge. Based on weekly fields from "The Mapped Sea Level Anomaly (MSLA)" data, produced by the CLS Space Oceanography Division (www.jason.oceanobs.com)

overflow (Fig. 1.14) and, although the overflow has variations, it has been demonstrated to be very persistent (Østerhus et al. 2001; Dickson and Brown 1994).

The overflow stability may be seen in terms of forcing. By generating a barotropic pressure gradient, wind stress can modulate the overflow (Biastoch et al. 2003), but, in addition, there is a pressure gradient at the depth of the overflow, which is generated by the accumulation of dense water in the Arctic Mediterranean (Hansen et al. 2001). This baroclinic pressure gradient is quite clearly responsible for accelerating the main overflow branches to the high speeds ($1\,\mathrm{m\ s^{-1}}$) that are observed.

This point was amply demonstrated by (Biastoch et al. 2003), who, in an idealized model experiment, changed the wind forcing from zero to four times the average observed. With increasing wind forcing, they found a shift in the overflow from east of Iceland to the Denmark Strait, but an essentially constant sum of both parts. They concluded that the total overflow can be changed only by altering the density contrast across the Ridge.

The link between overflow and Atlantic inflow may be illustrated by a simple model (Fig. 1.19). The baroclinic pressure gradient driving the overflow is maintained by the large reservoir of dense water in the Arctic Mediterranean. Even without any renewal, the amount of dense water north of the Ridge is sufficient to maintain an overflow for decades (Hansen and Østerhus 2000), which explains the overflow stability. But, a continuous overflow of 6 Sv will tend to depress the average sea-level of the Arctic Mediterranean by several centimetres each day, which is much more than required to drive the Atlantic inflow across the Ridge. The sea-level forcing will therefore rapidly adjust the volume flux of the Atlantic inflow towards balance.

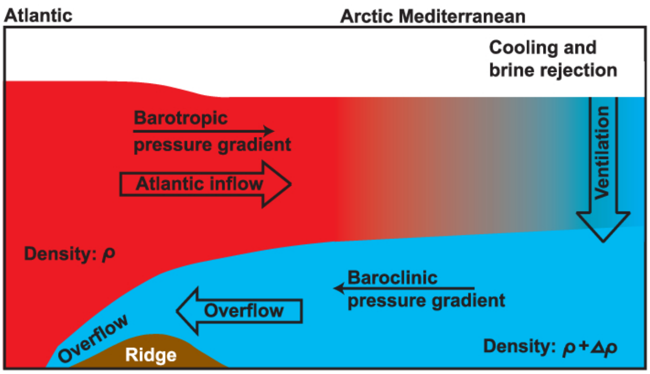

Fig. 1.19 A simple model of the overflow forcing of the Atlantic inflow. Cooling and brine rejection in the Arctic Mediterranean convert the incoming Atlantic water (red) into denser overflow water (blue), which accumulates at depth. The density contrast and sloping isopycnals generate a baroclinic pressure gradient that accelerates overflow water towards and across the Ridge. This removal of water from the Arctic Mediterranean induces a sea-level drop across the Ridge, which reduces the total pressure gradient acting on the overflow slightly (around 10% under present-day conditions) and drives an Atlantic inflow equal to the overflow for stationary conditions. Note that the vertical scales for sea-level slope and isopycnal slope are quite different

The system illustrated in Fig. 1.19 cannot explain all the Atlantic inflow. The surface outflow, associated with the estuarine circulation of the Arctic Mediterranean, and wind stress also contribute, but the fact that most of the Atlantic water returns as overflow (Fig. 1.15) is a clear indication that this is the dominant forcing mechanism and it can explain the relative stability of the Atlantic inflow.

1.9 Conclusions and Outlook

During the last decade, observations and modelling efforts have converged into a consistent description of the properties and intensity of the Atlantic inflow across the Greenland–Scotland Ridge. We know the average temperatures and salinities of the individual branches and have learned to link their decadal variations to the intensity of the Subpolar Gyre. In a series of projects, starting with Nordic WOCE, through VEINS and MAIA, to ASOF-MOEN, we have, for the first time, been able to measure the volume fluxes of all the branches with a relatively high accuracy and the measured average fluxes compare well with those calculated by the ASOF-MOEN model. When considering more rapid variations, the model indicates somewhat larger seasonal flux amplitudes than the observations, but not outside the combined observational and modelling uncertainties. These results highlight the pronounced stability of the Atlantic inflow across the Ridge and indicate that direct wind stress forcing is not likely to be the main driving force for the inflow, although it probably accounts for much of the inflow variability on timescales below a decade.

In the ASOF-MOEN project and its predecessors, an observational system has been established, which allows us to monitor the properties and intensities of all the inflow branches. This system would benefit from additional instrumentation but it can form the backbone of a dedicated monitoring system for the Atlantic inflow. In the coming decades, climate change is going to affect the ocean more and more and such a system will be essential if we are to be able rapidly to identify and quantify any changes to the Atlantic inflow.

Acknowledgements Through the years, a huge observational effort has been spent by the staff on the marine and fisheries research institutes of Iceland, the Faroe Islands, and Scotland and their research vessels. The monitoring system was originally established within the Nordic WOCE project with support from the Environmental Research Programme of the Nordic Council of Ministers (NMR) 1993–1998 and from national Nordic research councils. Later, support has been gained from European research funding programmes in several projects: VEINS, MAIA, and ASOF-MOEN. Continued support has been provided by the Danish DANCEA programme. The authors also wish to thank Beatriz Balino for her dedicated and efficient management of the ASOF-MOEN project.

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