Irminger Water Variability in the West Greenland Current

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ABSTRACT: Two data sets are used to examine the historical variability of Irminger Water along 3 sections across the West Greenland Current over the last 50 years. Significant variability in the salinity, size and position of the IW core are seen over time. The section data suggest that for the period 1984-2005, the transports are 2.8 ± 1.8 Sv, $5.5 \pm 3.2 \times 10^{13}$ J and 6.4 ± 3.4 mSv of salt referenced to 35.0. Transports have increased since 1995, with some of the saltiest and warmest IW ever recorded (comparable to previous maximums in the 1960s) being seen in recent years. The majority of this heat and salt is exported into the interior of the Labrador Sea between the Cape Farewell and Paamiut sections. The variability in IW is suggested to be driven by variability in sub-polar mode water formation east of the mid-Atlantic Ridge and is strongly anti-correlated (-0.51) with the North Atlantic Oscillation at a lag of one year. A similar correlation (0.51) and lag (1 year) are found between the amount of IW passing Cape Farewell and the formation of LSW for the period 1960-95.

1. Introduction

Strong wintertime buoyancy loss over the Labrador Sea leads to formation of Labrador Sea Water (LSW) through deep convection [*Lazier et al.*, 2002]. This water mass is then exported throughout the sub-polar North Atlantic [*Sy*, 1997; *Talley and McCartney*, 1982] as well south in the deep western boundary current as part of the lower limb of the meridional overturning circulation [*Talley and McCartney*, 1982]. Many authors have shown significant variability in convection depth and formation rate (e.g. Lazier *et al.* , 2002) over the past 50 years. This variability has been linked to changes in atmospheric buoyancy forcing [*Straneo*, 2006] as well as to periods of significant upper ocean freshwater input (e.g. Belkin *et al.* , 1998).

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, 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 watermasses during its transit to West Greenland waters.

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]; *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 partioning of IW. *Yashayaev* [2006] 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.

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

The first data set we use is taken from the International Council for the Exploration of the Sea (ICES) database. We focus on the Cape Farewell, Cape Desolation and Paamiut sections (Fig. 1) based on annual occupations of the sections between 1984 and 2005 (albeit with years missing for each section). Observations were normally performed annually 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 $\pm 0.003 - 0.004$ for salinity based on comparison with water samples. We carry out our analysis on the top 700 m, the deepest depth common to all years.

The second data set used is based upon a climatological analysis of the Labrador Sea [*Kulan and Myers*, 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 was divided into overlapping 3-year running mean triads covering the period 1949-1995. Each triad was defined to include all the data collected in a given year, as well as all available data in the preceding and following year. The data was binned into 2.5 degree (south of 55N) or 5.0 degree boxes (north of 55N) to provide a first guess for an objective 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 isopyncal framework using 44 density layers and 1/3 degree spatial grid for each year. The mapped time-varying triad fields were then interpolated to the location of the 5 stations along each ICES section. Although the mapped triad data is not a unique dataset in comparison to the ICES section data, it allows us to extend the analysis much further into the past. Correlations between the triad and section data are given for each section in table 1 and show solid agreement, suggesting that the triad data is suitable for extending the analysis.

The data (from both data sets) was used to calculate geostrophic, baroclinic velocities for the flow between the inner and outer station on each section, relative to 700 db. Note that the station positions are effectively the same each year and although in some years the inner stations were not occupied, they are in shallow water on the shelf where IW is not observed. 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 DFO 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 resulting model barotropic velocities were then interpolated to the Cape Farewell stations. Although without doubt a gross simplification, the calculated barotropic velocities of 5-7 cm/s are better than assuming no velocity at 700 m. The use of mean summer data for estimating the barotropic velocities through the diagnostic model was done so that all the variability would be contained within the baroclinic component based on the hydrography.

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 3.5-6C and salinities of 34.85-35.1. For the purpose of this study, we choose a broad definition including both pure and modified IW, with temperatures > 3.5C and salinities > 34.88.

3. Results

Time versus depth plots of temperature and salinity averaged across the stations in each section where IW was present, as well as hovemuller 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. Little 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 truely interannual variability, although *Buch et al.* [2004] suggest that IW transport was lower during the 1980s. Similar variability is seen at Cape Desolation although the salinities are generally lower. Significant IW can also be seen at Paamiut, effectively in phase with that at Cape Farewell. Additionally, a deep high salinity maximum can be seen at depths greater than 500 m in 1986-88 at Paamiut (with hints of this feature several years before at Cape Farewell). However, there is no signature in temperature and with a density > 27.75, this is probably Labrador Sea Water.

Timeseries of mean temperature and salinity for IW (Fig. 3) show a trend to saltier (+0.004 per year) and warmer (+0.03 C per year) IW entering the Labrador Sea over the period the section data is available. Transports are given in table 2. We see that an average of 2.8 ± 1.6 Sv of IW transporting $5.5 \pm 3.2 \times 10^{13}$ J of heat and 6.4 ± 3.4 mSv of salt entered the Labrador Sea over 1984-2005, with transports increasing post 1995 (although with decreased variability). 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 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. During 1995-2005, the transport of IW at Cape Desolation is actually larger than at Cape Farewell. We suspect this is because the Cape Desolation section extends farther offshore and thus more of the IW transport is measured.

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

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 in the section. Local maximums are seen in the early 1950s, the 1960s, early 1980s and the early 1990s, consistent with *Stein* [2004]. The depth of the IW core may also be slightly deepening with time. 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) - not shown. The high salinity water of density > 27.75 seen at depth on the Cape Farewell section in 1985 is also present in the triad analysis (and is the only year when densities reach such a value in the top 700 m). Mean salinity and temperature similarly vary on a quasi-decadal scale, decreasing from a maximum in the 1960s (Fig. 3). Transports are comparable to those from the section data (Table 2) 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. Summary and Discussion

Our results, consistent with previous work [*Buch et al.*, 2004] show that there is significant variability in time with respect to the amounts and the properties of IW that enter the Labrador Sea. Strong transport coinciding with high salinities and temperature are generally seen through the 1960s as well as the 2000s. Fresher and colder IW is seen during the intervening period, although small high salinity cores are seen during some years. Transport of this lower salinity mode of IW is variable, being significant in some years while in others very little if any IW is transported into the Labrador Sea. The depth of the core of the IW also seems to vary with time, from high up in the water column (200-400 m) to deeper (400-700 m).

The fairly high IW salinities seen through the late 1990s agrees with the findings of *Bersch* [2002] as well as *Cuny et al.* [2002]. The increase in salinity and slight decrease in depth of the IW core at the outer stations in the 2000s also agrees with a similar anomaly seen by *Bersch* [2002] in the A1E section starting around 1998. We find near record high salinities (comparing with a previous maximum in the 1960s) and record high transports (both volume and freshwater) 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* [2006] who found an increase in high salinity IW in the Labrador Sea over the last few years. Such changes were also seen at Cape Desolation and Fylla Bank [*Buch et al.*, 2004].

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]. *Thierry and Mercier* [2006] show that mode waters have become warmer, saltier 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 and Mercier* [2006] then suggested fed back upon the mode water properties.

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 -0.41, significant at the 99% level. If only the triad data set is considered (i.e. prior to 1995), the correlation coefficient increases to -0.51.

The small decreases in 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 in this region [*Cuny et al.*, 2002; *Heywood et al.*, 1994]. 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 implied changes in heat and salt content for the Labrador Sea due to IW (Table 3). 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 GJm^{-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 using a water mass formation approach, NCEP surface fluxes and interannually varying surface water properties. We find the maximum correlation between their classical LSW formation rates and the transport of IW at Cape Farewell to be 0.51 over the years 1960-1995, with a lag of one year, consistent with the idea that IW plays a key role in providing salt to the Labrador Sea to drive convection (e.g. [*Lazier et al.*, 2002; *Straneo*, 2006]). We would probably expect the correlation to increase if we were able to consider the period 1995-present, with the increased IW seen in recent years leading to increased LSW formation.

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Table 1. Correlations, significant at the 99% level, between the triad analysis and the ICES section data for each section. Correlations are calculated over all temperature (T) and salinity (S) values as well as over a smaller subset of the data associated with the off-shelf waters.

Section	All T and S		T>3C and S>34	
	T	S	T	S
Cape Farewell	0.71	0.87	0.65	0.53
Cape Desolation	0.43	0.88	0.62	0.83
Paamiut	0.72	0.93	0.49	0.82

Table 2. Volume, heat and freshwater transports of IW across each section - means with standard deviations. Volume transports are given in Sv, heat transports in $10^{13} Js^{-1}$ and the apparent salt flux across the section, in mSv. Heat transports are calculated with respect to a reference temperature of 0C while salt fluxes are computed with respect to a reference salinity of 35.0. Triad estimates (1949-95) are averaged only over those years with non-zero IW transport.

Section	Volume			Heat			Salt		
	1984-05	1995-05	1949-95	1984-05	1995-05	1949-95	1984-05	1995-05	1949-95
Cape Farewell	2.8 ± 1.6	3.6 ± 0.7	3.2 ± 0.8	5.5 ± 3.2	7.2 ± 1.8	5.8 ± 1.7	6.4 ± 3.4	8.1 ± 1.5	8.4 ± 1.8
Cape Desolation	2.5 ± 2.5	3.9 ± 2.4	2.9 *pm* 1.5	4.7 ± 4.9	7.4 ± 4.7	5.0 ± 2.7	6.7 ± 6.2	10.3 ± 5.6	8.0 ± 3.6
Paamiut	1.1 ± 1.2	1.3 ± 1.3	0.9 ± 0.5	2.1 ± 2.4	2.6 ± 2.8	1.6 ± 0.9	2.8 ± 2.8	3.2 ± 2.9	2.6 ± 1.3

Table 3. Mean lateral exchange of heat and salt into the Labrador Sea between the Cape Farewell and Paamiut sections. Heat transports are in $10^{13} Js^{-1}$ and the salt flux in mSv.

Heat			Salt		
1984-05	1995-05	1949-95	1984-05	1995-05	1949-95
3.4	4.6	3.8	3.6	4.9	4.5

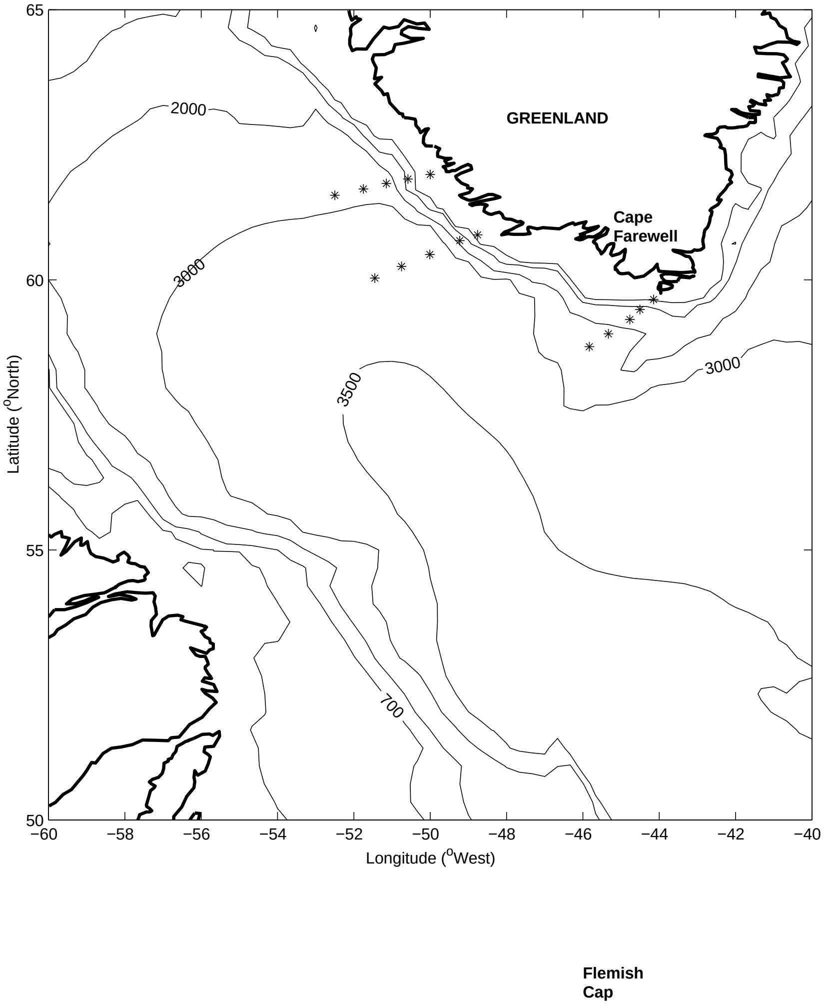

Figure 1. A map of our study region showing the locations of the sections used. Some of the major currents in the region are also indicated.

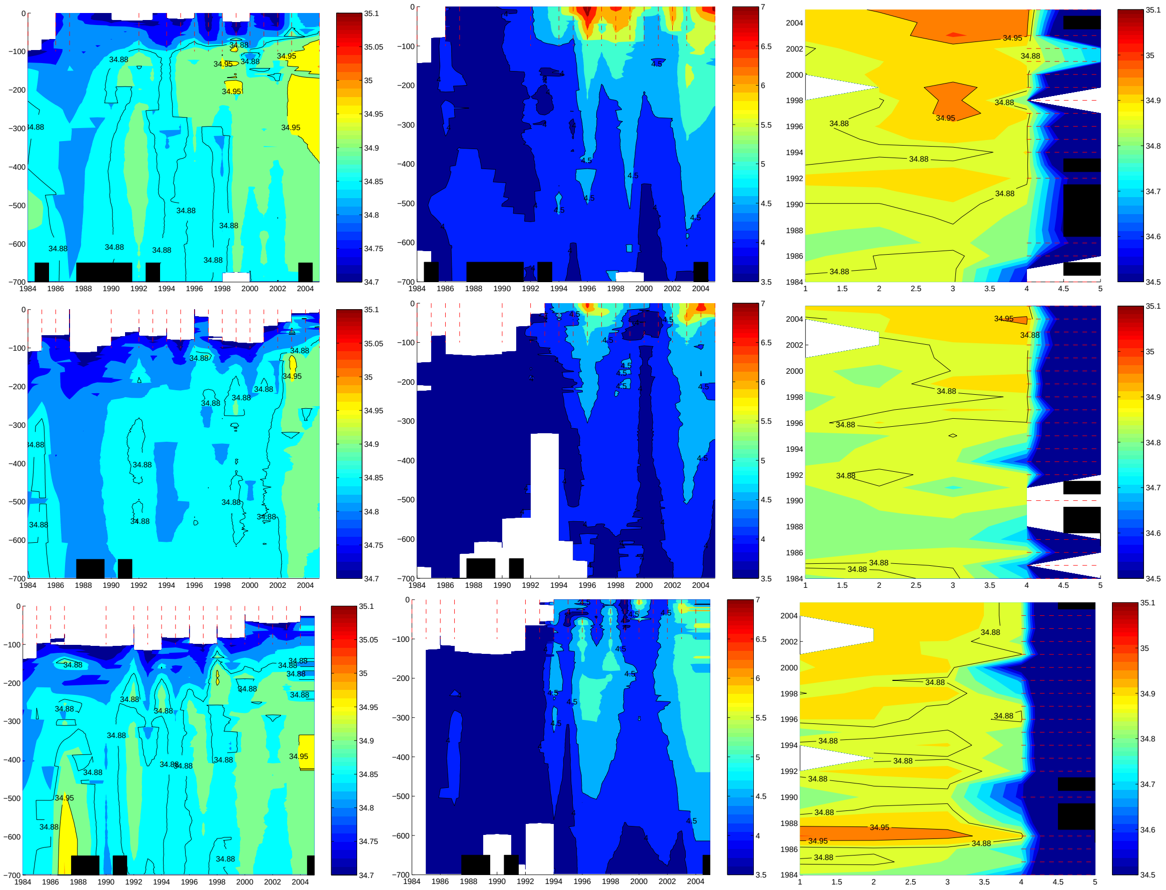

Figure 2. Time versus depth plots of salinity (left column) and temperature (middle column) averaged over the stations on each section (top - Cape Farewell, middle - Cape Desolation and bottom - Paamiut) and hoveuller plots of salinity (right column) formed by averaging in the vertical all data points with a salinity within 0.1 of the maximum salinity at that station. The coast is to the right on the hovemuller plots. The red dashed lines indicate the years when the data was collected while the the black boxes indicate the years for which no data was available, with these periods filled in by linear interpolation. The unshaded regions indicate regions 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 °.

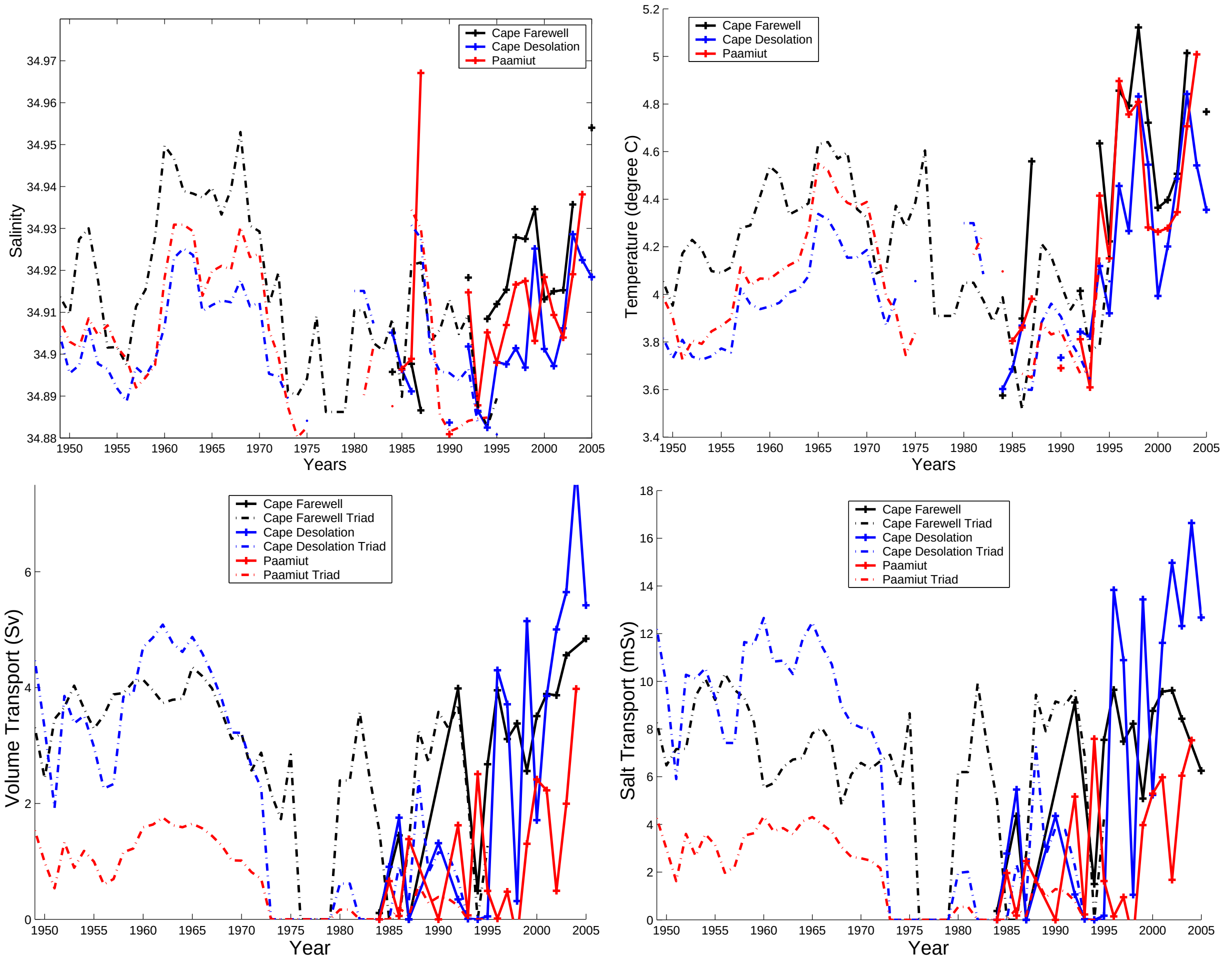

Figure 3. Timeseries of mean salinity of IW water (top left), mean temperature of IW water (top right), volume transport of IW water (bottom left) and salt transport (relative to 35.0) of IW (bottom right) for the 3 sections. A dot-dash pattern is used for the triad data while the section data is shown using a solid line.

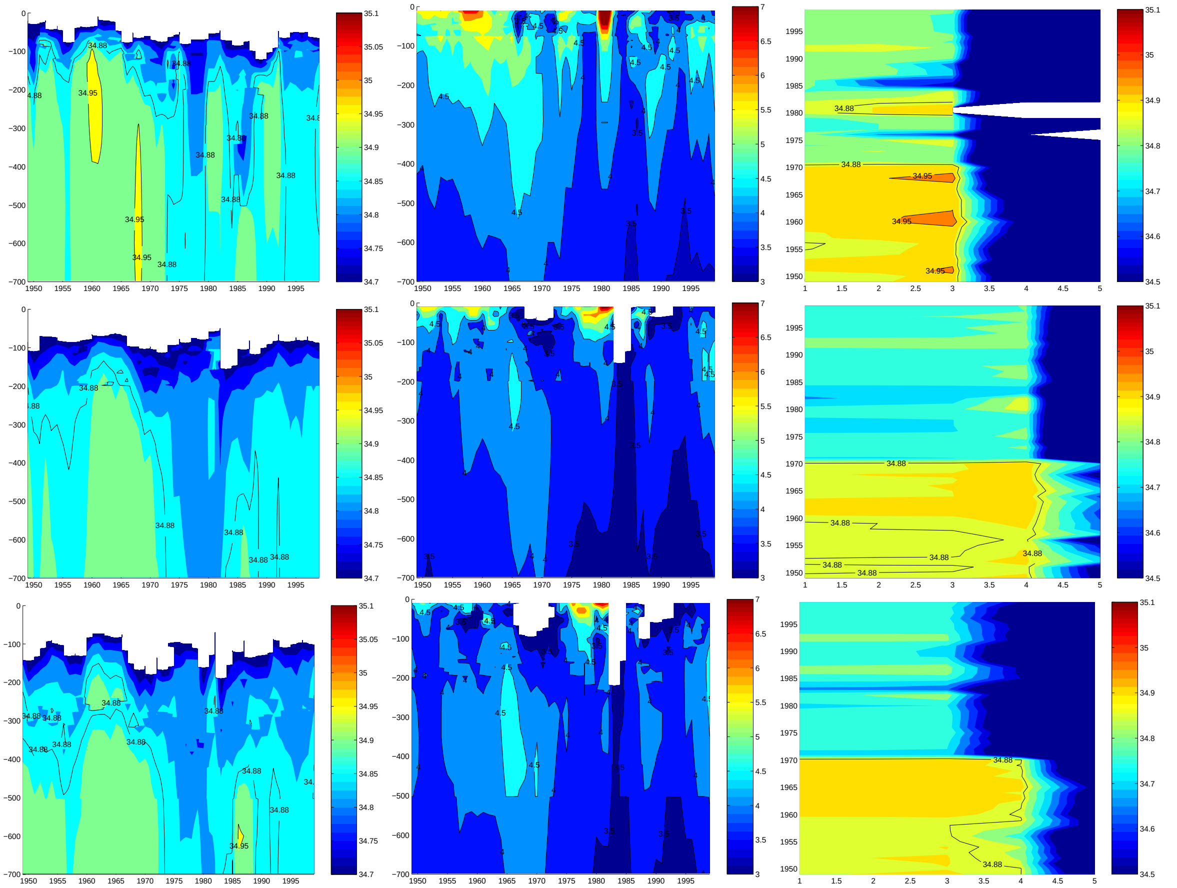

Figure 4. As per figure 2, except using the interpolated triad data.