Irminger Water variability in the West Greenland Current

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[1] We examine the historical variability of Irminger Water (IW) along 3 sections across the West Greenland Current over 1950–2005. Significant variability in the salinity, size and position of the IW core are seen over time. Some of the saltiest and warmest IW ever recorded have been seen since 1995 (comparable to previous maximums in the 1960s). During these periods, the volume of IW is also larger, leading to larger transports into the Labrador Sea. For the period 1984–2005 transports at Cape Farewell are 3.8 ± 0.9 Sv, 7.5 ± 2.2 × 10^{13} J and 8.5 ± 1.8 mSv of salt referenced to 35.0. LSW formation is also correlated to IW transport at Cape Farewell with a lag of one year (0.51).

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1. Introduction

[2] The principle source of heat to restratify the interior of the Labrador Sea after convection events is thought to be the input of warm and salty Irminger Water (IW) [*Cuny et al.*, 2002; *Lazier et al.*, 2002]. As a remnant of the sub-polar mode waters that have travelled cyclonically around the sub-polar gyre [*Cuny et al.*, 2002] it enters the Irminger Current along the east coast of Greenland [*Pickart et al.*, 2005] before rounding Cape Farewell and propagating north in the West Greenland Current (Figure 1), along the slope at 200–700 m [*Straneo*, 2006]. *Buch et al.* [2004] discuss how there are two components of this water mass, pure and modified, depending on how much mixing has occurred with surrounding water masses during its transit to West Greenland waters.

[3] *Bersch* [2002] showed temporal variations in the upper 600 m mean salinity between 1992 and 1998 along a section in the Irminger Sea, possibly responding to the North Atlantic Oscillation with a lag of 2 years. *Buch et al.* [2004] showed significant variability in IW at a single station at Cape Farewell, with a shift from pure IW prior to 1970 to modified Irminger Water in 1970–95 (with a potential return of pure IW after 1995). Both *Buch et al.* [2004] and *Stein* [2004] reported similar variability in IW along a section at Fylla Bank. However, *Cuny et al.* [2002] examined the maximum temperature and salinity of the IW in the Labrador Sea during the 1990's and found no clear trends along the West Greenland Current. *Straneo* [2006] found a large difference in the lateral heat transport into the interior of the Labrador Sea between the years of Ocean Weather Station Bravo (1964–74) and a later float data set (1996–2000), which she suggested may have been related to a change in the vertical partitioning of IW. I. Yashayaev (Recent changes in oceanographic conditions in the Labrador Sea, submitted to *Journal of Physical Oceanography*, 2007, hereinafter referred to as Yashayaev, submitted manuscript, 2007) noted a large volume of warm and salty water appearing over the continental slopes of West Greenland during the 2000s, which he thought came from the Irminger Sea. *Hatun et al.* [2005] reported on record high salinities in the inflow to the Nordic Seas and showed that the salinity change was linked to the dynamics of the sub-polar gyre circulation.

[4] Here we examine the temporal variability of IW (both properties and transports) along three sections across the West Greenland Current. Two datasets (not necessarily completely independent) are used to allow us to consider the entire period of 1949 to the present. We attempt to link the observed variability to variability elsewhere within the sub-polar gyre.

2. Data and Methods

[5] The first data set we use is a set of standard sections handled by the Danish Meteorological Institute every year in June–July on behalf of the Greenland Institute of Natural Resources. We focus on the Cape Farewell, Cape Desolation and Paamiut sections (Figure 1). Observations were normally performed annually between 1984 and 2005 (albeit with years missing for each section) on the same 5 repeat stations on each section, but in some years only 3–4 stations were occupied due to multi-year-ice on the inner stations. In most years, the section was performed in late June or early July, but during the period 1984–87, the occupations occurred earlier in spring (March through May). For 1984–87, the analysis is based on bottle data while CTD data is used for the other years. Accuracy is to the second digit on temperature and ±(0.003–0.004) for salinity based on comparison with water samples [*Ribergaard*, 2006]. We carry out our analysis on the top 700 m, the deepest depth common to all years.

[6] The second data set used is based upon a climatological analysis of the Labrador Sea (N. Kulan, and P. G. Myers, Comparing two climatologies of the Labrador Sea: Geopotential vs. isopycnal, submitted to *Atmosphere-Ocean*, 2007). All available stations with both temperature and salinity measurements that were in the Fisheries and Oceans Canada hydrographic climate database [*Gregory*, 2004] prior to 2000 were used. The data (with a precision of two decimal places) was divided into overlapping 3-year running mean triads covering the period 1949–1995. The data was binned into 2.5 degree (south of 55°N) or 5.0 degree boxes (north of 55°N) to provide a first guess for an objective

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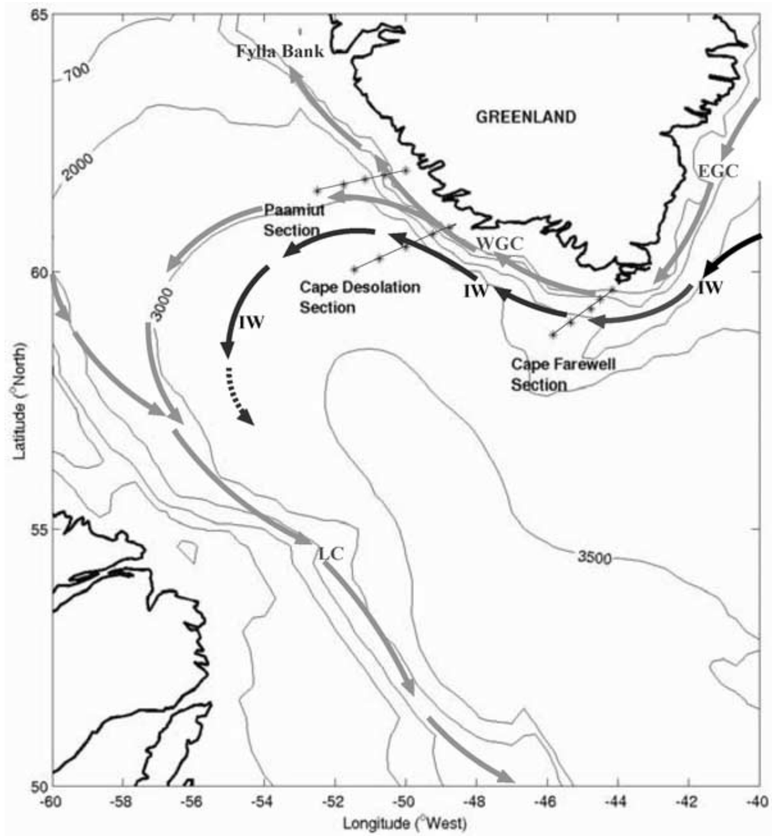

Figure 1. A map of our study region showing the locations of the sections used. Some of the major currents in the region are also indicated. The abbreviations used are: EGC, East Greenland Current; WGC, West Greenland Current; LC, Labrador Current; IW, Irminger Water.

analysis procedure that used three passes with decreasing correlation lengths of 600, 400 and 200 km, weighted by a topographic constraint to minimize mixing of waters across the shelf break. The mapping was carried out in an isopycnal framework using 44 density layers and 1/3 degree spatial resolution. The mapped time-varying triad fields were then interpolated to the location of the 5 stations along each section. Although the mapped triad data is not a unique dataset as it contains the section data, it allows us to extend the analysis much further into the past. Correlations between the triad and section data for Cape Farewell, Cape Desolation and Paamiut are respectively, 0.71, 0.43 and 0.72 for temperature and 0.87, 0.88 and 0.93 for salinity.

[7] We took the data and examined it for each station and each section on a 2 m grid in the vertical (requiring significant interpolation only in the early years with bottle data). To determine average properties, we examined each data point and included all of those where our definition for IW (see below) was satisfied. For transports, we determined the geostrophic velocity, relative to 700 db (or the bottom in shallower water) for each pair of stations at each depth, added an estimate of the barotropic velocity, calculated the resulting transports and then summed those whose T and S (which was interpolated onto the geostrophic velocity point) was consistent with IW. If data was missing, we did not include that point(s) in the calculation. That said, the inshore station is in shallow water with little or no IW present. Since IW is not found above 100 m, issues with the first measurement in a profile not being until 8–14 m should not effect the results. Missing measurements towards the base of a profile may be more significant.

[8] To obtain estimates of the barotropic component of the velocity, a mean spring (April to June) climatology of the Labrador Sea produced in a similar manner to the triads (but using all data collected in the given months available in the Department of Fisheries and Oceans Canada climate database prior to 2000 irregardless of the year collected) was used as input to a regional ocean general circulation model of the sub-polar gyre [*Myers*, 2002] run in diagnostic mode. The component of the barotropic velocity perpendicular to the sections was then used, giving estimates of 5–7 cm/s, which based on *Clarke* [1984], *Pickart et al.* [2005] are probably a lower bound. The use of mean spring data for estimating the barotropic velocities was done so that all the variability would be contained within the baroclinic component based on the hydrography (i.e. the barotropic variability is not included).

[9] A number of definitions of pure and modified IW exist [*Buch et al.*, 2004; *Cuny et al.*, 2002; *Clarke*, 1984; *Reynaud et al.*, 1995; *Ribergaard*, 2006], although all generally consider the water mass to have temperatures and salinities within the range 3.5–6°C and 34.85–35.10. For the purpose of this study, we choose a broad definition

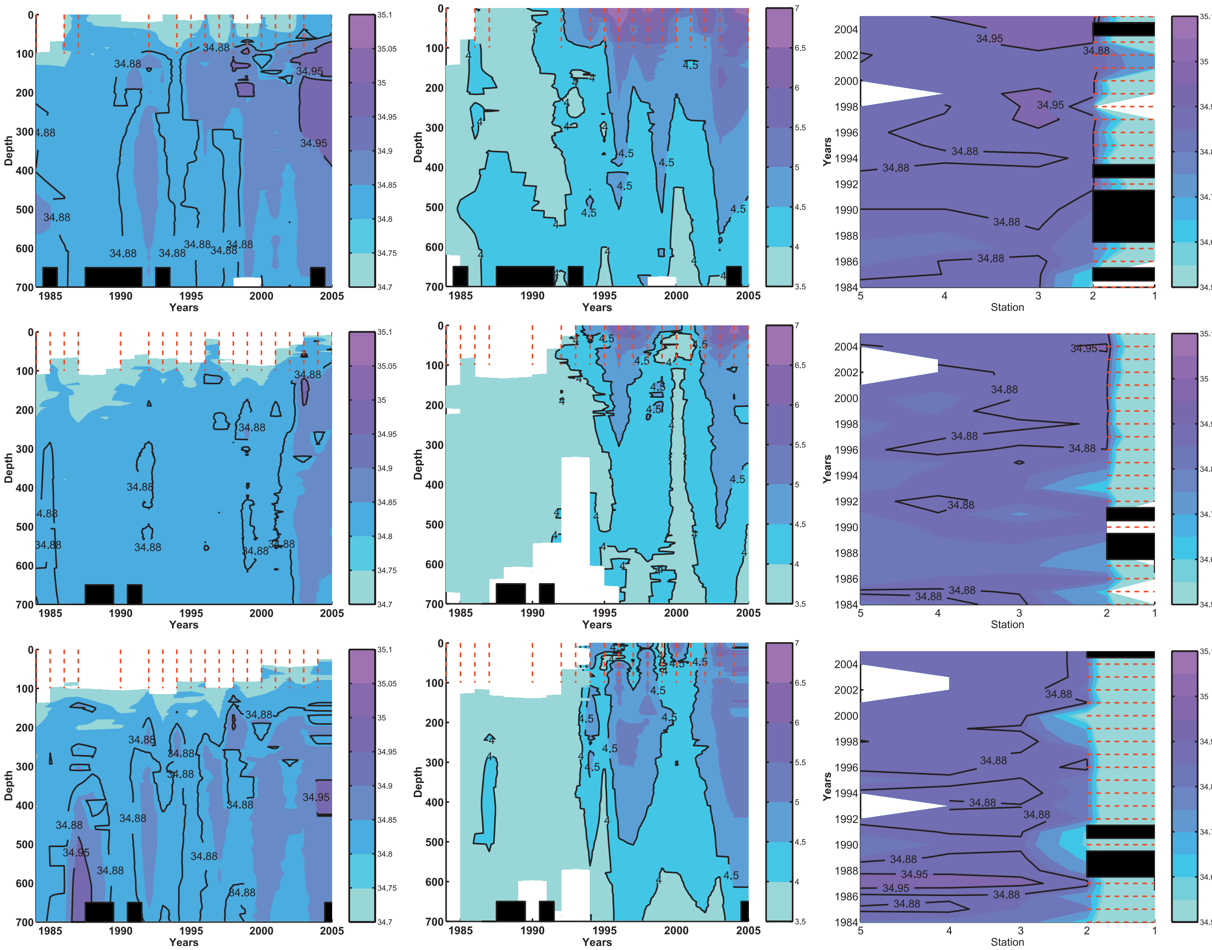

Figure 2. (left column) Time versus depth plots of salinity and (middle column) temperature averaged over the stations on each section. (top) Cape Farewell, (middle) Cape Desolation, (bottom) Paamiut. Grid points with salinities less than 34.7 and temperatures less than 3.5°C were not included in the averages. The right column shows Hovmoller plots of salinity formed by averaging in the vertical all data points with a salinity within 0.1 of the maximum salinity at that station. The red dashed lines indicate the years when the data was collected while the black boxes indicate the years for which no data was available, with these periods filled in by linear interpolation. The white regions indicate where the temperature or salinity at no station along the given section was in the ranges plotted (34.7–35.1 and 3.5–7°C).

including both pure and modified IW, with temperatures >3.5°C and salinities >34.88.

3. Results

[10] Time versus depth plots of temperature and salinity averaged across the stations in each section where IW was present, as well as Hovmoller diagrams of the vertical average of salinity around the salinity maximum at each station are shown in Figure 2. At Cape Farewell, salty and warm IW can be seen at 150–300 m in the early 1990s and post 1995 with two periods of maximum salinity occurring 1997–99 and post 2003. Reduced quantities of IW can be seen during the mid-1980s. Based on the single section per year, we can't differentiate whether this lack of IW is a seasonal feature with the IW not showing up at Cape Farewell until later in the spring or is truly interannual variability, although *Buch et al.* [2004] suggest that IW transport was lower during the 1980s. Similar variability is seen at the other sections although the salinities are generally lower.

[11] Timeseries of mean temperature and salinity for IW (Figure 3) show a trend to saltier (+0.004 per year) and warmer (+0.03°C per year) IW entering the Labrador Sea over 1984–2005. Transports are given in Table 1, increasing post 1995 (albeit with decreased standard deviation). The mean transport of IW at Cape Farewell was estimated as 8.5–11 Sv [*Clarke*, 1984] from a cruise in 1978 while *Pickart et al.* [2005] estimated a transport of 13.6 Sv for the Irminger Current just east of Cape Farewell in 2001. Our transports of IW (3.8 ± 0.9 Sv over 1984–2005) are much smaller than that reported by *Clarke* [1984] but this is not surprising since his estimate is based on a much broader definition of the Greenland slope and is a snap shot from a single hydrographic section.

[12] We present the triad data interpolated to the 5 stations on each section in the same way as the station data (Figure 4). Due to interpolation issues, the front separating the fresh coastal component of the West Greenland Current

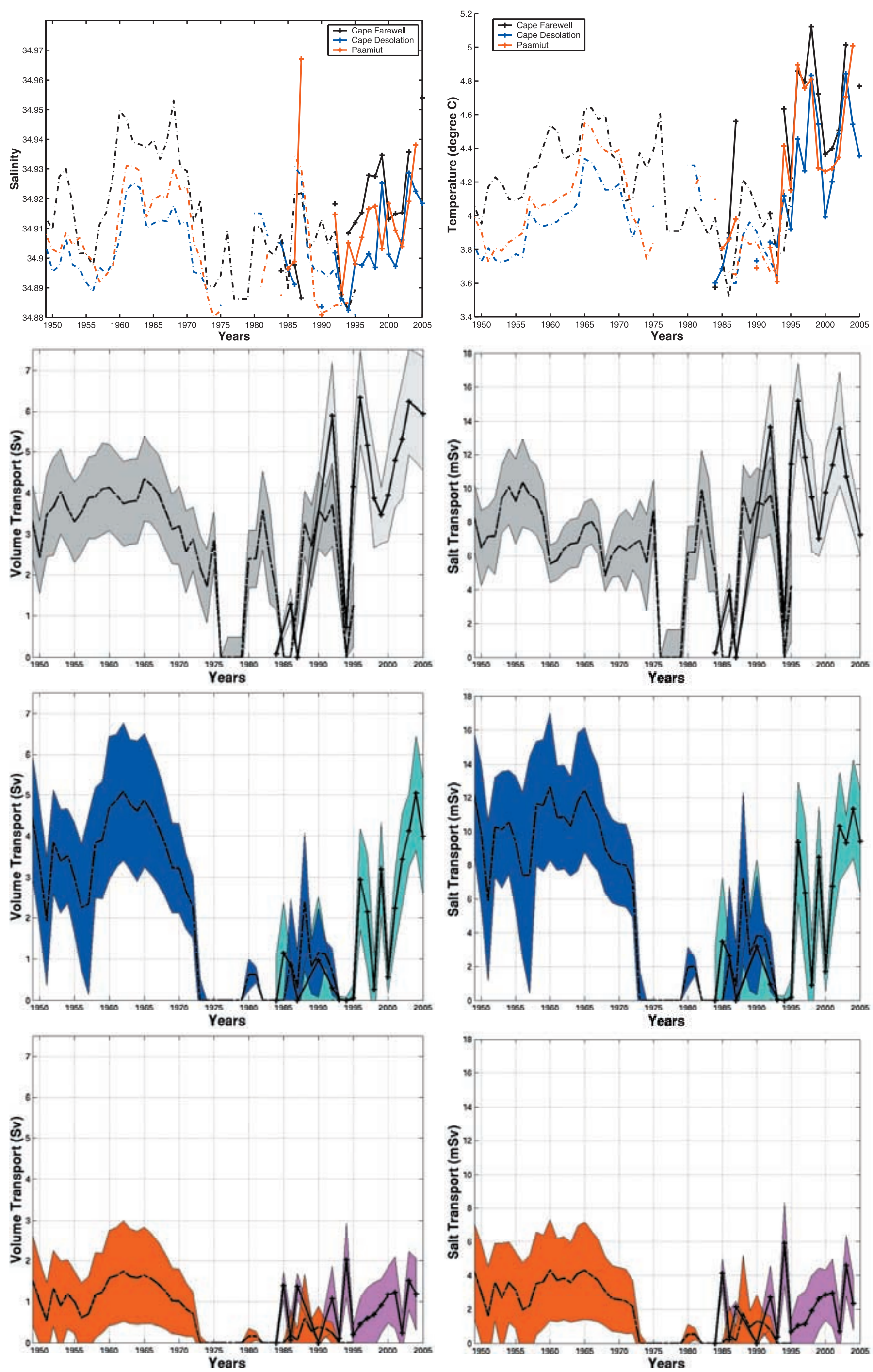

Figure 3. IW timeseries of mean (top left) salinity, (top right) temperature, (left) volume transport, and (right) salt transport (relative to 35.0) for the 3 sections (Cape Farewell, Row 2; Cape Desolation, Row 3; Paamiut, Row 4). A dash pattern is used for the triad data (with uncertainties given in dark gray, blue, and red respectively) while the section data is shown using a solid line (and lighter shades). The uncertainty estimates for transports are based on the given precisions for property measurements and an estimate of 2.0 cm/s for the barotropic velocity.

and the IW is farther offshore in the triad data set. Despite this obvious issue, the basic characteristics of the two timeseries agree for the most part during the periods that they overlap, with very low salinities in the middle 1980s and salinities increasing in the 1990s.

[13] The main core of the IW is in the same 100–200 m range in the triad data, although there are some years with very low salinities and little if any IW present. Local maximums are seen in the early 1950s, the 1960s, early 1980s and the early 1990s, consistent with the results presented by *Stein* [2004]. The depth of the IW core is

Table 1. Volume, Heat, and Freshwater Transports of IW Means With Uncertainty[a]

Section	Volume, Sv			Heat, 10^{13} J s^{-1}			Salt, mSv		
	1984–2005	1949–1995	1995–2005	1984–2005	1949–1995	1995–2005	1984–2005	1949–1995	1995–2005
Cape Farewell	3.8 ± 0.9	3.2 ± 0.6	4.9 ± 1.1	7.5 ± 1.7	5.8 ± 1.2	9.8 ± 2.2	8.5 ± 1.8	8.4 ± 0.4	10.8 ± 2.2
Cape Desolation	2.4 ± 0.9	2.9 ± 1.2	3.1 ± 1.2	4.1 ± 1.6	5.0 ± 2.0	5.3 ± 2.2	6.0 ± 2.5	8.0 ± 3.0	7.4 ± 3.1
Paamiut	1.0 ± 0.6	0.9 ± 0.6	0.8 ± 0.7	1.8 ± 1.1	1.6 ± 1.3	1.8 ± 1.5	2.4 ± 1.5	2.6 ± 2.0	2.2 ± 1.8

[a]References used are 0°C for temperature and 35.0 for salinity. Triad estimates (1949–1995) are averaged only over those years with non-zero IW transport.

deepening through the analysis period. It is also supportive of two different core depths for the IW, a shallow regime around 200–400 m (early 1950s, early 1960s and early 1980s) and a deep regime with a core at depths >500 m (middle 1960s and early 1990s). Mean salinity and temperature similarly vary on a quasi-decadal scale, decreasing from a maximum in the 1960s (Figure 3) although increasing again from the mid-1990s. Transports are comparable to those from the section data (Table 1) except maybe at Paamiut, where too small transports are probably the result of slightly too cold and/or fresh waters just missing the criterion we use for IW.

4. Discussion

[14] The increase in IW salinities seen through the late 1990s agrees with the findings of *Bersch* [2002]. During the most recent years, we find near record high salinities (comparing with a previous maximum in the 1960s) and record high transports (both volume and salt) into the Labrador Sea. This is consistent with the record high salinities that *Hatun et al.* [2005] found at the entrance to the Nordic Seas as well as Yashayaev (submitted manuscript, 2007) who found an increase in high salinity IW in the Labrador Sea over the last few years. Such changes were also seen at Fylla Bank [*Ribergaard*, 2006].

[15] IW is a dense form of sub-polar mode water (SPMW) on its progression around the sub-polar gyre before entering the Labrador Sea [*Talley and McCartney*, 1982; V. Thierry et al., Rapid changes in the properties of the Reykjanes Ridge Mode Water over 1990–2006, submitted to *Journal of Geophysical Research*, 2007, hereinafter referred to as Thierry et al., submitted manuscript, 2007] show that mode waters have become warmer, saltier

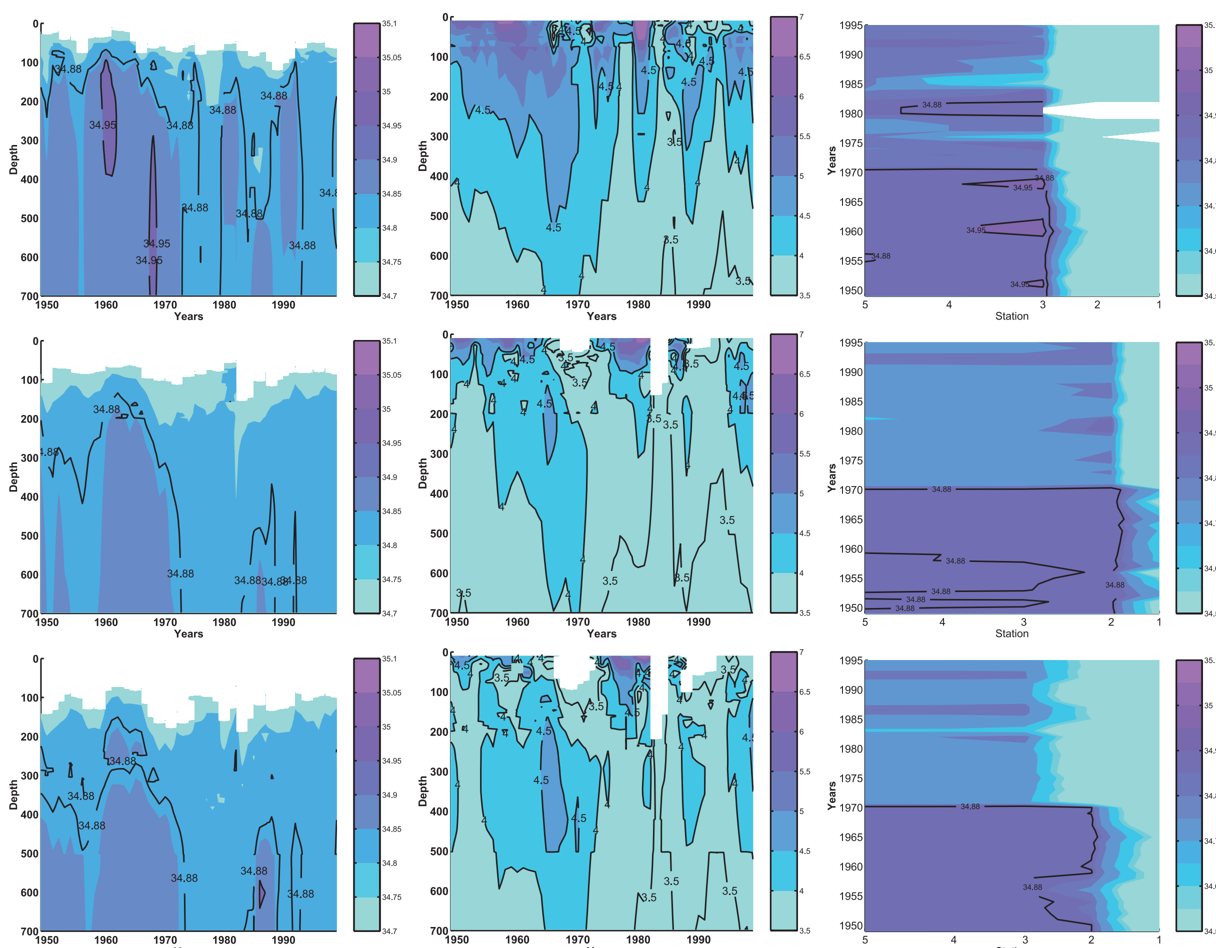

Figure 4. Same as for Figure 2, except using the interpolated triad data.

and lighter since the late 1980s in the Iceland Basin and along Reykjanes Ridge, as we find farther downstream. Observations suggest that the sub-polar gyre circulation has weakened over this same period [*Hakkinen and Rhines*, 2004; *Hatun et al.*, 2005], leading to changes in the North Atlantic Current transport and thus the salinity in the north–east Atlantic, which *Thierry et al.* [submitted manuscript, 2007] then suggested fed back upon the mode water properties.

[16] Besides impacting the circulation of the sub-polar gyre [*Curry and McCartney*, 2001], the low (high) phase of the North Atlantic Oscillation (NAO) can be associated with strong (weak) heat losses in the eastern basin [*Hurrell et al.*, 2003] leading to an increased (decreased) production of SPMW [*Joyce et al.*, 2000]. Such forcing also plays a role in driving IW variability. We use the indices produced by the National Oceanic and Atmospheric Administration/National Weather Service Climate Prediction Center (CPC). Considering only the years with non-zero IW transport in our data, the correlation coefficient between the volume transport and the winter (JFM) NAO index is maximum at a lag of one year, at -0.42 (-0.52 for the triad data prior to 1995), significant at the 99% level.

[17] The small decreases in salt transports between the Cape Farewell and Cape Desolation sections show that little of the IW is transported offshore into the Labrador Sea in this region, with much of this exchange occurring between the Cape Desolation and Paamiut sections, consistent with estimates of high eddy kinetic energy and eddy fluxes in this region [*Jakobsen et al.*, 2003]. Taking the export into the interior of the Labrador Sea (taken to have an area of 10^6 km^2) as simply the difference in transports between the Cape Farewell and Paamiut sections and assuming, for lack of better estimates, that our snapshot estimates are representative for the whole year, we find mean lateral exchanges of heat and salt into the Labrador Sea due to IW for 1984–05 of 5.7×10^{13} J/s and 6.1 mSv. Larger values (8.0 and 3.8×10^{13} J/s; 4.9 and 8.0 mSv) are found for 1995–05 and 1949–95.

[18] These estimates agree with the contention of *Straneo* [2006] that the mean annual heat loss to the atmosphere over the central Labrador Sea of 1 GJ m^{-2} is balanced (and exceeded) by the subsurface transport of heat by IW. The larger IW heat transport to the interior of the Labrador Sea during the 1990s/2000s is also consistent with the increase in heat content seen during this period [*Lazier et al.*, 2002; *Straneo*, 2006]. *Myers and Donnelly* [2007] estimated LSW formation using a water mass formation approach and interannually varying fluxes and surface water properties. We find the maximum correlation between their estimates of dense LSW formation and the transport of IW at Cape Farewell to be 0.51 over the years 1960–1995, with a lag of one year, which may suggest that IW plays a role in providing salt to the Labrador Sea to drive convection [e.g., *Lazier et al.*, 2002; *Straneo*, 2006].

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