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Geostrophic Circulation and Heat Flux Across the Flemish Cap, 1988-2000[1]

by

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ABSTRACT

Historical NODC, and OISST data together with CTD data from bottom trawl surveys preformed in the Flemish Cap were used to analyse geostrophic circulation over the bank and heat flux across the 47º line. The recurrence of the casuistic of anticyclonic circulation around Flemish Cap in July permits us to infer that the topographic constraint in terms of the bank's location and dimension primarily determines dynamics over the bank. A coherent cold flow skirts through the northeastern flank with enhanced southerly geostrophic velocities of ~7 cm s-1 and partly recirculates around the southern and southwestern flanks in a mean poleward flow of ~3 cm s-1 to isolate a central anticyclonic core. The gyre bears warmer and less saline waters than its surroundings anchored on the topography of the bank. The most significant source of variability of the water masses over Flemish Cap was linked to the variability of the advective flows, namely the offshore branch of the LC and oscillations of the NAC's north wall. The series of geostrophic heat flux anomaly was also estimated to be well balanced and in the order of 2.3 TW in each direction, although the long-term trend of the heat flux series seems to be slightly directed towards a net shift from positive (poleward) in the late 80's to slightly negative (equatorward) net heat flux in the second half of the 90´s. This was attributed to the southward off-shore branch of the Labrador Current over the bank. An approach was made by attempting to examine linkages between the shift in the Coastal Slope Water System and the NAO. In any case, enhanced Labrador Current over the 1995-2000 quinquennuim was observed to strengthen the anticyclonic gyre anchored on the topography of the bank.

INTRODUCTION

The area off eastern Canada is located under the influence of the large buoyancy driven coastal Labrador Current (LC) extending from the west coast of Greenland to the Middle Atlantic Bight. The LC, a combination of the Baffin Land Current and the West Greenland Currents transports about 5 Sv (1 Sv = 106 m3·s-1) (Petrie and Anderson, 1983) southwards closely following the slope of the Peninsula of Labrador (Trites et al., 1982). When the LC gets to the vicinity of the Grand Banks it manifests as a shelf edge phenomenon (Hill et al., MS 1973); its flow diverts into two branches (Smith et al., 1937); the inshore branch remains confined to the coast surrounding the Newfoundland coast through the Avalon Channel towards the East to form the Cape Breton Current; the offshore branch of the LC carries 90% of the transport (Greenberg and Petrie, 1988) following the north-eastern shelf-break of the Grand Bank (Trites et al., 1982). On reaching Flemish Pass, a southward sub-branch flows through the pass while the westward

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sub-branch skirts Flemish Cap around its north north-eastern slope (Smith et al., 1937). The offshore branch is warmer, deeper and more rapid than the inshore one. Interannual changes of the offshore branch are common and associated with variations in the West Greenland Current (Heywood et al., 1994).

Climatological analysis shows intense temperature and salinity gradients that delineate a front between cold, fresh and warm, salty water in the region (Lozier et al., 1995). At the southern part of the Grand Bank these gradients mark the boundary between the Slope Water (derived from 3 Sv of maximum LC transport along the shelf-break (McLellan, 1957)) and the northward-flowing Gulf Stream (GS). Around the tail of the Grand Banks of Newfoundland both flows meet in the NAC front giving rise to the North Atlantic Current (NAC). The NAC carries 30 Sv of GS and LC waters towards the European basins between Portugal and Greenland, and plays a major role in the regulation of the European climate (Krauss et at., 1987). Further north and east the thermohaline gradient is preserved by the contrast between the Mixed Water (Worthington (1976), derived from the southward flow of the Labrador Current) and waters carried by the NAC. At the 31.85 sigma-1 level (below 300 m) the NAC shows strong convergence near 45ºN in the vicinity of Flemish Cap, where a southward flowing Labrador Current partially feeds into the northward flowing NAC.

Lozier et al. (1995) showed that large temperature variability also occurs in this region, which is attributed to strong temporal meandering of the NAC. In this sense Russian studies pointed to strong variability of NAC behavior (Baranov and Ginkul, 1984; Fofonoff and Hendry, 1985), although on average it circulates parallel to the 4000 m isobath (Krauss et al., 1987) with both strong topographic control and baroclinic effects (Heywood et al. (1994). The variability of the current system south of the subarctic front seems to be closely related to meanders and eddies near the NAC front (Krauss et al., 1987). Peak values of eddy kinetic energy (EKE) have been found in deep waters (4000 m) from the NAC extension flowing around Flemish Cap (Heywood et al., 1994), where intensive mixing of waters can be inferred. Near the Newfoundland Rise a quasi-permanent meander forces the NAC to manoeuver its way some 50-150 miles off the bank, a dynamic low being formed between the NAC-LC current systems (Robe, 1974), in the middle of which the Flemish Cap Bank is found.

Flemish Cap is a plateau of approximately 200 km radius situated in NAFO Division 3M, centered at 47ºN 45ºW (Figure 1). It is separated from the Grand Bank of Newfoundland by the ca. 1100 m deep Flemish Pass, which relatively isolates Flemish Cap (Akenhead, 1986; Colbourne, 1997; Gil et al., 1998). Bathymetry ranges between 125 m in the central part of the bank and 700 m around its edge. The characteristic slope is very steep at the southern half of the plateau, where depth increases from 200 to 1000 m over a distance of 20 miles, whereas the slope increases gradually northwards before reaching Flemish Pass.

Flemish Cap is located within an area of transition between cold subpolar waters, influenced by fluctuations in the Labrador Current and warm temperate waters, influenced by fluctuations in the NAC to the south. Labrador Current Slope Water flows from the north with temperatures < 2ºC and salinity < 34.3 psu, and North Atlantic Current Water flows from the south with temperatures > 4ºC and salinity > 34.8 psu. The boundary condition the Grand Banks exert on circulation and the permanence of the NAC and LC system permits the area to be well understood in terms of average dynamics (Robe, 1974).

Previous studies based either on geostrophy (Kudlo and Borokov, 1975; Robe, 1974; Borokov and Kudlo, 1980), drifting buoys (Ross, 1981) or ADCP measurements (Colbourne and Foote, 2000) show that anticyclonic circulation is a quasi-permanent feature over Flemish Cap. The quasi-permanent anticyclonic vortex appears as a topographically generated Taylor column over the bank, as shown theoretically by some authors (Huppert, 1975; Huppert and Bryan, 1976) and experimentally by others (Taylor, 1923; Davis, 1972). From observations made in 1977-1982 Kudlo et al. (1984) observed the occurrence of four types of geostrophic flow, of which the most common situation (67%) happened to be made by anticyclonic patterns following quiescence of wind stress activity, whereas the passage of major storm events and associated wind forcing breaks stratification and contributes to gyre attenuation and the appearance of meandering flows across the bank, or leads to a series of intermediate dynamic situations. Average current speed of 10 cm/s and a nominal width of 200 km have been reported elsewhere.

The anticyclonic gyre imposes homogeneous water properties on waters over Flemish Cap. They are often referred to as mixed waters of LC and NAC (Hayes et al., 1977; Anderson, 1984). Several papers report the existence of subsurface temperature minima below the summer thermocline (Karasyev, 1962; Templeman, 1974; Colbourne, 1993, 1995, 1996 and 1997; Cerviño and Prego, 1997; Gil et al., 1998), associated with seasonal thermohaline fronts and emergence of LC. Akenhead (1986) provided evidence that the Labrador Current was the sole source of water within the upper 100 m range on the Cap itself but the importance of accounting the influence of both LC and NAC was put forward by Greenberg and Petrie (1988). Their barotropic model was purely based on the LC and was seen to underestimate the strength of the clockwise circulation over Flemish Cap. Dynamics of the frontal system associated with the confluence of warm and cold waters is postulated as the driving mechanism that determines circulation dynamics in the region.

Recent studies have pointed out the existence of a coupled slope water system (CSWS) operating in the Northwest Atlantic. The CSWS responds in a similar manner to climate forcing over a broad range of time scales, and two characteristic modes have been identified (Pickart et al., 1999). The maximum mode is characterized by deep and intense convection in the Labrador Sea; a relatively cool, fresh and thick layer of Labrador Sea Water (LSW) is formed; and the volume transport in the Deep Western Boundary Current increases while volume transport in the Labrador Current diminishes. Conversely the minimum mode corresponds to a system state in which convection becomes weaker and shallower in the Labrador Sea, LSW becomes warmer, saltier and thinner and volume transport in the DWBC diminishes while the volume transport in the shallow Labrador Current increases (Dickson et al., 1996; Dickson, 1997; Curry et al., 1998). Although the linkage is not straightforward it is tempting to associate maximum (minimum) modes of the CSWS with positive (negative) phases of the NAO (Pickart et al., 1999). In addition, the position of the Gulf Stream's north wall appears to shift in response to NAO, northward during positive phases of the NAO and southward during negative phases, although the physical mechanisms are not well known (Planque and Taylor, 1998).

METHODS

Climatological maps were presented based on individual hydrographic stations taken along the Grand Banks of Newfoundland from 1900-1998. The data were obtained from the World Ocean Database 1998, version 2.0, January 2000, prepared by the Ocean Climate Laboratory from the National Oceanographic Data Center (NODC, 2000). In order to evaluate mean summer conditions, only profiles acquired between June-August were included. Data were quality-controlled to eliminate erroneous or badly referred data. Data processing protocol basically followed Lozier et al (1995), who had already worked with the NODC database.

Monthly,optimally interpolated SST (OISST, source: http://ingrid.Ideo.columbia.edu/SOURCES/.IGOSS/.nmc/. monthly/) data were used to create monthly time series of SST anomalies from July 1988 to July 2000 at various locations in the western North Atlantic. Regional maps of July OISST anomalies for several years were also produced.

Hydrographic data presented here came from SBE 19 and SBE 25 CTD profiles acquired during the series of annual bottom-trawl surveys carried out in Flemish Cap on board R/V Cornide de Saavedra in July from 1988-2000. Cruises ran for between 10-15 days during which approximately 110 sampling stations per cruise with mean station separation of 10 nautical miles were randomly stratified following standard methods described in Vázquez (2000). Reports of the hydrographic conditions in Flemish Cap as observed during these surveys are published elsewhere and are available via web query (www.nafo.ca/publications). For technical reasons datasets corresponding to 1992 and 1994 were not accessible and could not be included in this analysis. A total of 1296 stations were included.

For each station the data were linearly interpolated in the vertical to project a property value onto a set of pressure surfaces. To avoid sparse distribution in space, data were objectively analyzed with a simple isotropic univariate interpolation method. An objective technique with scale separation developed by Doswell (1977) and Maddox (1980) based on the filtering properties of the objective analysis scheme developed by Barnes (1964) was applied. It was based on the values of neighboring data and on an assumed correlation function of the field. The correlation scale was set to 15 km while the number of interpolations was set to 70 to permit an error variance O(0.05) that accounted for previous smoothing of the observations. Additionally, the analysis operated as a low-pass filter, filtering the observed data field to define the analyzed field. The band-pass response of filtering was centered at a wavelength of 40 km. The short wavelength noise was therefore filtered out. The result was a gridded matrix with 0.07º longitude by 0.05º latitude (8x4 km2) cells, vertically interpolated every 10 m.

Dynamic calculations were based upon the dynamic topography field, which was computed upon the 120 dbar reference level. Previous studies of water circulation in Flemish Cap (Kudlo and Burmakin, 1972; Kudlo and

Borovkov, 1975; Kudlo et al, 1974) used the 200 dbar as the reference level for the computation of geopotential anomaly but we chose a shallower level for the sake of fronts definition and to retain information over the bank's shallower part.

The magnitude of interannual variations in the strength of circulation across Flemish Cap was computed along the standard transect that crosses the Flemish Cap over its shallower part along 47ºN (47ºN line) based on near-synoptic July measurements. Geostrophic velocity, T-S anomalies, geostrophic transports and heat flux anomalies were calculated along this line. As in other climate research studies the basic diagnostic tool employed was based on the estimation of anomalies, in the sense of differences from normal. The normal was attributed to the mean fields over 1988-2000. Mean states have been computed based on gridded x-z fields of temperature, salinity and derived quantities from the transect stations along the 47ºN parallel. For geostrophic velocity calculations along this line dynamic topography was computed at all depths down to 400 m or the bottom if at lesser depth with the reference level set at the sea surface, given the relative importance of bottom and mid-depth currents (e.g. Colbourne and Foote, 2000). The time varying part of the geostrophic velocity field (v'(x,z)) was computed by subtracting the average velocity distribution along 47ºN from a section observed during any single cruise. The averaging period was 1988-2000. Geostrophic transports were computed along 47ºN integrating the velocity structure according to the algorithm presented by Unesco (1991):

The temperature (T'(x,z)) and salinity (S'(x,z)) anomaly fields were similarly computed. Thus were also able to compute the time varying part of the heat flux (H') in a way similar to Freeman (2002):

where Cp is the specific heat and ρ the mean density of sea water.

RESULTS

Climatological analysis

The summer regional climatological setting shows that the area east of Newfoundland is characterized by an intense thermohaline contrast that delineates a front between cold, fresh and warm, salty water that extends from the surface well to subsurface layers (Figure 2). To the southwest of the Grand Banks these gradients mark a clear boundary between the Slope Water and the Gulf Stream. Further north and east, the thermohaline gradient is preserved by the contrast between the Mixed Water derived from the southward flow of the Labrador Current (Worthington, 1976) and waters carried by the NAC. At the 31.85 sigma-1 level (below 300 m) the NAC shows strong convergence near 45ºN in the vicinity of Flemish Cap, where a southward flowing Labrador Current partially feeds into the northward flowing NAC.

Isotherms and isohalines follow the bathymetry, a fact which gives evidence of topographic control of the climatological flows (Figure 3). Warmer and saline NAC waters typify the area offshore of the shelf-break whereas regions inshore of the continental slope seem to be occupied by Mixed Water. The particular location and topographic characteristics of Flemish Cap force the area of the plateau to be primordially influenced by the Labrador Current, whereas NAC traces seem to remain along Flemish Cap's offshore flank, around which they bend cyclonically. Climatological temperature gradients along 47ºN are ~6ºC at 200 km at sea surface versus more than 9ºC over the same distance at 100 m. Sharper contrast between slope and offshore waters is observed at subsurface levels, which may be a source of baroclinic instabilities that eventually reach inshore regions. The literature reveals the NAC signal through the Flemish Pass and around the fringe of the southern plateau (e.g. Cerviño and Prego, 1997; Gil et al., 1998; Cerviño et al, 1999).

Temperature and Salinity Sections

The upper layer experiences the development of a seasonal thermocline commencing in early May and running until late August when maximum surface temperatures may reach 12-13º. This seasonal cycle is accompanied by a salinity cycle with low salinities in the summer months produced by north-south advection of the Labrador Current and ice melting on the shelf (Colbourne and Foote, 2000). Both cycles in water properties occur in the upper layers. Below 50 m amplitude of both temperature and salinity cycles decreases considerably, while phase increases and temperature and salinity seem to be nearly constant throughout the year (Colbourne and Foote, 2000).

Distribution of mean July hydrographic properties over the 10 years comprised in the 1988-2000 series analyzed along the 47ºN line shows a well developed thermocline partially deepened over the central region of the bank (Figure 4). Below it a zonal temperature gradient is set, with maximum values towards Flemish Pass and minimum temperatures along the easternmost slope. Salinity increases with depth but the bowl-shaped salinity contours reflect homogenisation over the mount's topography, which is indicative of anticyclonic circulation around the periphery of the bank.

Water mass properties are strongly determined by anticyclonic dynamics over the bank. Mean velocities show a narrow, intense southward flow centered at ~44.50ºW with geostrophic velocities O(10) cm/s and a wider but less intense northward flow along the westernmost slope before reaching Flemish Pass. Both streamers are coherent with the existence of a semi-permanent gyre anchored on the topography of the bank, which isolates the central region and imposes strong homogenisation of water masses.

The standard deviation maps of thermohaline properties (Figure 4) show that major variability occurred over the upper 25 m. Variability diminished sharply with depth and below 200 m the mean values had a very small deviation with respect to the single cruise values. There was a saddle of low variability in the central part that gave a measure of the recursive existence of a homogeneous gyre anchored on the topography of the bank. Maximum temperature and salinity variability was located at the periphery of the bank, although it was concentrated at the westernmost part, associated with the northward flow. We infer that ocean advection must play a major role in the net heat flux over Flemish Cap. Colbourne and Foote's (2000) atmosphere-ocean heat flux calculations showed there were strong dissimilarities between in situ temperatures and net heat flux and ocean advection are also proposed as a primary mechanism determining variability of mater mass characteristics in central Flemish Cap.

Contrary to the variability structure of salinity and temperature fields, it is noteworthy that most of the variability associated with the geostrophic velocity is distributed along the easternmost part, i.e. associated with the southward flow. The northward, return flow seems more stable in terms of velocity structure and temporal variability.

Dynamic topographies

The analysis of dynamic topographies shows that water circulation is essentially anticyclonic around Flemish Cap in July over 1988-2000 (Figure 5a, b, c). However, significant year to year differences have been revealed. Circulation patterns may be classified into two flow types: (i) a well-developed anticyclone with a well-defined core anchored on the bank's topography encircled by a coherent flow (ii) noisy circulation patterns whereby the central core is undefined and mesoscale structures occur. Years 1988 and 1989 belong to this second type, where the finer scale analysis revealed a number of mesoscale nuclei distributed over Flemish Cap, although the larger scale pattern showed an anticyclonic trend anyway.

In 1990 a blend of types (i) and (ii) occurred, characterized by a clear anticyclonic circulation but neither a definite gyre core nor a coherent surrounding flow were evidenced. To the contrary, in 1991 a well developed type (i) occurred. A clear central core made up of a number of mesoscale nuclei appeared encircled by anticyclonic flows. This situation was repeated in 1993, when horizontal gradients through the gyre center appeared much more smoothed. The next sampled year (1995) was likewise characterized by two large anticyclonic nuclei torn off by a relative low that ran through the bank from NE-SW.

In 1996 a coherent anticyclonic circulation encircled a central core in which a number of stations bore approximately the same geopotential anomaly. Conversely, in 1997 the bank's dynamic topography appeared split up into two areas: the northern zone was occupied by relative dynamic lows, whereas the southern one bore the highest dynamic height values. The most feasible hypothesis is that the gyre was displaced southwards and was not fully sampled. In this case a meandering flow crossed the bank from NW to SE in a situation already observed by Kudlo et al. (1984). 1998 presented the most conspicuous gyre observed in the 1988-2000 series. Anticyclonic circulation was perfectly identifiable except for the SE area, in which the lack of stations prevented us from determining whether the gyre was perfectly closed. Of the remaining years, 1999 also showed a large high center at the topography of the bank, surrounded by an arch of variable and incoherent lows. 2000 also presented a well developed anticyclonic gyre made up of smaller submesoscale individuals.

Water Masses and Dynamic Topographies

According to the Taylor-Proudman theorem, the quasi-permanent anticyclonic vortex appears as a topographically generated Taylor column over the bank when the flow is fairly homogeneous (Huppert, 1975; Huppert and Bryan, 1976; Taylor, 1923; Davis, 1972). For stratified flows, conservation of Ertel's potential vorticity (e.g. Gill, 1982 p 315) means that the shrinking of vortex lines on approaching a topographic anomaly brings about the generation of anticyclonic circulation over the bathymetric feature. Fluid particles are displaced to the left over the obstacle giving accelerated flow over the left flank where anticyclonic flow associated with the hill enhances the mean flow. Numerical models (Huppert and Bryan, 1976) observe that stratified time-dependent flows interacting with varying bottom topography develop anticyclonic circulation. In the present paper the LC assumes the role of the time-dependent flow and Flemish Cap is the bathymetric obstacle. The development of anticyclonic circulation anchored on the topography of the bank brings about differential seasonal processes of water mass modification with respect to surrounding water masses. In summer in Flemish Cap dynamic isolation of a defined structure versus colder northerly flows may be observed as a significant relative increase of temperature values versus surrounding waters (Figure 4).

The offshore branch of the LC that skirts around Flemish Cap supplies the source waters for the development of the anticyclonic circulation over Flemish Cap. This streamer is warmer, deeper and more rapid that the inshore one that flows through Flemish Pass (Greenberg and Petrie, 1988). Water masses retained within the anticyclonic gyre bear lower salinity values than surrounding masses, especially along the southern tip, where ambient salinity values are influenced by subtropical masses. On the other hand, water mass salinities over the deeper levels associated with the anticyclonic gyre must closely resemble the source waters retained within the gyre since atmospheric effects (solar warming, evaporation and precipitation) are expected to be less accentuated at these levels than over the surface mixed layer. Pressure levels corresponding to both 40 and 120 dbar have been chosen to analyse the water mass field in relation with the circulation scheme described in the previous section, and to discuss the sources of interannual variability.

The well-developed anticyclone sampled in July 1991 corresponds to coherent temperature and salinity distributions (Figure 6). These maps mirror dynamic topography maps. The gyre core at 40 dbar bears higher temperatures and lower relative salinities versus the saltier, warmer surroundings. These properties are maintained down to just above the bank's shallower part at 120 dbar, where a perfectly delimited low salinity nucleus vertically coincided with the high center. This thermohaline situation was repeated in 1998, yet in this case the core temperature was very homogeneous (~ 3.8 ºC) and slightly higher than its surroundings (Figure 7).

The development of the anticyclonic gyre may show variability in relation with variations in stratifications of source flows. The development of seasonal stratification from May through September has recently been evidenced (Colbourne and Foote, 2000). Departures from Taylor-Proudman's constriction of homogeneous flow causes quasi-geostrophic velocities and an anticyclonic gyre independent of the stagnant Taylor column. This situation is reflected in deeper water mass properties. 1991 and 1998 showed good anticyclonic gyre development. Although mid-depth temperatures were slightly warmer in 1998 than in 1991, at 220 dbar around the topography of the bank, the westernmost flank bore warmer temperatures than elsewhere (Figure 8). Both situations are compatible with the generation of anticyclonic circulation over Flemish Cap whereby fluid particles are displaced to the left over the obstacle giving accelerated flow over the left flank where anticyclonic flow associated with the bank enhances the mean flow.

Topographic constraint in terms of the bank's location and dimension primarily determine dynamics over Flemish Cap, but variability of the incoming flows determines the snapshot of circulation. Besides the issue concerning the stratification constraint, variability associated with the baroclinicity, intensity and persistence of LC and NAC over Flemish Cap is inferred to be the key to modulating topographic-steered dynamics over Flemish Cap. Occurrence of NAC-sourced water masses onto Flemish Cap through Flemish Pass and around the fringe of the southern plateau is observed to be a quite recurrent feature in the literature (e.g. Cerviño and Prego, 1997; Gil et al., 1998; Cerviño et al, 1999). Similar to the dynamic topography chart, temperature and salinity plots at 220 dbar in 1988 reveal no well developed anticyclonic gyre situation (Figure 8). This situation may correspond to either the fading out of a pre-existing anticyclonic structure or to a weakly dynamic situation with stagnated masses in the form of small-scale nuclei. Around the SW flank of the bank a coherent structure with high temperature and salinity signals extended.

This may provide a clue that poleward-flowing NAC waters may have entered the region and if there was an anticyclonic gyre, this could have been displaced northeastwards.

In the remaining years characterized by clear anticyclonic circulation (1990, 1993, 1996, 1999, 2000), temperature and salinity distributions follow the general pattern described for 1991 and 1998 (not shown). Although all these years show clear contrast between the homogeneous, low salinity anticyclonic core and the surrounding waters, a positive correlation between the definition of dynamic topography and temperature-salinity distributions at 40 and 120 dbar was observed.

In contrast, in 1988 when the situation was characterized by a number of small anticyclonic nuclei throughout the Cap, no salinity signal could be associated with any dynamic feature (Figure 10), and while at 40 dbar a succession of low and high salinity structures was observed, mean salinities ~ 34.1 were delimited by high salinities ~34.4 to the SW over subsurface levels (120 dbar). Likewise, temperature distribution at 40 dbar also showed a wavy distribution with warm (6ºC) versus cold (3ºC) nuclei. At 120 dbar intense horizontal temperature gradients were associated with an anomalously warm signal to the SW (5-6ºC).

The situation in 1995 was characterized by an anticyclone interrupted by a relative low through the middle in the NE-SW direction . This dynamic picture was mirrored by both the salinity and temperature fields at 40 and 120 dbar, where low salinity, warm waters < 34.0 were torn off by a higher salinity, colder water wedge, Figure 11.

In any year the anticyclonic gyre bears light waters versus the colder, denser eastern masses. Below 60 dbar the NAC-sourced waters are extremely warm versus the low salinity, colder ambient waters, whereby lighter densities are borne for the former (Figure 12). Subsurface (e.g. 120 dbar) densities are quite similar for both the NAC-type and the anticyclonic gyre waters. In the case that both water types are approached along the southern or southwestern flank, the density gradient between these may dwindle, even causing the anticyclonic gyre to die out.

Geostrophic Transports and Heat Flux

Geostrophic transports are presented in Figure 13a. The integration of geostrophic transports was performed over the 47ºN line and restricted to cover the bank feature, in an attempt to account for computations just over Flemish Cap and in order to avoid features outside the bank. The total northward (positive) and southward (negative) transports were computed for each year and provided an estimate of the strength of the north-south component of the gyre circulation around the Cap (Colbourne and Foote, 2000). The graph shows that mean transports were O(0.3) Sv, and that the overall budget is well balanced in terms of average calculations. This assertion is compatible with the notion of semi-permanent anticyclonic circulation reported in the literature, which seems to be a recurrent feature as observed in the July surveys. However there are a number of years in which southward transport exceeded the northward flux (1991, 1995, 1996 and 2000). Conversely, in 1997 and 1998 the net volume transport integrated along 47ºN was directed polewards.

The resulting series of geostrophic heat flux anomaly is displayed in Figure 13b. The mean heat flux is estimated to be well balanced in terms of average calculations and being in the order of 2.3 TW (TerraWatts, 1012 W) in each direction. Opposite to the volume transport estimates, the long-term trend of the heat flux series seems to be slightly directed towards a net shift from positive (poleward) in the late 80's to slightly negative (equatorward) net heat flux in the second half of the 90´s.

What are the causes of this shift in the net heat flux pattern, if any? Presentation of mean July SST anomalies computed over the whole of Flemish Cap from 1988-2000 show a quasi-complete decadal cycle wavelength (Figure 14). SST anomalies reached a minimum of –2ºC in 1991 after which it increased steadily to reach +2ºC in 1999. The notion of positive anomalies of air-heat flux during the early- to mid-1990's together with SST's below the long-term average on Flemish Cap led Colbourne and Foote (2000) to conclude that advection is the primary mechanism determining variability in water mass characteristics in central Flemish Cap. Additionally, the July mean SST anomaly cycle on Flemish Cap is tightly linked to mean July SST anomalies along a location further north (51.5ºN), which is assumed to be more directly influenced by the Labrador Current and where retention processes are not to be expected (Figure 14). The overlapping of both series may allow us to infer that July SST anomalies and heat flux across the 47º N line must be a consequence of the variability of the Labrador Current as being the main incoming flux over the bank. Inverse correlation between the southward heat flux anomaly and mean SST over Flemish Cap has been observed (Figure 15), this association not being so obvious for the northward counterpart of the heat transport. The stronger the southward heat flux, the greater the temperature anomaly and therefore heat transport seems to be associated with intensity of the Labrador Current over the bank. This feature may appear contradictory, since one should expect a linkage between Labrador Current intensity and its particular salinity-temperature anomaly.

The years 1989 to 1994 have been reported as extremely cold years with temperature anomalies that resembled the 'Great Salinity Anomaly' years (e.g. Buch, 1998; Colbourne and Foote, 2000). However, the effect upon sea surface salinities was not so easily identifiable, being an index of little freshwater input across the Strait of Denmark, although cold and dry conditions across western Greenland and the Canadian Arctic was a feature. These cold atmospheric conditions were reflected in waters off SW Greenland by temperatures below normal in the upper 400 m (Buch, 1998). Relative to this period we were only able to present data from 1988-1993, but the considerable fall in mean upper layer temperature over Flemish Cap was plausible (Figure 14). This temperature fall came together with a net poleward export of heat flux during the cold years to balance the heat deficit to the north. In contrast, the quinquennuim 1995-2000 was characterized by an increase in both SST and subsurface temperatures (e.g. Colbourne and Foote, 2000), which was accompanied by null to equatorward net heat flux across the 47ºN line (Figure 13b).

Both processes may be linked in a coupled slope water system (CSWS) responding in a similar manner to climate forcing over a broad range of time scales. Two characteristic CSWS modes have been identified over the Northwest Atlantic (Pickart et al., 1999). The maximum mode is characterized by deep and intense convection in the Labrador Sea; a relatively cool, fresh and thick layer of Labrador Sea Water (LSW) is formed; and the volume transport in the Deep Western Boundary Current increases while volume transport in the Labrador Current diminishes. Conversely, the minimum mode corresponds to a system state in which convective renewal of intermediate and deep waters in the Labrador Sea is weaker and shallower; LSW becomes warmer, saltier and thinner and volume transport in the DWBC diminishes while the volume transport in the shallow Labrador Current increases (Dickson et al., 1996; Dickson, 1997; Curry et al., 1998). Heywood et al. (1994) already reported that interannual changes in the LC offshore branch are common and pointed out that these could be associated with variations in the West Greenland Current.

Although the linkage is not straightforward it is tempting to associate maximum (minimum) modes of the CSWS with positive (negative) phases of the North Atlantic Oscillation (NAO) (Pickart et al., 1999). Hurrel (2000) discussed that the intensity of the winter convective activity in the North Atlantic is characterized by interannual as well as interdecadal variability that appears to be synchronized with the NAO.

The winter (December through March) index of the NAO based on the difference of normalized sea level pressure anomalies between Lisbon, Portugal and Stykkisholmur/Reykjavik, Iceland is presented in Figure 16. Data presented here are maintained by Jim Hurrell from the National Center for Atmospheric Research (Climate Analysis Section) and are public online through their website (http://www.cgd.ucar.edu/~jhurrell/nao.html). The atmospheric conditions over the North Atlantic area are highly dominated by the NAO, which has been in a persistent and exceptionally positive phase since the early 80's, which started to show symptoms of decline towards the mid 90's. In the NAO positive phase, higher than normal surface pressures south of 55ºN combine with a broad region of anomalously low pressure throughout the Arctic, which brings about, among other things, anomalously strong cold northerly flow across western Greenland and the Canadian Arctic, descending land and SST over the whole northwest Atlantic (Hurrell, 2000). Intense convective activity in the Labrador Sea until the early 90's has been documented (e.g. Hurrell, 2000). LSW has become progressively colder and fresher but associated surface LC transports had fallen to a minimum by the mid 90's as inferred from the heat flux estimates presented here and which belong to a clear maximum CSWS mode.

Conversely, the weakening of the NAO phase after the mid 90's may be associated with a shift towards a minimum CSWS mode that brings about the formation of a large volume of warmer and saltier LSW. This shift in CSWS is noted at Flemish Cap and brings about the relative increase of both northward/southward components of the heat flux and a tendency towards a net balanced or equatorward heat flux perpendicular to the 47ºN line across the bank. This, together with the concommitant relative increase of both northward/southward components of the geostrophic volume transport in this five-year period, also indicates good development of the anticyclonic circulation associated with Flemish Cap and we can conclude that the stronger the Labrador Current, the better the conditions for the development of the anticyclonic gyre anchored on the topography of the bank.

CONCLUSIONS AND SUMMARY

The recurrence of the casuistic of anticyclonic circulation around Flemish Cap in July permits us to infer that the topographic constraint in terms of the bank's location and dimension primarily determines dynamics over the bank. A coherent cold flow skirts through the northeastern flank and partly recirculates around the southern and southwestern flanks to isolate a central anticyclonic core. This summer circulation is held responsible for the intense temperature gradients between the Cap's center and its surroundings. We can conclude that the most significant source of variability of the water masses over Flemish Cap was linked to the variability of the advective flows, namely the offshore branch of the LC and oscillations of the NAC's north wall.

The sampling methodology made up of a high station density in a spiral track from the borders towards the center in some 8-10 days permitted the intensity and shape of the structures to be determined to a reasonable degree of accuracy. The resulting picture shows a mean asymmetric anticyclonic circulation following the bathymetry with enhanced southerly geostrophic velocities of ~7 cm s-1 with respect to the poleward flow of ~3 cm s-1. From the surface through the bottom the gyre bears warmer and less saline waters than its surroundings anchored on the topography of the bank. The most conspicuous anticyclonic gyre was sampled in 1998 when isolines closed around the Cap almost completely. The occurrence of warm and saline waters along the southern flank of the Cap has also been observed, which may well have their source in subtropical NAC waters. These showed a maximum of occurrence in 1988 with salinity values above 34.8 and temperatures ~6 ºC at 120 dbar.

These results confirm that the incoming flows over Flemish Cap are adequate enough in terms of intensity and stratification to be subjected to the principle of conservation of Ertel's Potential Vorticity and give rise to anticyclonic circulation on the bank. This seems to be a recurrent process on the observational basis but in an intermittent manner, as drifting buoy data (e.g. Loder et al., 1988) suggest that residence times are significantly lower than recirculation times. Kudlo et al. (1984) put forward that the frequency of storm passages may exert influence on the break up of the gyre. We have observed that NAC intrusions may be held responsible for weakening of the gyre. In this sense, the gyre bears the lighter waters versus the colder, denser eastern masses. Below 60 dbar the NAC-sourced waters are extremely warm versus the low salinity, colder ambient waters, whereby lighter densities are borne for the former. Subsurface (e.g. 120 dbar) densities are quite similar for both the NAC-type and the anticyclonic gyre waters. In the case that both water types are approached along the southern or southwestern flank, the density gradient between these may dwindle, even causing the anticyclonic gyre to die out.

The total northward (positive) and southward (negative) transports were O(0.3) Sv with stronger intensity of both components after 1995, the overall budget being well balanced in terms of average calculations. The series of geostrophic heat flux anomaly was also estimated to be well balanced and being in the order of 2.3 TW in each direction, although the long-term trend of the heat flux series seems to be slightly directed towards a net shift from positive (poleward) in the late 80's to slightly negative (equatorward) net heat flux in the second half of the 90´s.

The first period studied (1989-1994) seemed to belong to the maximum CSWS mode, characterized by deep and intense convection in the Labrador Sea and reduced volume transport in a colder, fresher Labrador Current, which materialized in a generalized temperature fall in the Flemish Cap that came together with a net poleward export of heat flux during the cold years to balance the heat deficit to the north.. The second period studied (1995-2000) was characterized by an increase in both SST and subsurface temperatures, which was accompanied by null to equatorward net heat flux across the 47ºN line and therefore seems to be affected by a shift towards a minimum CSWS mode, with weaker convective renewal of intermediate and deep waters in the Labrador Sea and an enhanced warmer, more saline Labrador Current. July SST anomalies and heat flux across the 47º N line were seen to be a consequence of the variability of the Labrador Current as being the main incoming flux over the bank rather than a consequence of seasonal processes within the retained waters. Intensity of the southward off-shore branch of the Labrador Current over the bank is concluded to be the primary mechanism determining variability in heat transport across Flemish Cap.

It must be questioned whether the source of LC variability and the shift in the CSWS is linked to the NAO, but this shift in CSWS is noted at Flemish Cap and brings about the relative increase of both northward/southward components of the heat flux and a tendency towards a net balanced or equatorward heat flux perpendicular to the 47ºN line across the bank. This, together with the concommitant relative increase of both northward/southward components of the geostrophic volume transport in this five-year period, would also indicate good development of the anticyclonic circulation associated with Flemish Cap and we can conclude that the stronger the Labrador Current the better the conditions for the development of the anticyclonic gyre anchored on the topography of the bank.

REFERENCES

Anderson, J.T. 1984. Early life history of redfish (Sebastes spp.) on Flemish Cap. Can. J. Fish. Aquat. Sci., 41(7): 1106-1116

Akenhead, S.A. 1986. Water retention over Flemish Cap. The role of freshwater outflow in coastal marine ecosystems. NATO ASI Series, G7, 283-293. S. Skreslet ed.

Baranov, E.I. and V.G. Ginkul, 1984. Dynamics of waters of the Newfoundland energy active zone. Meteorogiya di Gidrologiya, 12, 78-84

Barnes. 1964. A technique to maximizing details in numerical weather map analyses. J. Appl. Meteor., 3, 396-409

Borovkov, V.A., and B.P. Kudlo. 1980. Results of USSR oceanographic observations on Flemish Cap, 197-78. ICNAF Sel. Pap., 6: 47:52

Buch, E. 1998. Oceanographic investigations in West Greenland Waters, 1981-1997. NAFO SCR Doc. 98/24. Serial No. N3008

Cerviño, S. and R. Prego, 1997. Hydrographic Conditions on the Flemish Cap in July 1996. NAFO SCR Doc., 17, Serial No. N2847

Cerviño, S., R. Prego and M. Gómez-Gesteira. New local-scale hydrographic observation on Flemish Cap in July 1996. NAFO Sci. Coun. Studies, 32: 17-24

Colbourne, E. 1993. Oceanographic conditions on the Flemish Cap during the summer of 1993, with comparisons to the Long-Term average. NAFO SCR Doc., 93, Serial No. N2300

Colbourne, E. 1995. Oceanographic conditions on the Flemish Cap during the summer 1995, with comparisons to the 1961-1990 average. NAFO SCR Doc., 102, Serial No. N2625

Colbourne, E. 1996. Oceanographic conditions on the Flemish Cap during the summer 1996, with comparisons to the previous year and the 1961-1990 average. NAFO SCR Doc., 96, Serial No. N2770

Colbourne, E. 1997. Oceanographic conditions on the Flemish Cap during the summer of 1997, with comparisons to the previous year and the 1961-1990 average. NAFO SCR Doc., 84, Serial No. N2930

Colbourne, E. and K.D. Foote. 2000. Variability of the Stratification and Circulation on the Flemish Cap during the Decades of 1950s and 1990s. J. Northw. Atl. Fish Sci. 26: 103-122

Curry, R.G.., M.S. McCartney and T.M. Joyce. 1998. Oceanic transport of subpolar climate signals to mid-depth subtropical waters. Nature. 388: 575-577

Davis, P.A. 1972. Experiments on Taylor columns in rotating stratified fluid. J. Fluid Mech., 54, 691-717

Dickson, R. J. Lazier, J. Meinke, P. Rhines and J. Swift. 1996. Long-term coordinate changes in the convective activity of the North Atlantic. Progr. Oceanogr. 38: 241-295

Dickson, R. 1997. From the Labrador Sea to global change. Nature. 386: 649-650

Doswell, C.A.1977. Obtaining meteorologically significant surface divergence fields through the filtering property of objective analysis. Mon. Wea. Rev., 105, 885-892

Fofonof, N.P. and R.M. Hendry, 1985. Current variability near the southeast Newfoundland Ridge. Journal of Physical Oceanography, 15, 963-984

Freeman, H. The heat flux across Line-P 1996-1999. Atm-Ocean. 40(1), 81-89

Gil, J., R.F. Sánchez, M. Blanco and I. Reguera. 1998. Hydrodynamics associated to the confluence of Labrador and North Atlantic currents over Flemish Cap in July 1997. NAFO SCR Doc. 98/32 Serial No. N3019

Gill, A.E. 1982. Atomsphere-ocean dynamics. International Geophysics Series, Vol 30. Academic Press, Inc. 662pp

Greenberg, D.A. and B.D. Petrie, 1988. The mean barotropic circulation on the Newfoundland Shelf and Slope. Journal of Geophysical Research, 93, No. C12, 15541-15550

ayes, R.M., D.G. Mountain and T.C. Wolford. MS 1977. Physical oceanography and the abiotic influence on cod recruitment in the Flemish Cap region. ICNAF Res. Doc. 54, serial No. 5107, 33pp.

Heywood, K.J., E.L. McDonald and M.A. White. 1994. Eddy kinetic energy of the North Atlantic subpolar gyre from satellite altimetry. Journal of Geophysical Research, 99, C11, 22525-22539

Huppert, H.E. 1975. Some remarks on the initiation of inertial Taylor Columns. J. Fluid Mech., 67: 397-412

Huppert, H.E. and K. Bryan. 1976. Topographically generated eddies. Deep-sea Res., 23: 655-679

Hurrell, J.W. 2000. Climate: North Atlantic and Arctic Oscillation (NAO/AO). In: Encyclopedia of Atmospheric Sciences. Eds. J: Holton, J. Pyle and J. Curry. Academic Press.

Karasayev, B.E. 1962. Hydrologic investigations carried out aboard the vessels of the Kaliningrad Council for People's Economy in 1961 in ICNAF Subareas 2 and 3. ICNAF Res. Doc., 71, Serial No. 1014, 10 pp.

Krauss, W., E. Fahrbach, A. Aitsam, J. Elken and P. Koske. 1987. The North Atlantic Current and its associated eddy field southeast of Flemish Cap. Deep-Sea Research 34 No 7:1163-1185

Kudlo, B.P. and V.V. Burmakin. 1972. Water circulation in the south Labrador and Newfoundland areas in 1970-71. ICNAF Redbook, Vol III: 27-33

Kudlo, B.P., B.A. Borokov, 1975. Circulation of waters in the ICNAF area in 1974-75. ICNAF Res. Doc. 79, Serial No. 3874, 14 pp

Kudlo, B.P., B.A. Borokov and N.G. Sapronetskaya. 1984. Waters circulation pattern on Flemish Cap from observations in 1977-82. NAFO Sci. Coun. Studies, 7, 27-38

Lee, A.J. 1968 NORTHEASTATLANT Surveys: Physical Oceanography. ICNAF Spec. Publ. Part I. 31-54

Lozier, M.S., W.B. Owens and R.G. Curry. 1995. The climatology of the North Atlantic. Progr, Oceanogr. 36: 1-44

Maddox, R.A. 1980. An objective technique for separating macroscale and mesoscale features in meteorological data. Mon. Wea. Rev., 108, 1108-1121

McLellan, H. J. 1957. On the distinctness and origin of the slope water off the Scotian Shelf and its easterly flow south of the Grand Banks, J. Fish. Res. B., Can., 14, No. 2

NODC. (2000). World Ocean Database 1998. Documentation and Quality Control. Version 2.0. Ocean Climatic Lab. NODC Internal Report 14. Silver Spring MD. January 2000. (http://www.nodc.noaa.gov/oc5/pr_wodv2.html)Petrie and Anderson. 1983.

Pickart, R.S., T.K. McKee, D.J. Torres and S.A. Harrington. 1999. Mean structure and interannual variability of the slope water system south of Newfoundland. J. Phys. Oceanogr. 29: 2541-2558

Planque, B. and A.H. Taylor. 1998. Long-term changes in zooplankton and the climate of the North Atlantic. ICES J. Mar. Sci. 55: 644-654

Robe, R.Q. MS 1974. Circulation east of the Grand Banks, 1973. + Addendum. ICNAF Res. Doc. 107 Serial No. N3345, 1 p + 8 p addendum

Ross, C.K. 1981. The drift of satellite-tracked buoys on Flemish Cap, 1979-80. NAFO Sci. Coun. Studies, 1: 47-50

Smith, E.H., F.M. Soule, and O. Mosby, 1937: The Marion and General Green expeditions to Davis Strait and the Labrador Sea, Bull. U.S. Coast Guard, 19, 259 pp.

Stein, M. 1996. Flemish Cap – A review on Research Activities With Focus on Oceanographic Conditions. NAFO Sci. Coun. Studies, 25: 1-24

Taylor, G.I. 1923. Experiments on the motion of solid bodies in rotating fluids. Proc. R. Soc. London, Ser. A, 104: 213-218

Templeman, W. 1974. Temperature and salinities in the eastern Newfoundland area in 1973. ICNAF Res. Doc., 71, Serial No. 5104, 13 pp.

Trites et al. 1982
Unesco, 1991. Processing of Oceanographic Station Data, 1991. JPOTS editorial panel.
Vázquez, A. 2000. Results of the bottom trawl survey on Flemish Cap of July 1999. NAFO SCR Doc. 00/9. Serial No. 4228. 50 pp
Worthington, L.V. 1976. On the North Atlantic Circulation. The Johns Hopkins Oceanographic Studies, 6. 110pp

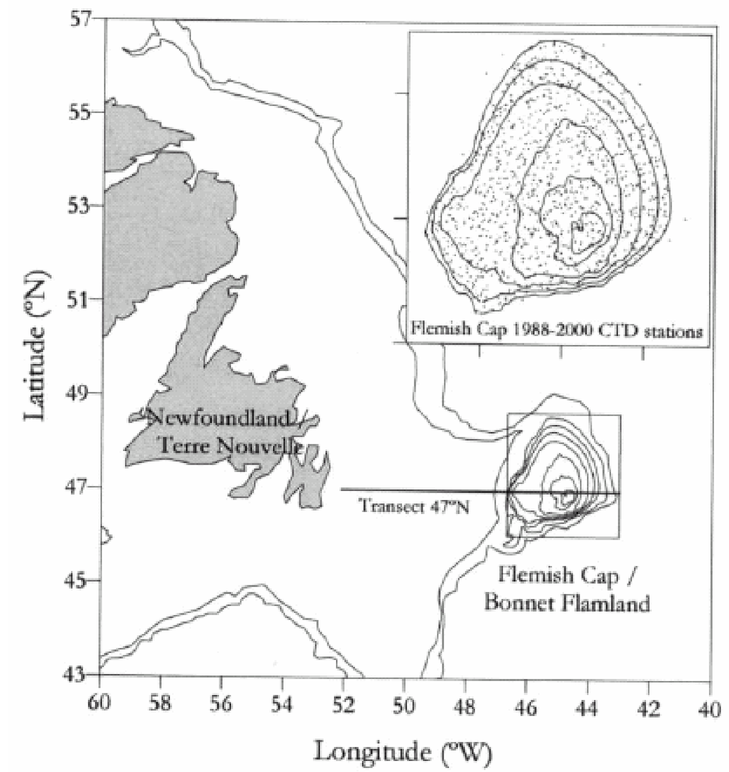

Fig. 1. Location map of the Flemish Cap area. Bathymetric lines are 140, 180, 250, 360, 550, 720, 1000 and 2000 m. Detail of the Flemish Cap is enlarged in the inset.

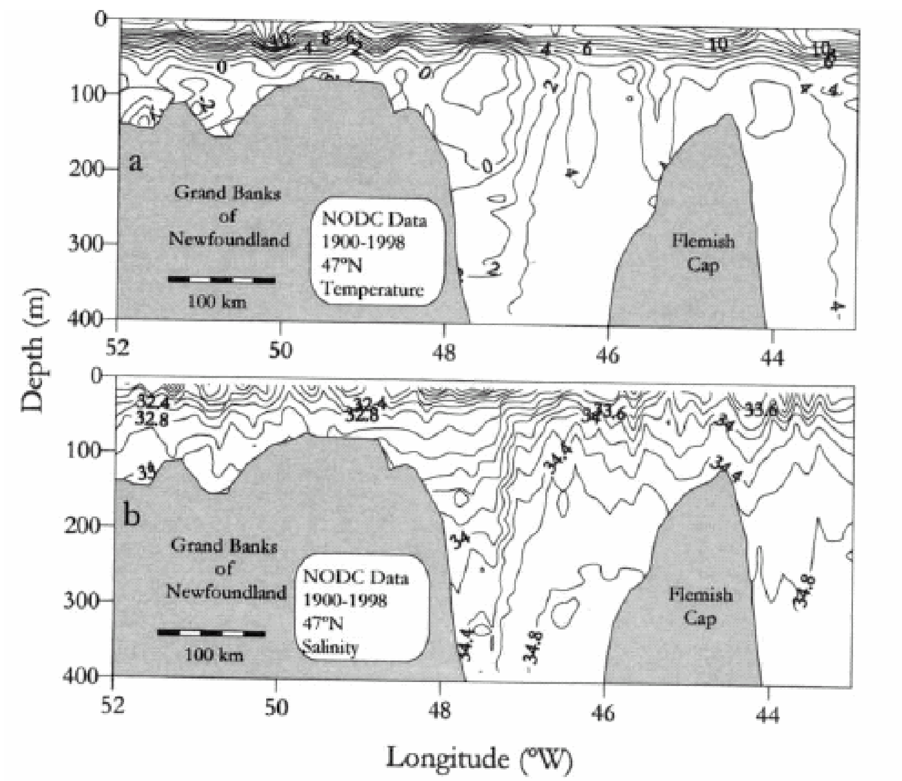

Fig. 2. Climatological transects across the 47ºN line. (a) Temperature (ºC). (b) Salinity. Source: NODC 1900-1998

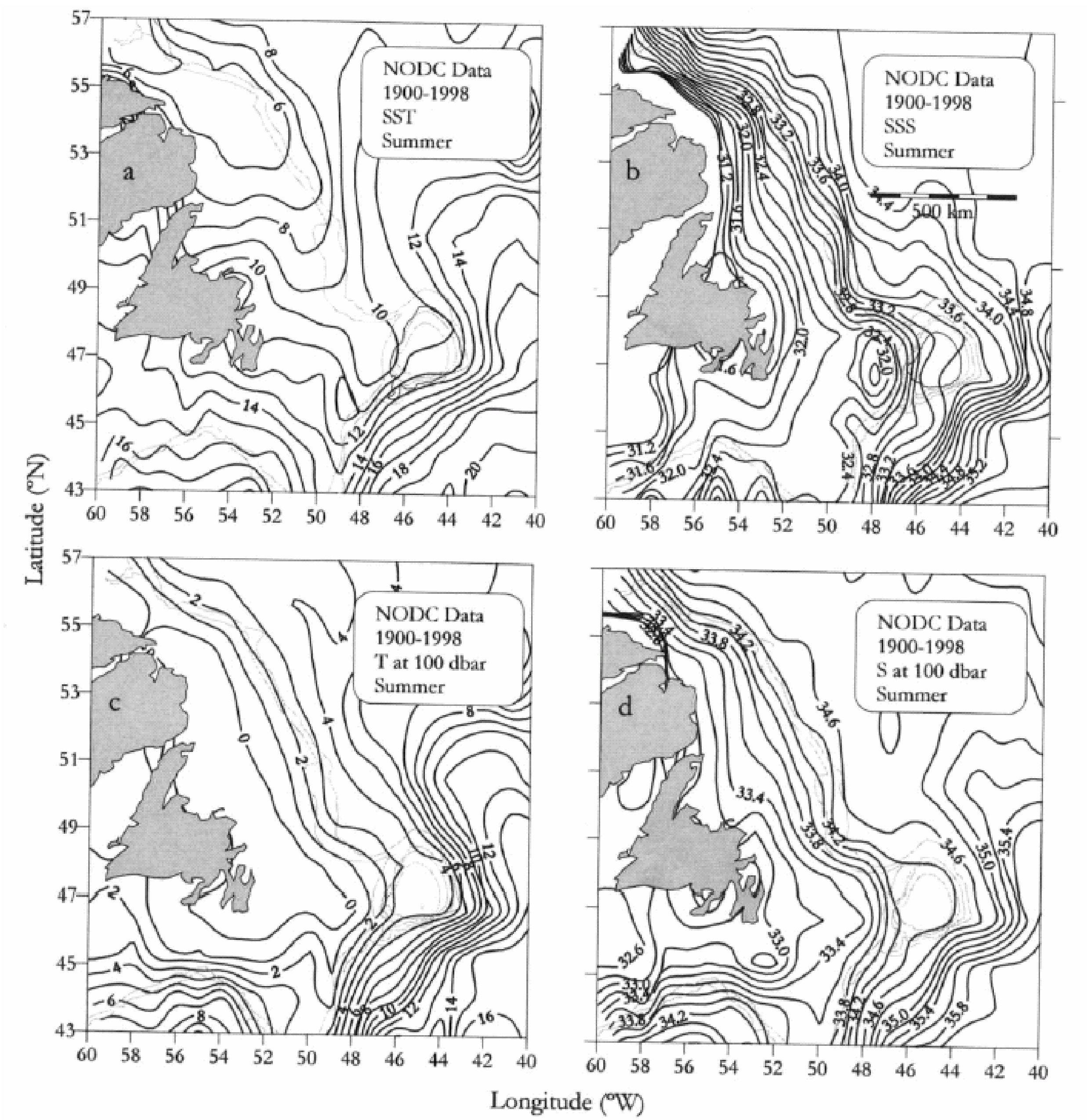

Fig. 3. Climatological plots of temperature (ºC) and salinity of the Newfoundland area. (a) and (b) at sea surface; (c) and (d) at 100 dbar.

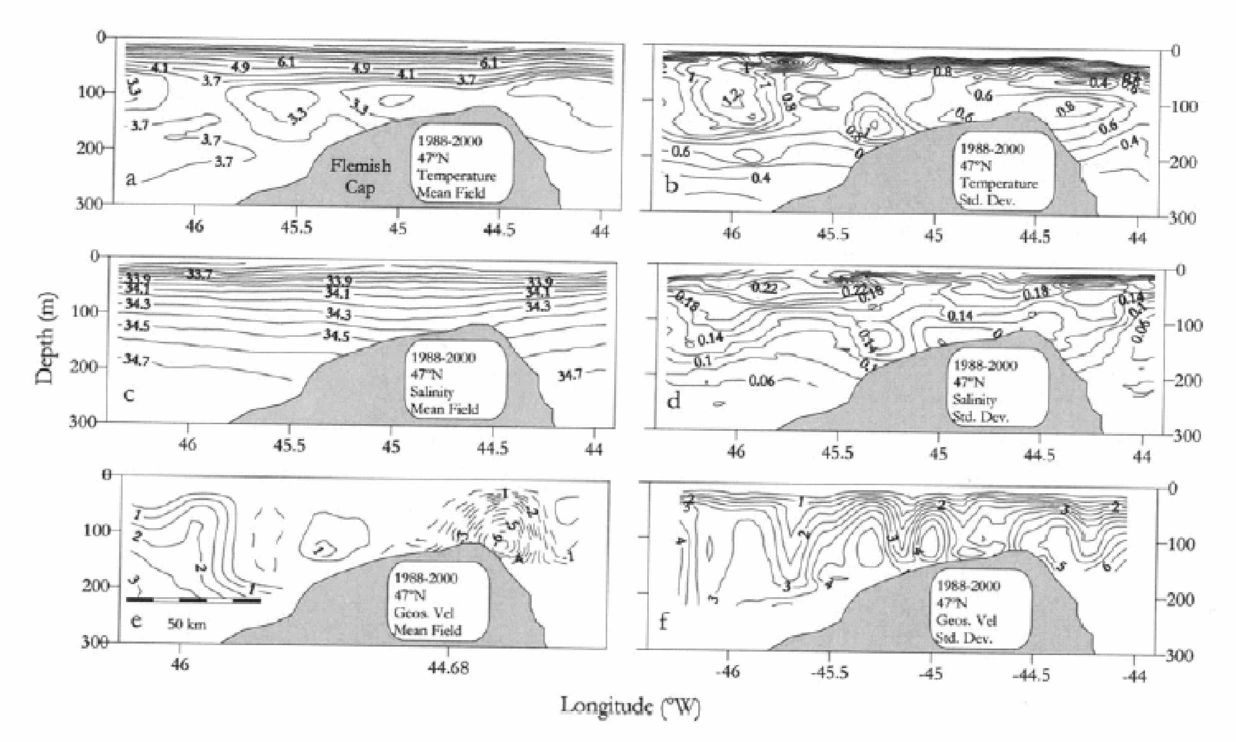

Fig. 4. 1988-2000 mean and standard deviation cross-sections along the 47ºN line from bottom trawl surveys. (a) and (b) temperature (ºC); (c) and (d) salinity; (e) and (f) geostrophic velocity (m s-1). Broken lines in Figure 4(e)

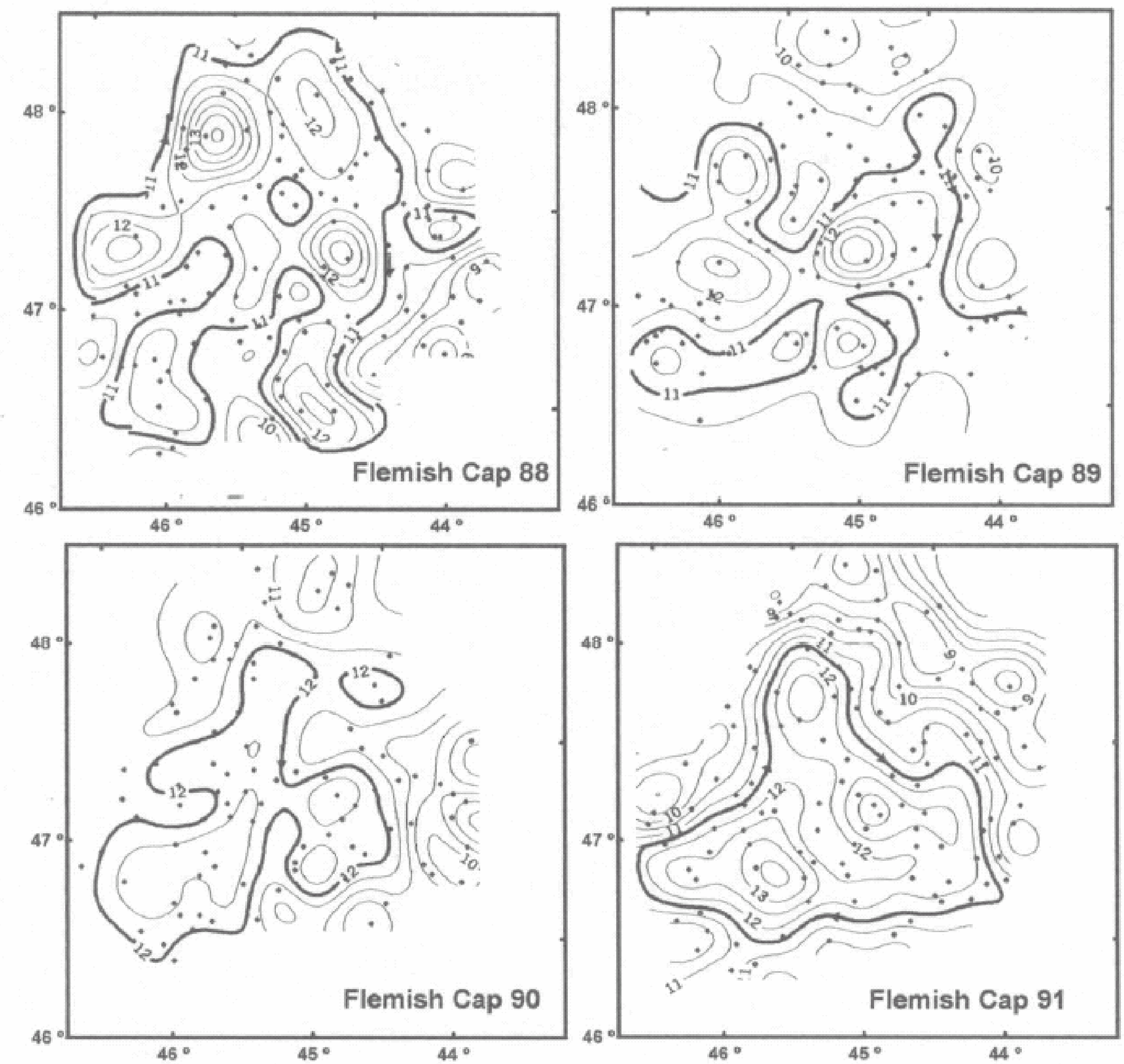

Fig. 5a. 1988-2000 series of dynamic topographies (dyn. cm). (i) 1988-1991

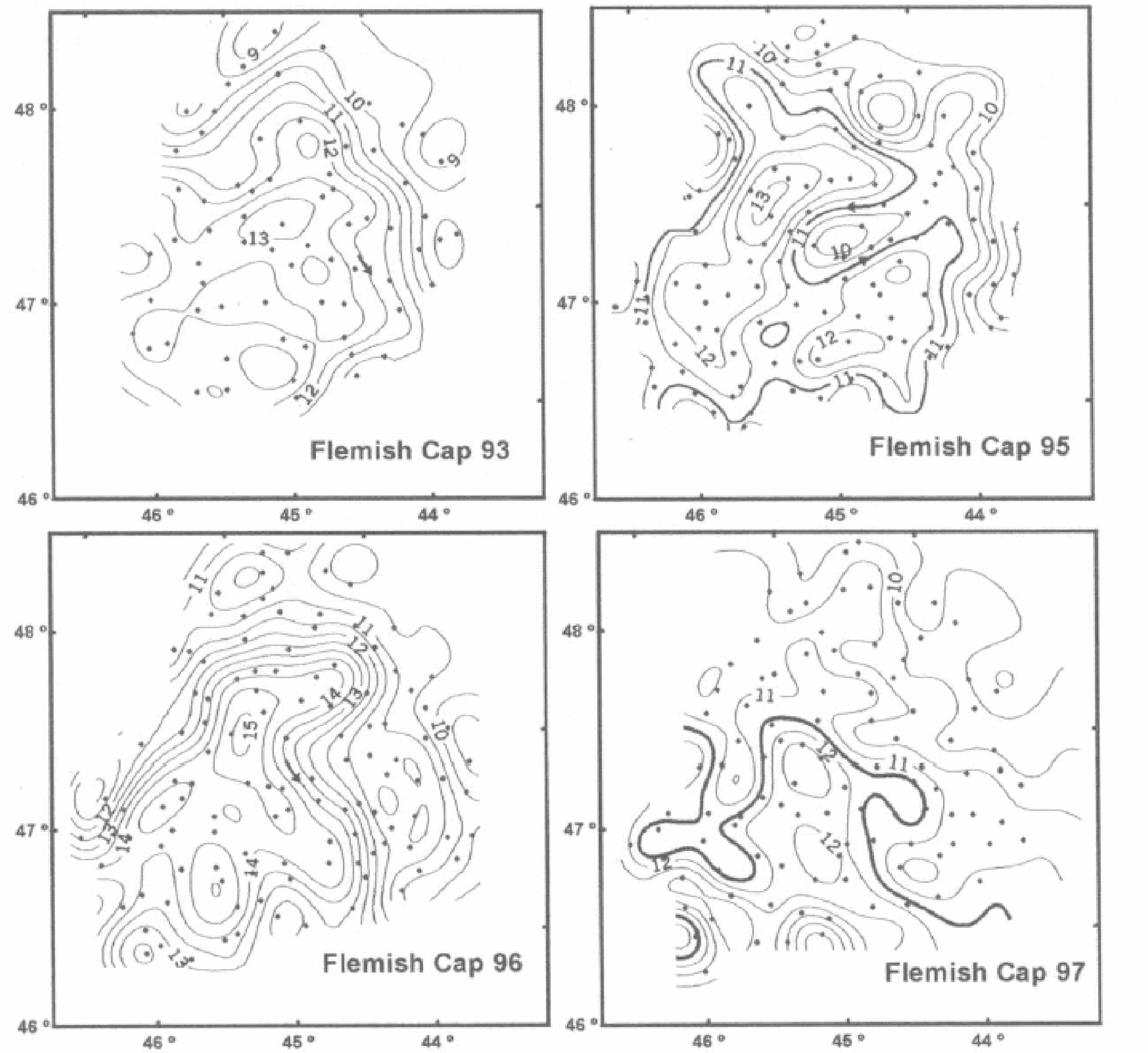

Fig. 5b. 1988-2000 series of dynamic topographies (dyn. cm). (ii) continued: 1993-1997

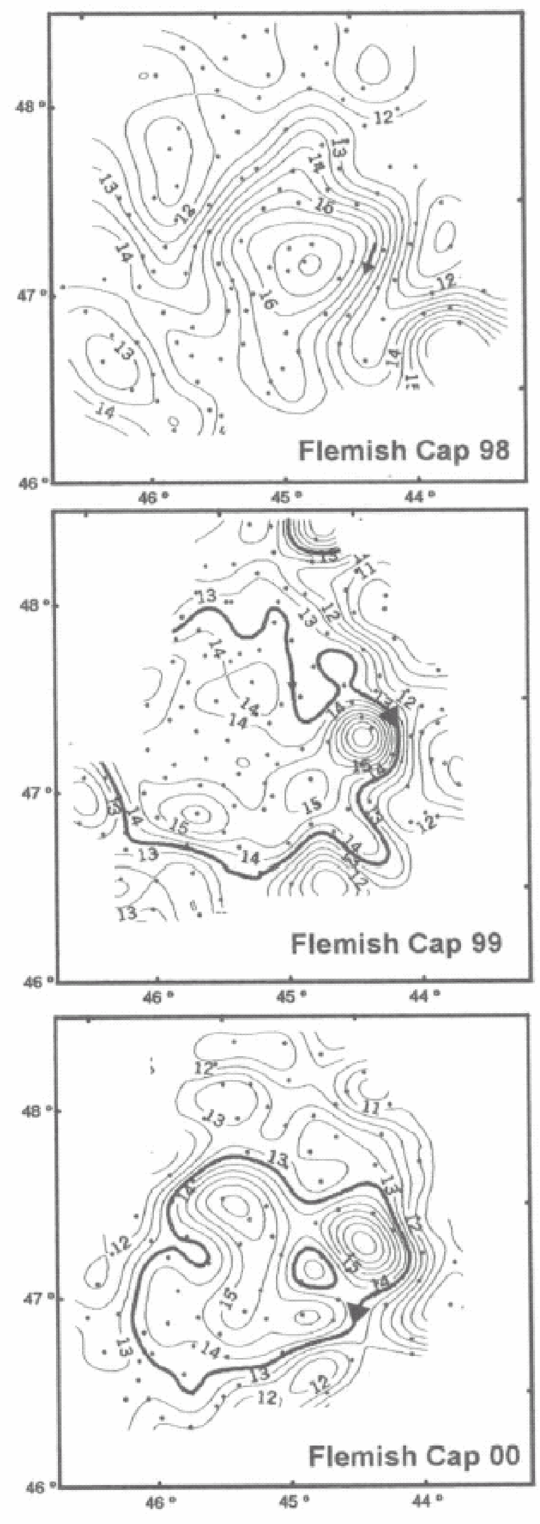

Fig. 5c. 1988-2000 series of dynamic topographies (dyn. cm). (iii) continued: 1998-2000

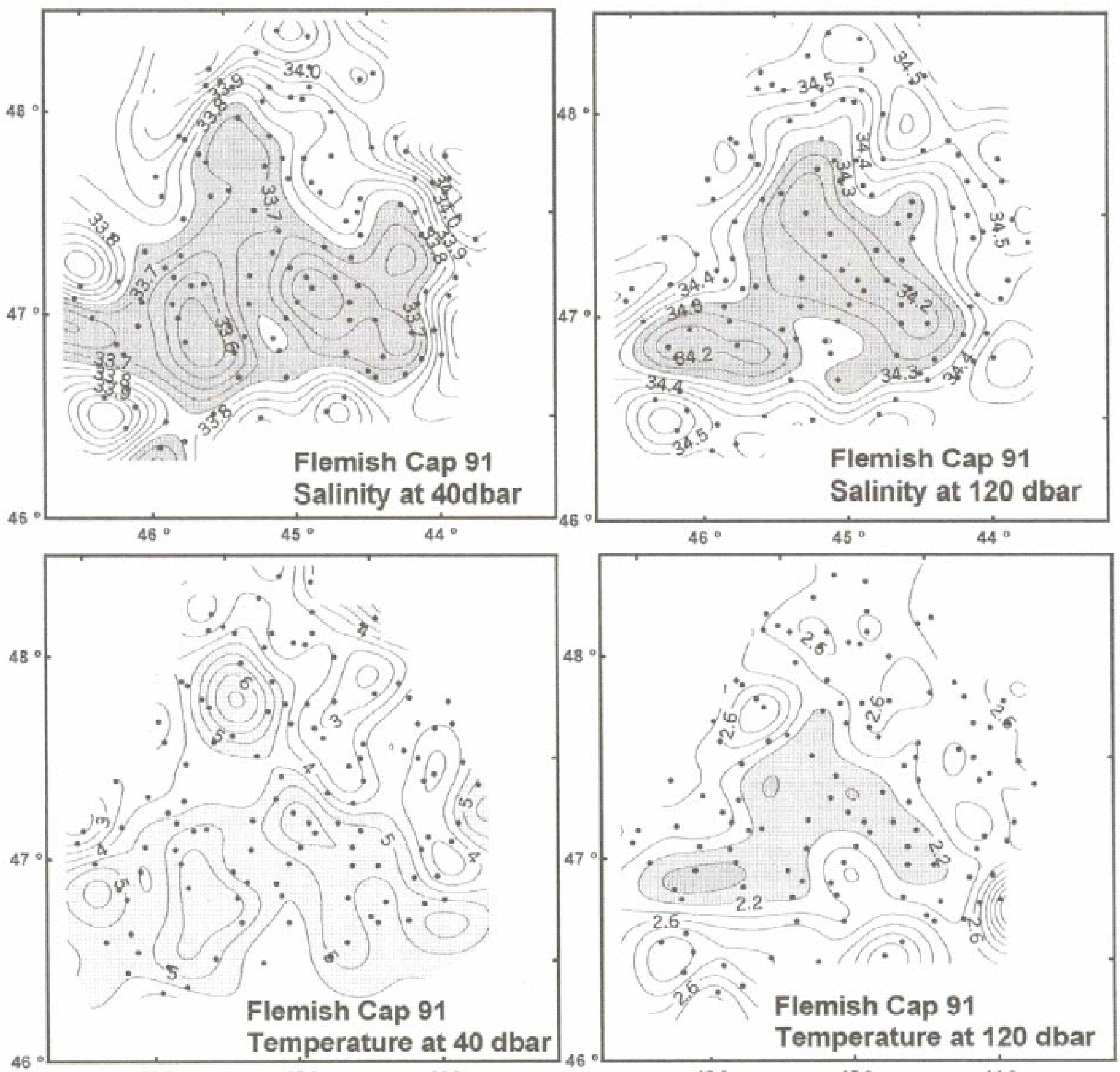

Fig. 6. 1991. Salinity at 40 and 120 dbar and temperature (ºC) at 40 and 120 dbar.

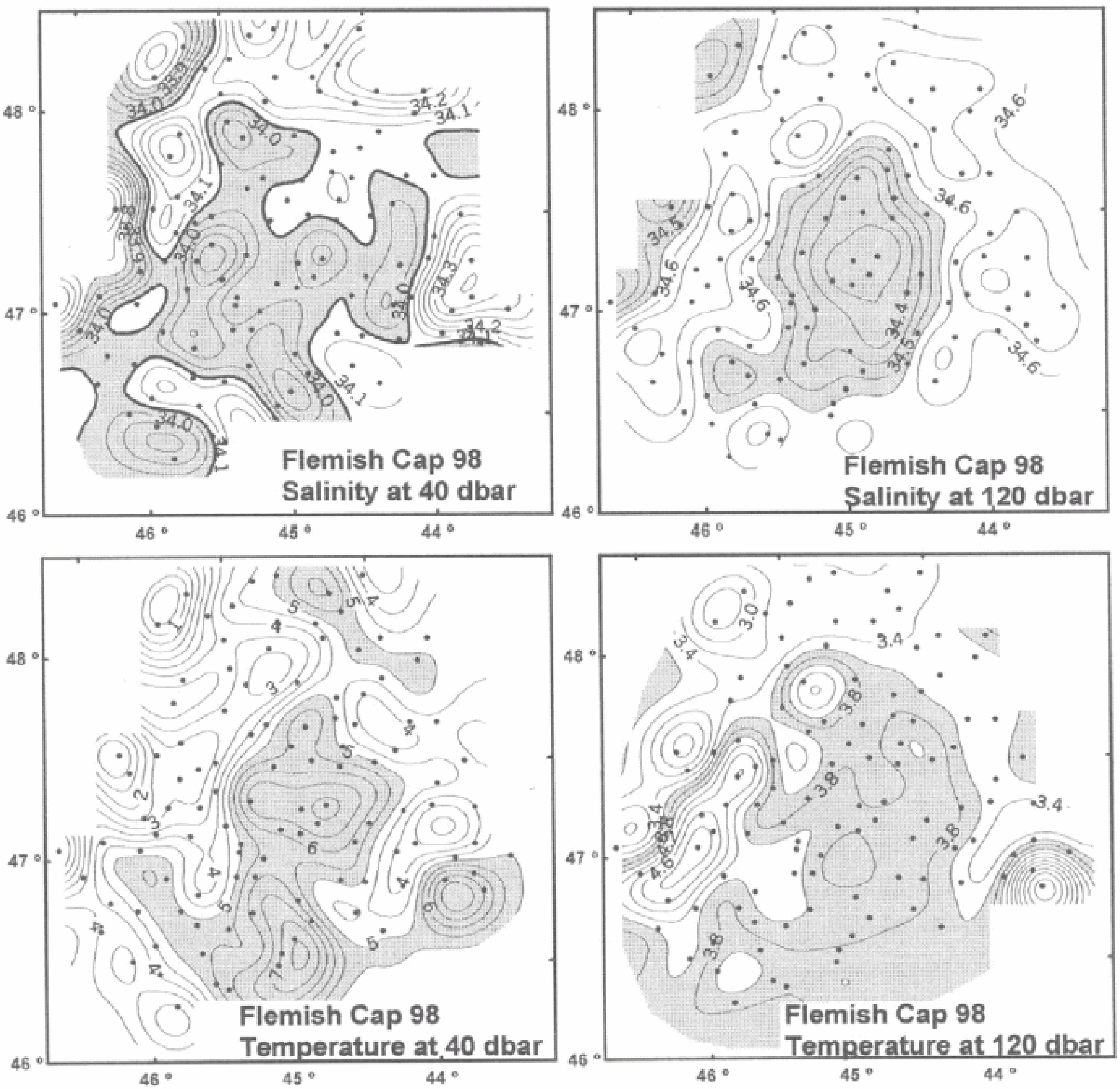

Fig. 7. 1998. Salinity at 40 and 120 dbar and temperature (ºC) at 40 and 120 dbar.

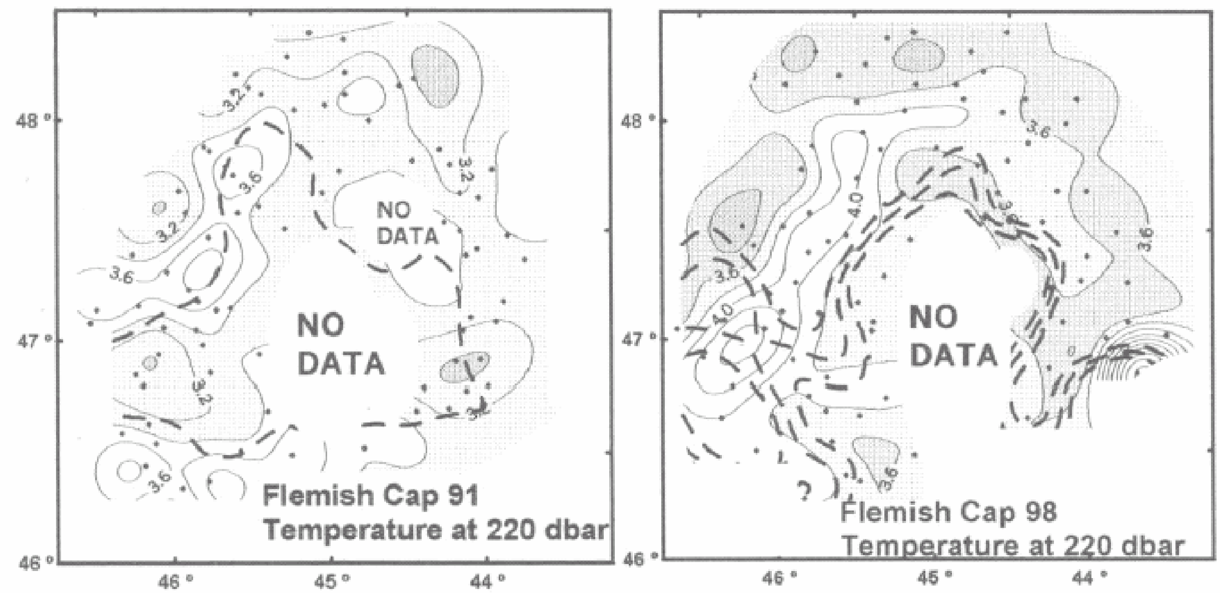

Fig. 8. Temperature (ºC) at 220 dbar in (a) July 1991 and (b) July 1991.

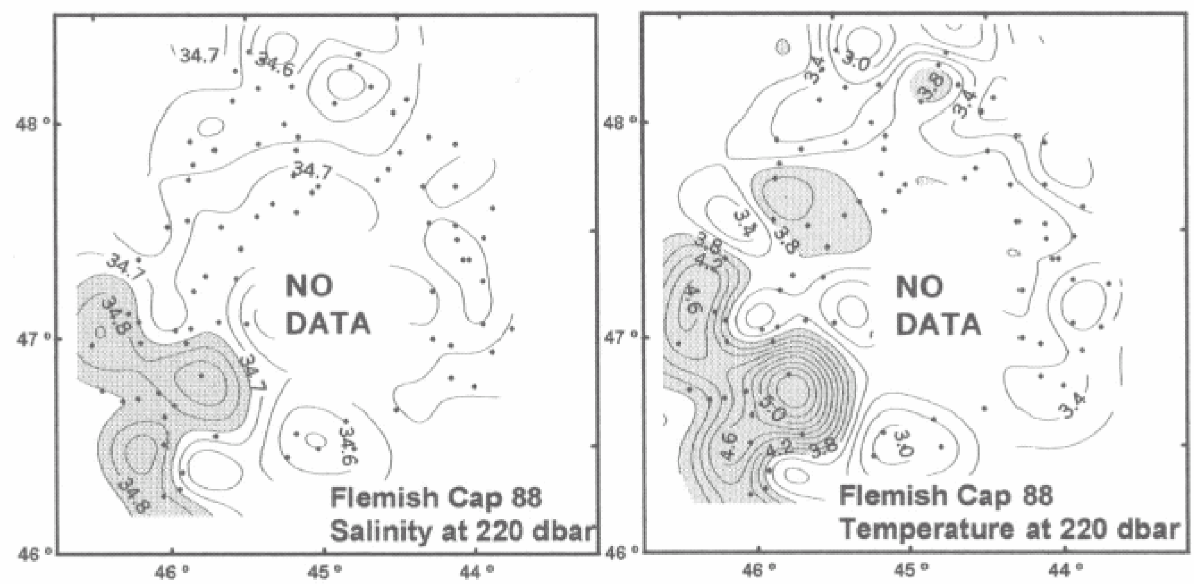

Fig. 9. Salinity and temperature at 220 dbar in July 1988.

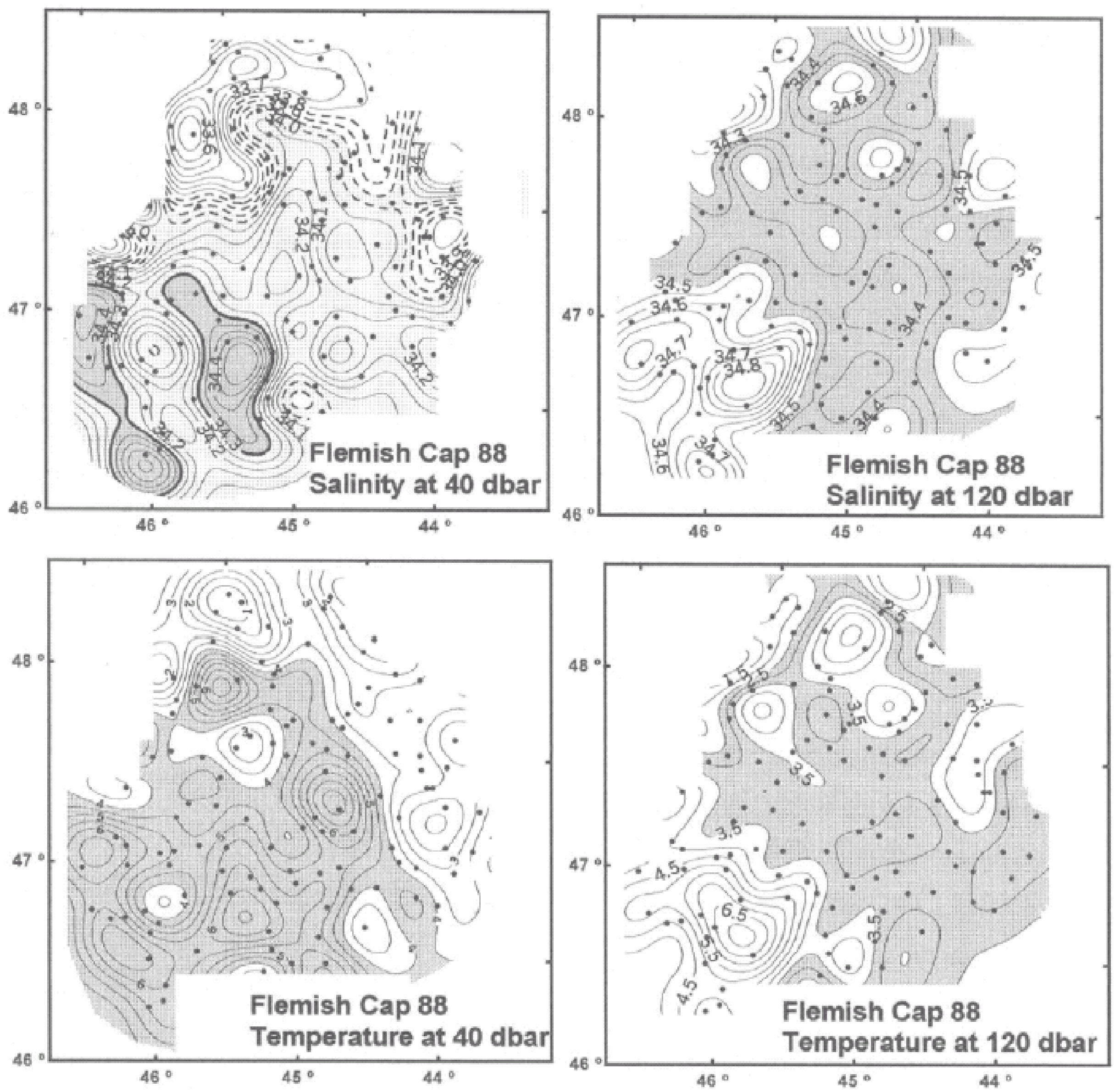

Fig. 10. 1988. Salinity at 40 and 120 dbar and temperature (ºC) at 40 and 120 dbar.

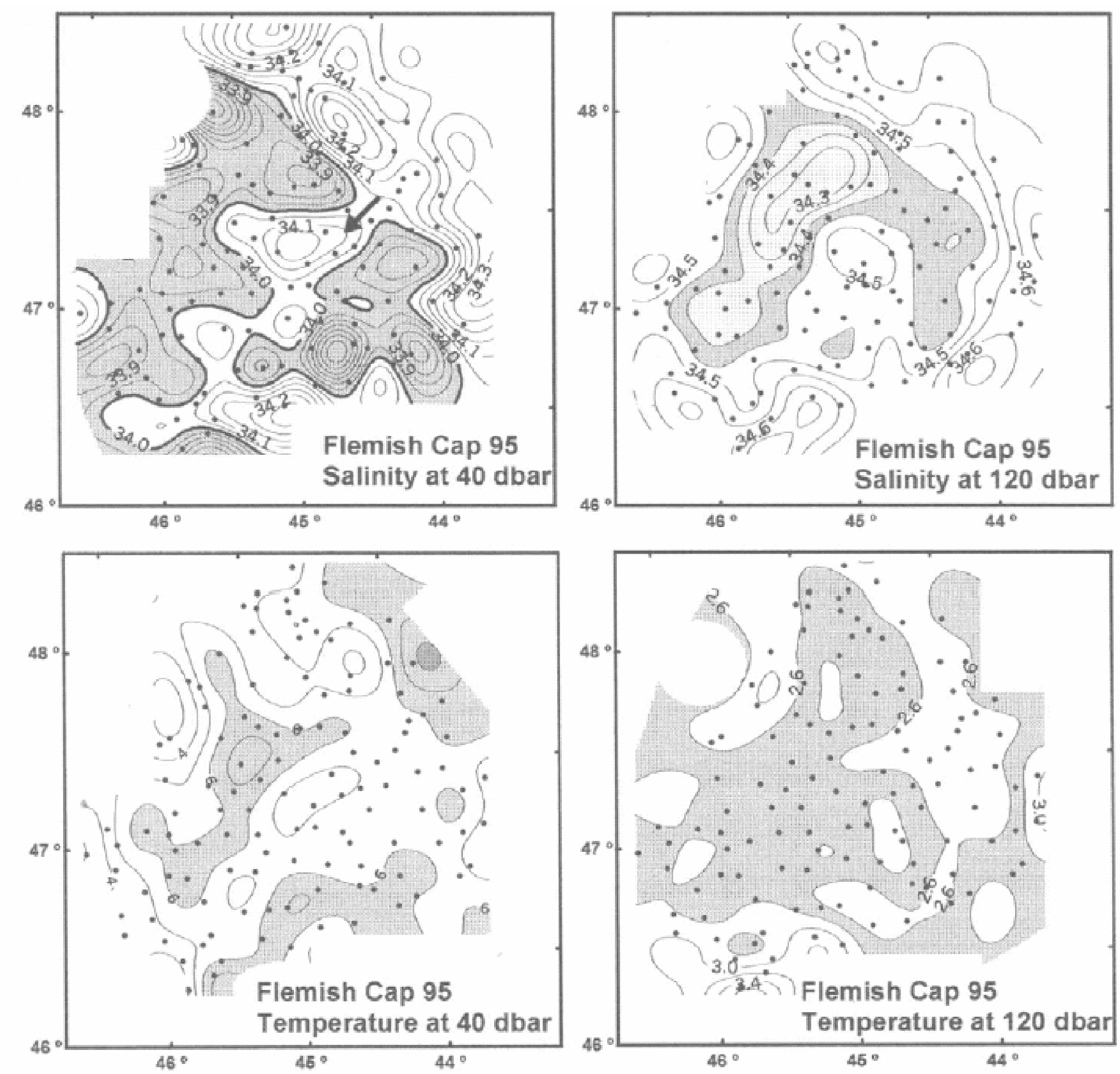

Fig. 11. 1995. Salinity at 40 and 120 dbar and temperature (ºC) at 40 and 120 dbar.

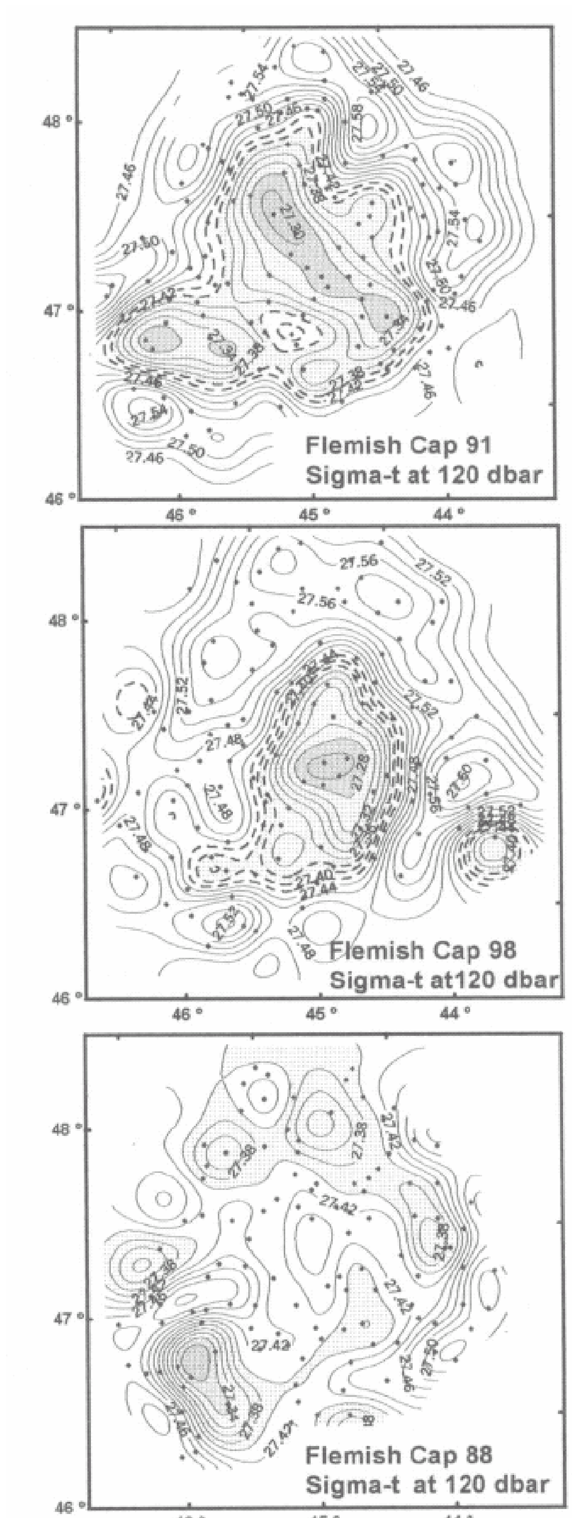

Fig. 12. Sigma-t at 120 dbar in (a) July 1991; (b) July 1998; (c) July 1988

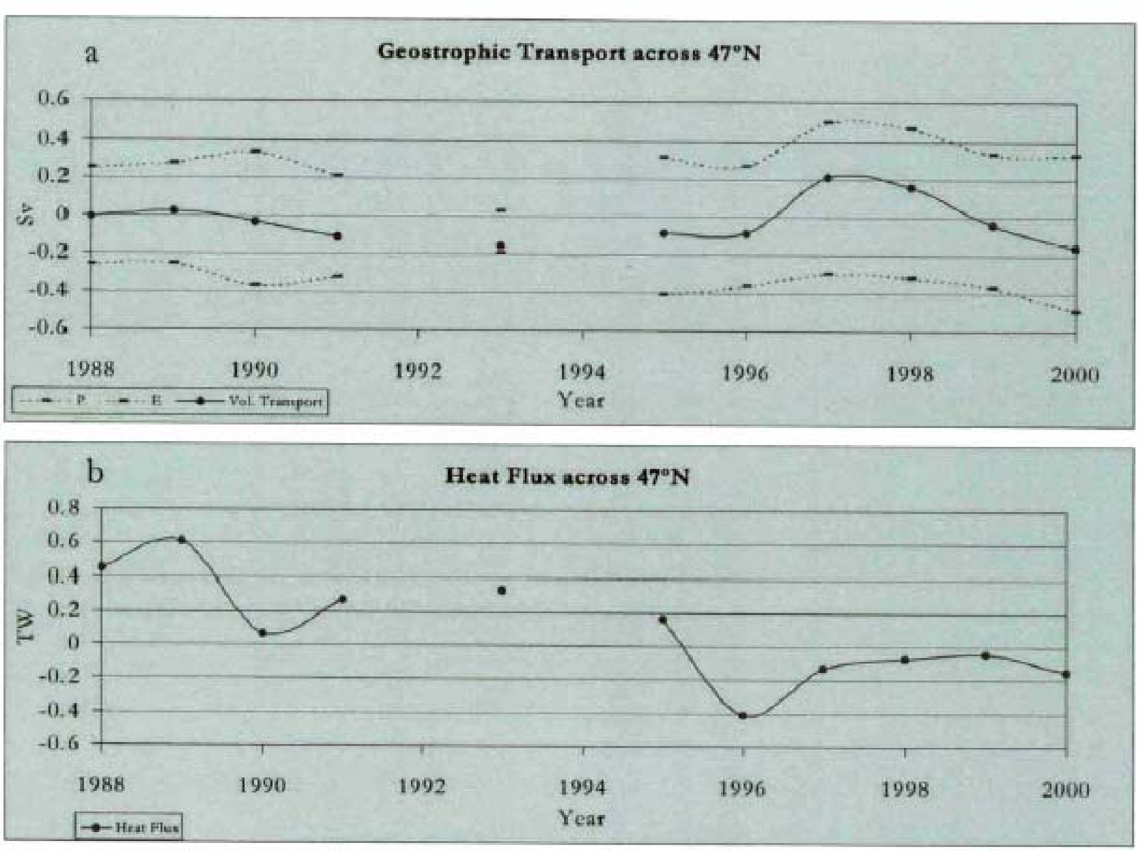

Fig. 13. (a) 1988-2000 geostrophic volume transport (Sv.) (1 Sv = 106 m3 s-1) across 47ºN over Flemish Cap. P states for poleward and E for equatorward components of the overall flow. (b) 1988-2000 net heat flux (Terra Watts) (1 TW = 1012 W) across 47ºN over Flemish Cap.

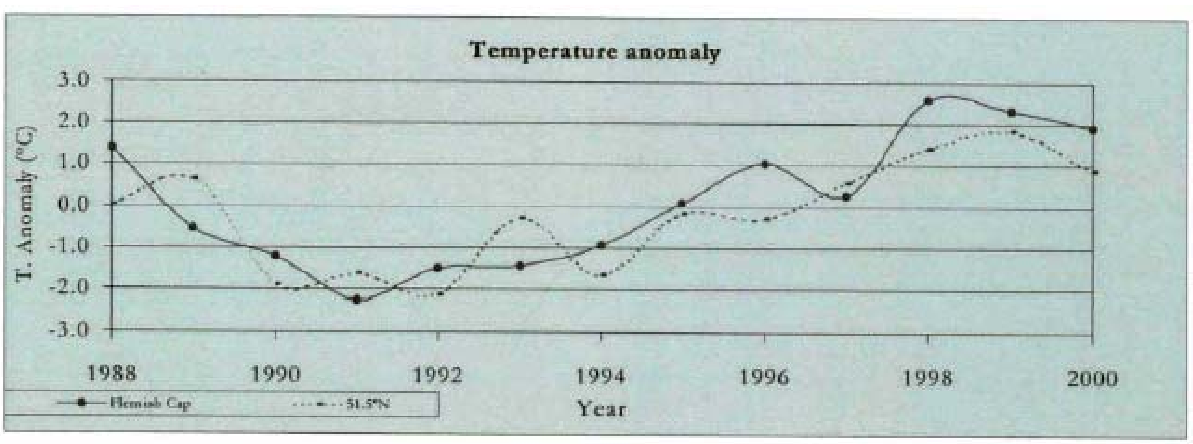

Fig. 14. Optimally interpolated SST (OISST) anomalies (ºC) from July 1988 to July 2000 at various locations in the western North Atlantic. Bold line: computed mean over Flemish Cap. Broken line: at 51.5ºN.

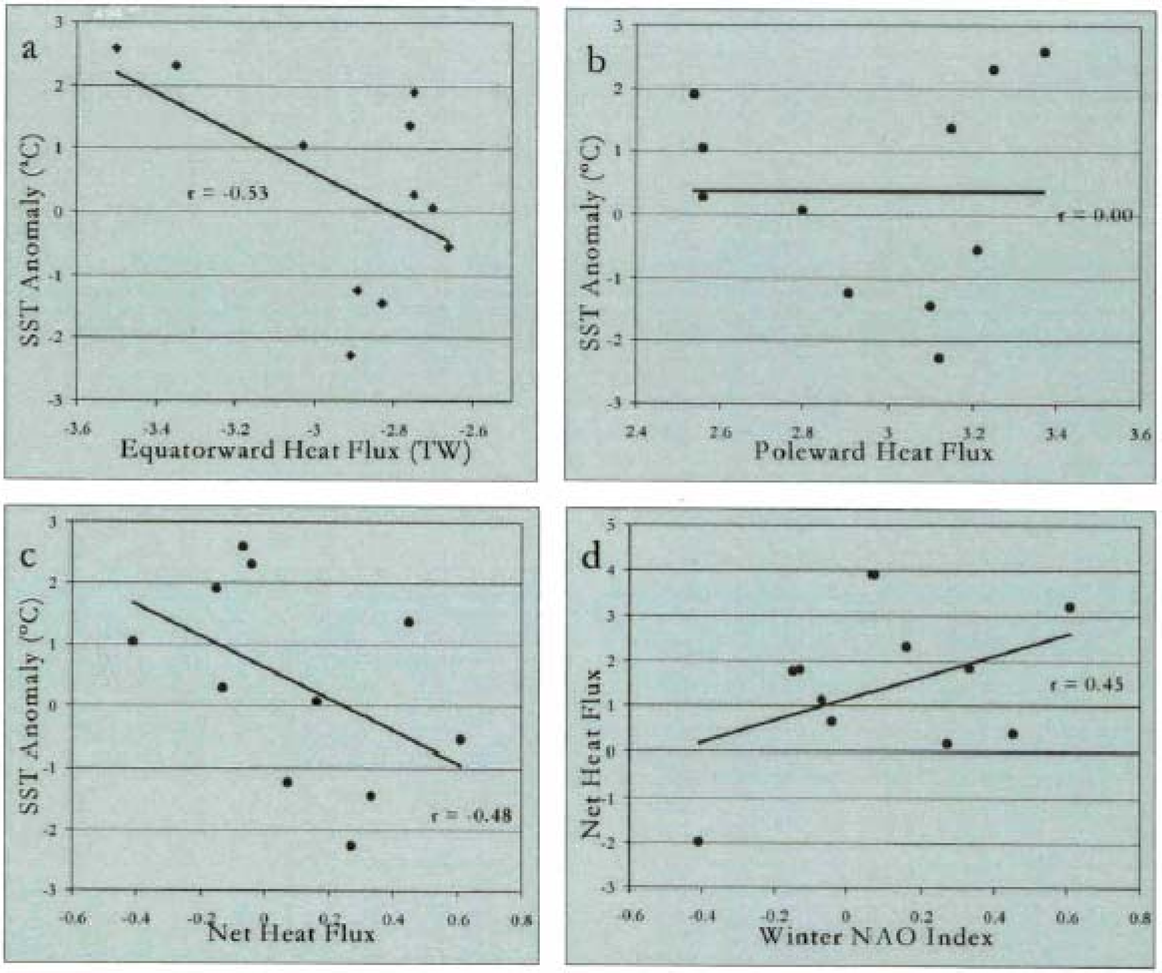

Fig. 15. Scatter plots of : (a) Flemish Cap OISST anomalies (ºC) versus Equatorward heat flux (TW) across 47ºN over Flemish Cap; (b) Flemish Cap OISST anomalies (ºC) versus Poleward heat flux (TW) across 47ºN over Flemish Cap; (c) Flemish Cap OISST anomalies (ºC) versus net heat flux (TW) across 47ºN over Flemish Cap; (d) net heat flux (TW) across 47ºN over Flemish Cap versus winter NAO index.

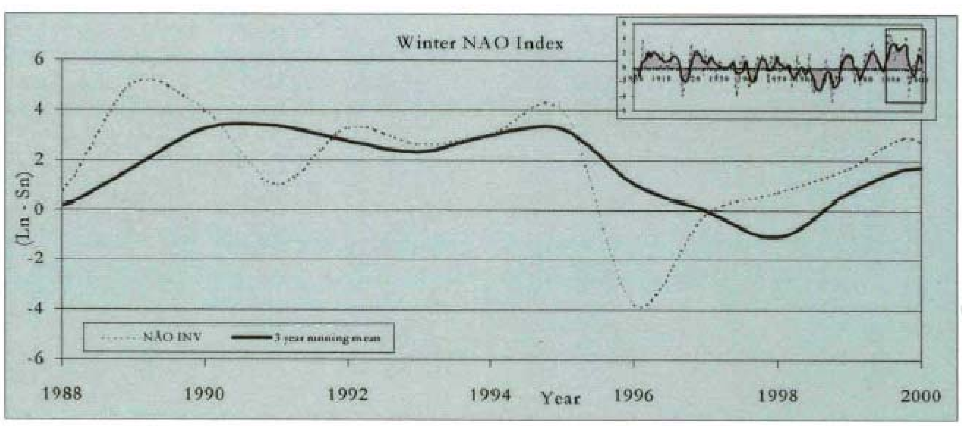

Fig. 16. 1988-2000 winter North Atlantic Oscillation (NAO) index (difference of normalised sea level pressure anomalies between Lisbon, Portugal and Stykkisholmur/Reykjavik, Iceland). The inset presents the 1900-2000 series. Bold line: 3-year running mean.