Hydrological regime and water volume transport in the Faeroe-Shetland Channel in summer of 1988 and 1989

Faeroe-Shetland Channel
Water masses
Dynamic calculations
Year-to-year variability

Passage Faeroe-Shetland
Masses d'eau
Calcul dynamiques
Variabilité annuelle

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ABSTRACT

This paper deals with the hydrological regime and the water volume transport in the Faeroe-Shetland Channel in early summer of 1988 and 1989. The oceanographic conditions in the channel and in its vicinity are reviewed briefly. Then, some thermohaline and dynamical analyses are presented. The results are based on the vertical CTD soundings carried out during the cruises of R/V *Oceania* in the region of the channel in July 1988 and in July 1989. The upper part of the water column in the channel, down to the bottom of the main pycnocline, was warmer and more saline in 1988. However, the net transport of warm waters with $\theta \geq 3°C$ was oriented to the north, and was rather larger in 1989 (1.8 ± 2.0 Sv) than in 1988 (1.0 ± 1.0 Sv). The cold water ($\theta < 3°C$) net transport (0.65 ± 0.35 Sv) was directed into the Norwegian Sea in 1988, in contrast with 1989, when it (- 0.65 ± 0.56 Sv) was directed into the Atlantic Ocean.

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RÉSUMÉ

Régime hydrologique et transport d'eau dans le Passage Faeroe-Shetland en été 1988 et 1989

L'article porte sur le régime hydrologique et le transport d'eau dans le Passage Faeroe-Shetland au début de l'été, en 1988 et 1989. Les conditions océanographiques dans le passage et dans son voisinage sont brièvement décrites, et certaines analyses thermohalines et dynamiques sont présentées. Les résultats ont été obtenus à partir des stations hydrologiques effectuées par le navire océanographique *Oceania* durant les campagnes dans le passage en juillet 1988 et en juillet 1989. La partie supérieure de la colonne d'eau jusqu'au fond de la pycnocline principale était plus chaude et plus salée en 1988. Néanmoins, le transport net d'eau chaude avec $\theta \geq 3°C$ était orienté vers le Nord et était plus important en 1989 (1,8 ± 2,0 Sv) qu'en 1988 (1,0 ± 1,0 Sv). Le transport net (0,65 ± 0,35 Sv) d'eau froide ($\theta < 3°C$) était dirigé vers la Mer de Norvège en 1988, en contradiction avec 1989 où il (- 0,65 ± 0,56 Sv) était dirigé vers l'Océan Atlantique.

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INTRODUCTION

The Arctic Ocean and the Nordic Seas, *i. e.*, the Greenland, Iceland and Norwegian Seas, collectively designated as "the Arctic Mediterranean", play an unique role in the world climate system because of the general oceanographic significance of the dense waters formed there in winter and ventilating the North Atlantic deep and bottom waters (Swift, 1984]. In the Arctic Ocean, the production of relatively warm and saline dense water results from brine release, probably due to freezing on the shelves, and contributes to the Arctic Ocean Deep Water (Aagaard *et al.*, 1985; Rudels and Quadfasel, 1991). This exits from the Arctic Ocean along the Greenland slope (for location of most of geographical sites mentioned in this paper, *see* Fig. 1, redrawn from Perry *et al.*, 1980) to mix with the Greenland Sea Deep Water formed probably through haline convection in a weakly stratified water column of the Greenland basin (Clarke *et al.*, 1990). The final product is a nearly homogenous water mass (θ = -1.05°C and S = 34.91), the Norwegian Sea Deep Water (NSDW) found below 2000 m depth in the Norwegian Sea [NS (Swift and Koltermann, 1988)]. As the entire deep water system of the Arctic Mediterranean, this water mass is constrained by the continental ridges between Greenland and Scotland (Fig. 1) to circulate internally.

NSDW is usually defined as including a body of somewhat warmer and more saline water ($\theta \leq$ - 0.5°C or even $\theta \leq$ 0°C and S $\approx$ 34.92) probably resulting from cooling of the intermediate waters formed in the Iceland Sea (Swift, 1984) and/or from mixing of waters in the volumetrically dominant range of properties, upward into the intermediate waters, as it is displaced by new deep water (Aagaard *et al.*, 1985). This warmer fraction of NSDW, found at the sills, is thought to be of considerable importance for the production of deep water in the North Atlantic (NA). It is easily traced in the Faeroese Channels, comprising the Faeroe-Shetland Channel (FSC) and the Faeroe-Bank Channel (Fig. 1), and overflows the Faeroe Bank sill at a maximum depth of 830 m.

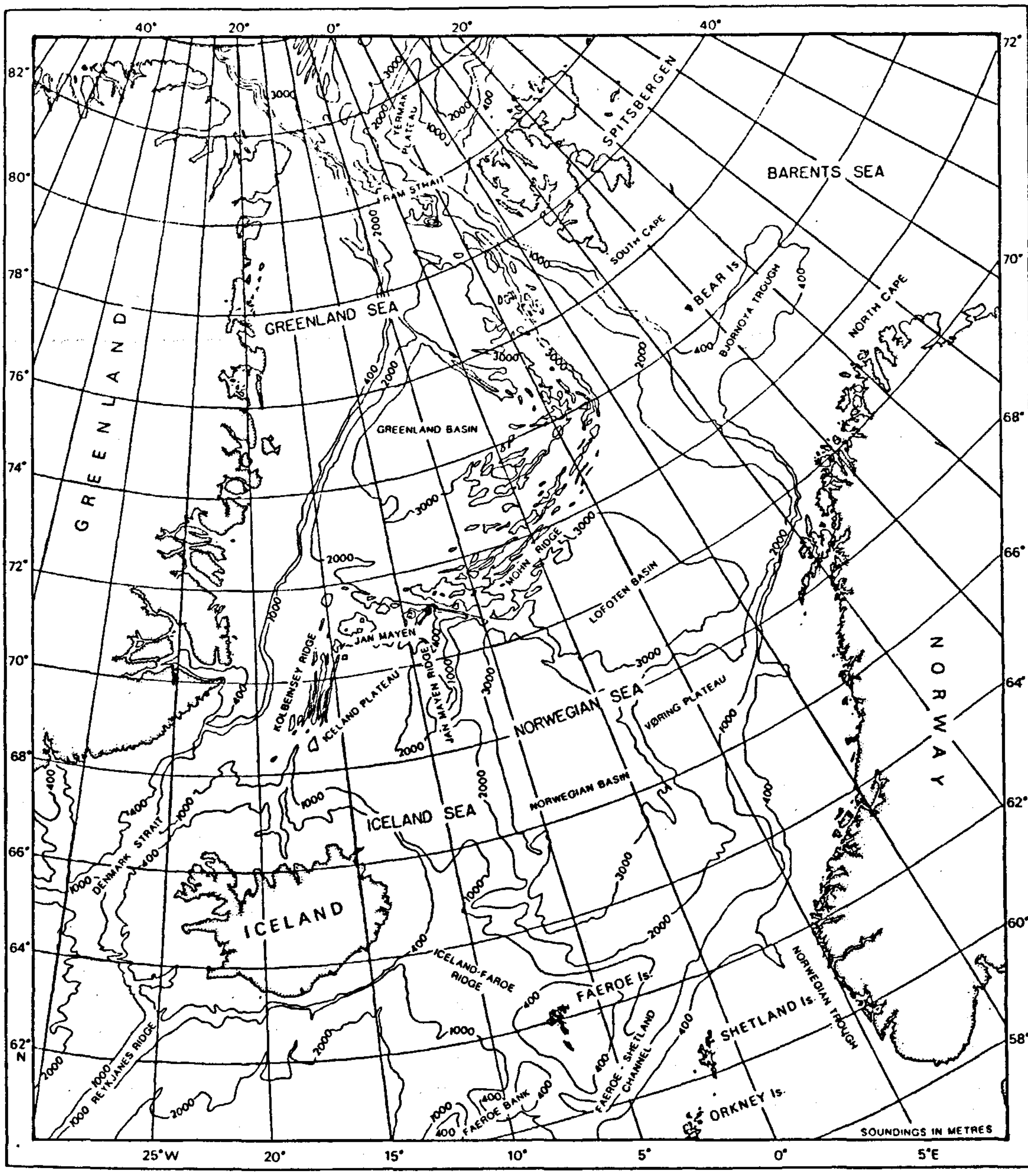


Figure 1

Topography and geographical names of major features in the Nordic Seas (redrawn from Perry et al., *1980). Depth contours are in metres.*

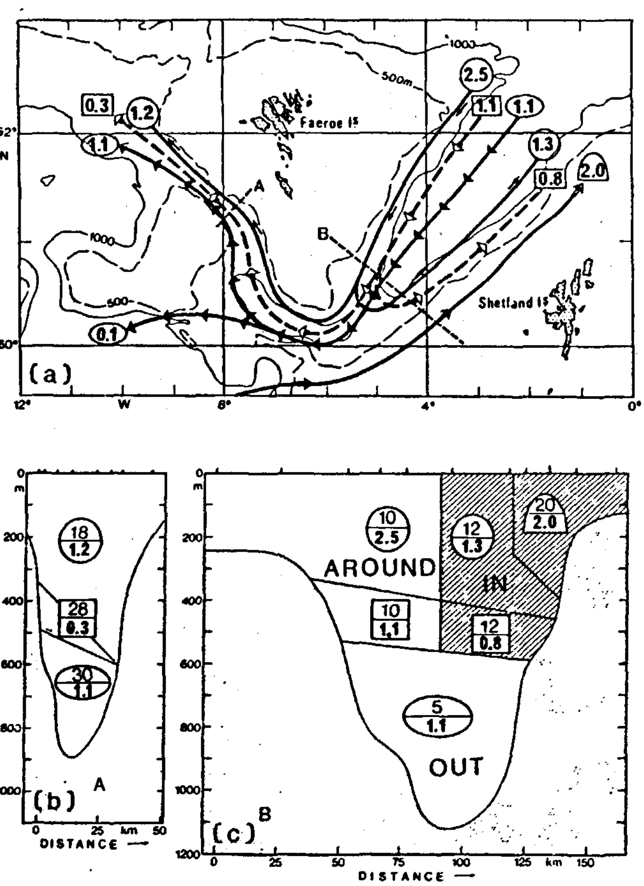


Figure 2

Scheme (from Dooley and Meincke, 1981) of the amounts and routs of the various water mass transports in the Faeroese Channel: a) locations of cross-sections A and B, routs of the transports and transports in Sv (circle - MNAW, square - intermediate waters, ellipse - NSDW, semi-ellipse - NAW); b) section A across the Faeroe Bank Channel; c) section B across the Faeroe-Shetland Channel. The symbols in (b) and (c) represent the same water masses as in (a) with the upper number giving the mean speed and the lower one the transport. Sections A and B face into the Norwegian Sea.

Another primary outflow to NA occurs in the Denmark Strait in the East Greenland current. This current carries not only the low salinity cold surface waters from the Arctic Ocean at the rate of about 3.5 Sv (1 Sv = 10^6 m^3s^{-1}) (Hopkins, 1988) but also contributes in great measure to the overflow of dense waters in the intermediate range of temperatures [$0°C \leq \theta \leq 3°C$ (Swift, 1984)]. This water is in part formed in winter at the surface in the Iceland Sea (Swift, 1986). The overflow at the Iceland-Faeroe ridge is of less importance because the bathymetric constriction there is such that the water flow favours the along-ridge direction.

The overall outflow of deep and intermediate waters between Greenland and Scotland may attain as much as 5 Sv some 2 Sv of which may have characteristics (in the broad sense) of NSDW (Hopkins, 1988). The routes of various water masses in the Faeroese Channels (the major outflow path for NSDW) are depicted schematically in Fig. 2, redrawn from Dooley and Meincke (1981). With the aid of dynamic computations constrained with current meter measurements from the "Overflow 73" experiment, these authors estimated the Faeroese Channels outflow of NSDW as 1.1 Sv and that of the intermediate waters as 0.3 Sv. More recent observations give slightly higher values. On the basis of hydrographic data and current measurements, Borenäs and Lundberg (1988) report a 1.5-1.9 Sv outflow of cold waters ($\theta < 3°C$) through the Faeroe-Bank Channel in May 1983. Saunders (1990), combining CTD and ADCP (Acoustic Doppler Current Profiler) data, determined the flux of waters in the same temperature range in May 1987 to be 1.9 ± 0.4 Sv.

The thermohaline circulation and the water mass structure within the whole Arctic Mediterranean strongly depend on the inflow of warm and saline waters from NA. These waters separate themselves from the North Atlantic Current and enter the Nordic Seas, some through the Faeroese Channels and the rest west of the Faeroe Islands, over the Iceland-Faeroe ridge. Both branches continue their way northward together, as the Norwegian Atlantic Current. At the very entrance, the westernmost extraction of the Atlantic waters enters into direct contact with the cold low salinity water of the East Icelandic Current which branches to the east from the East Greenland Current. It results in a sharp front which forms part of the frontal system separating in the Nordic Seas the Atlantic domain from the Arctic one (Swift, 1986).

The heat and water budgets of NS need a total influx of about 8 Sv of the Atlantic waters (Worthington, 1970). However, the strength of this inflow is not well established and estimates vary between 2 and 8 Sv. For a long time the Faeroese Channels have been considered as the major entrance of the Atlantic waters into NS. Dooley and Meincke (1981) obtained, in their dynamic calculations with a level of "no motion" at 550 m, an estimate of 2 Sv transport of waters flowing directly from the south and and another 1.3 Sv northward flow of recirculating fresher Atlantic waters. Notice (Fig. 2) that another 1.2 Sv of fresher Atlantic waters keeps on moving clockwise around the Faeroe Islands. These results do not differ significantly from Tait's (1957) calculations in FSC, relative to the S = 35 ppt isohaline, which gave a mean (from 1927 to 1952) inflow of 2.5 Sv. In agreement with these estimates are also the results of the van Aken's (1988) inversion of hydrographic data collected in the vicinity of the Faeroese Channels in summer of 1983. Van Aken found a 2 Sv net inflow of Atlantic waters through FSC. The clockwise circulation around the Faeroe Islands was absent in his solution. On the other hand, his analysis revealed a much stronger (6 Sv) inflow of warm waters into the NS north of the Faeroe Islands. Hansen *et al.*'s (1986) ten-day (June 1986) mean current meter observations of the eastward transport of the Atlantic waters at S = 35.2 ppt along the north Faeroe shelf gave a value of 2.9 Sv. Saunder's (1990) estimate of the eastward transport of waters with $\theta > 5°C$ is even smaller (0.5 ± 1.0 Sv). His findings are also quite in variance with the van Aken's results as far as the northward transport through FSC is concerned. According to his measurements, the waters with $\theta > 7°C$ flow to the north at the rate of 5.3 ± 1.5 Sv. Gould *et al.*'s (1985) annual (summer 1983-summer 1984) mean current measurements revealed an even stronger inflow of Atlantic waters only on the Shetland shelf (7.5 Sv). The differences in the above and other estimates may be attributed not only to different methods of calculations but also to strong interannual and seasonal variability of the Atlantic influx (Hopkins, 1988). On the seasonal time scale a maximum occurs in winter and a minimum in summer (Gould *et al.*, 1985) sometimes with a secondary maximum in June (Tait, 1957).

The aim of present work is to describe the year-to-year variability of the hydrological regime and the water volume transport through FSC between summer 1988 and summer 1989. The paper is constructed as follows. The next section describes the hydrographic data and measurement equipment. The following one gives general information about hydrological conditions in the channel in July 1988 and 1989. Then, one can find some estimates of the water volume exchange through the channel obtained by geostrophic computations and a simple box-type inverse method. The last section summarizes the results of investigations and gives conclusions of our study.

THE DATA

In summer of 1988 and of 1989 (8-10 July each year), during the second and third oceanological expeditions of R/V *Oceania* to NS (Zielinski and Siwecki, 1989), vertical CTD soundings were carried out for a range of hydrographic stations. In both years, fifteen stations of interest (Fig. 3)

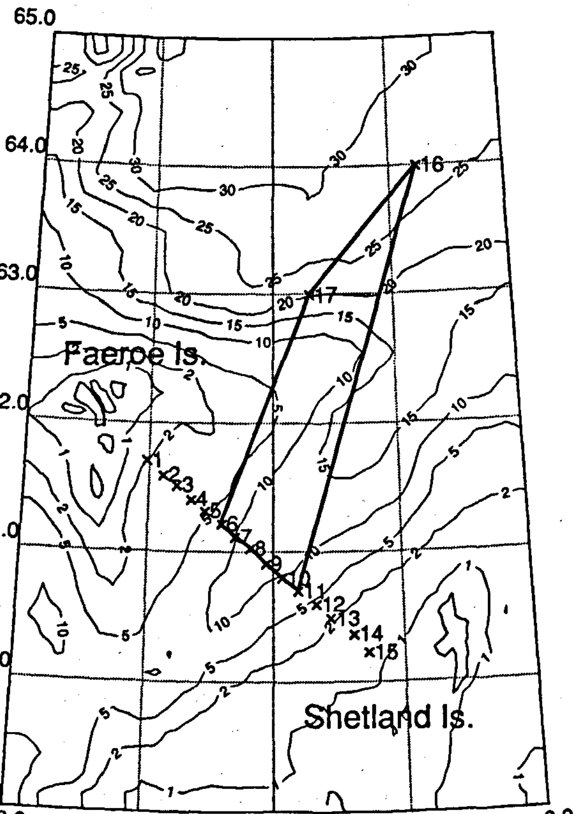


Figure 3

Schematic location of stations at the research area in 1988. Depths of the isobaths are in hectometres.

were occupied in a cross-section perpendicular to FSC and another two (stations 16 and 17) to the north of the channel. From now on, the section will be called the Faeroe-Shetland Section (FSS). Stations in FSS are numbered from 1 to 15, beginning from the closest station to the Faeroe Islands. The mean distance between two neighbouring stations in FSS was 18.0 ± 5.5 km in 1988 and 17.9 ± 2.1 in 1989. In 1988 (89), station 17 was located about 200 (210) km away from station 6 and about 195 (145) km from station 16. The position of station 17 was shifted one degree to the east in 1989 in comparison with 1988. The measurements were carried down to the bottom or to the depth of 1 000 m if the bottom was not reached. Station 17 in 1988 is an exception. Its last observational level was 670 m. The six deep stations in FSS (stations 6-11) will be referred to as the deep section (DS6) while the four deepest ones (stations 7-10) as DS4.

The CTD vertical soundings were done using Guildline Instruments sonde model 8 770 with the following characteristics: pressure sensor-accuracy 0.123 dbar and time constant 1 ms; temperature sensor-accuracy 0.00152°C and time constant 60 ms; measurements of conductivity were made indirectly by coefficient CR (ratio of *in situ* conductivity to its laboratory calibration value calculated for constant temperature and pressure) with accuracy $195. 10^{-6}$ and time constant 60 ms (Operating manual for Model 8770, portable CTD system, 1988). The salinity of sea water was calculated by the aid of standard formulae (UNESCO technical papers in marine sciences, 1983). The data were de-spiked and corrected for temperature and salinity by using a digital low-pass filter (Hamming, 1977). The absolute accuracy of the data after correction was for temperature < 0.02°C, for salinity < 0.01 and < 0.01 σ units for density, also calculated using the standard formulae.

GENERAL FEATURES OF HYDROLOGICAL REGIME

A preliminary discussion of the hydrological regime in the Faeroe-Shetland Channel in summer of 1988 was presented in (Druet and Jankowski, 1991; Jankowski, 1991). Here, we provide the more detailed information that is needed for comparison of the thermohaline conditions between the summer periods of 1988 and of 1989. In Figures 4, 5 and 6 the distributions of the sea water potential temperature, salinity and potential density excess ($\sigma_\theta = \rho\ (\theta, S, 0) - 1\ 000\ \mathrm{kgm^{-3}}$) along FSS in both summer periods are displayed. These distributions show some differences between the two periods, especially in the deepest part of the cross-section, say below the isohaline S = 35 (dashed line in Fig. 5). At the lower depth levels in the year 1988, in which the deep waters are on the average less saline (*see* the depth mean θ and S profiles in Fig. 7), we find a tongue of relatively low saline waters at the Faeroe Islands side. This structure with $S \approx 34.88$, is in agreement with van Aken and Eisma's (1987) findings for summer 1983. Their picture of the salinity distribution in the $\sigma_\theta = 27.9$ plane leads to the conclusion that the fresher deep and intermediate waters originate from the NS. A geostrophically constrained outflow of these waters favours the right-hand side (looking downs-

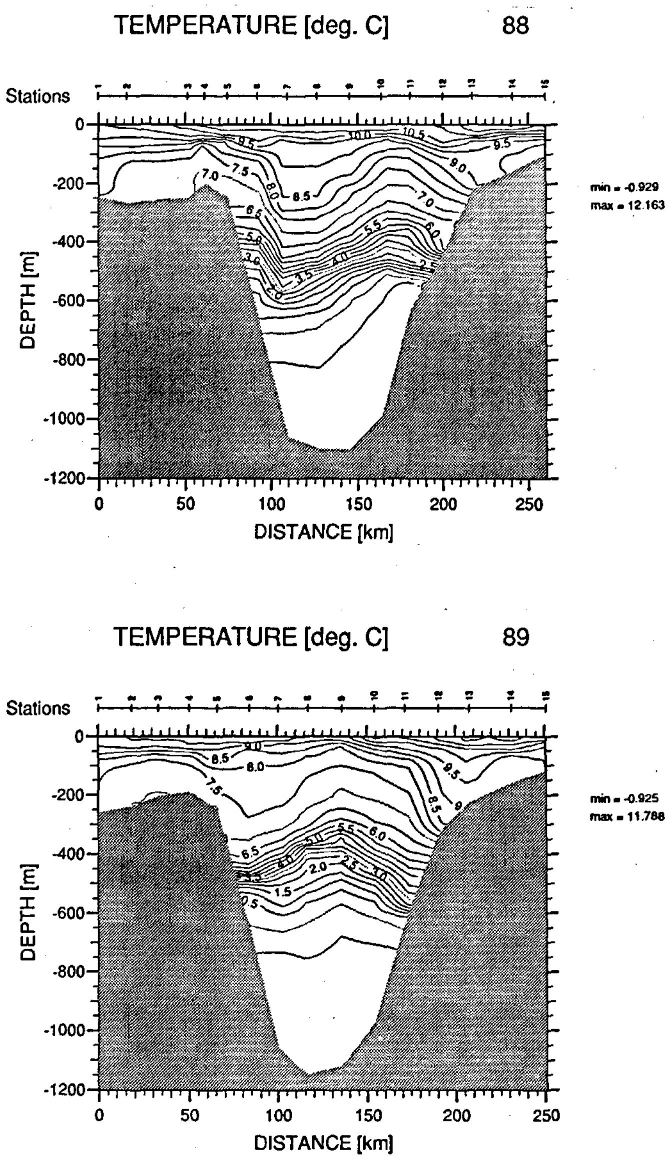


Figure 4

Distribution of potential temperature (°C) versus *depth (m) along the cross-section in the Faeroe-Shetland Channel in summer of 1988 and in summer of 1989.*

tream) and causes the interface separating the upper and lower waters to tilt up at the Faeroe slope. Such a topography of isotherms and isopycnals (here the salinity is of minor importance for the density) can be also found in the 1987 Saunders' (1990) data set. In the 1988 section one can see just the opposite tilt, extremely pronounced along the $\theta = -0.5°C$ isotherm (Fig. 5).

The upper waters are warmer in 1988 ($\theta_{max} = 12.16°C$) compared to 1989 ($\theta_{max} = 11.79°C$) and more saline, down to the bottom of the main thermocline and halocline, respectively (Fig. 7). They also reach deeper levels in that year. The mean depth of the S = 35 isohaline is about 440 m in 1988 and 405 m in 1989. This suggests a stronger inflow of the Atlantic waters in 1988 (not necessarily through FSS). The presence of warmer surface waters in 1988 results in a rather steeper seasonal pycnocline (Fig. 6) in that year, enhanced at the Shetland Island by the occurrence there of a very low salinity surface water (Fig. 5). On the other hand, the main pycnocline seems to be better pronounced in1989.

The most important common feature of the two sections is the presence of strong disturbances of the runs of isotherms, isohalines and isopycnals, seen nearly through the whole range of depths. They may be caused either by the local wind forcing or by quasi geostrophically-balanced eddies (of a 30 km scale) that propagate with the mean flow (Dooley and Meincke, 1981).

To proceed further with description of the year-to-year variability a θ-S analysis was carried out. In Figure 8, in which the summary θ-S diagrams for the channel stations are given, the following water masses can be identified (Dooley and Meincke, 1981; Johannessen, 1986; Swift, 1986; van Aken and Eisma, 1987):

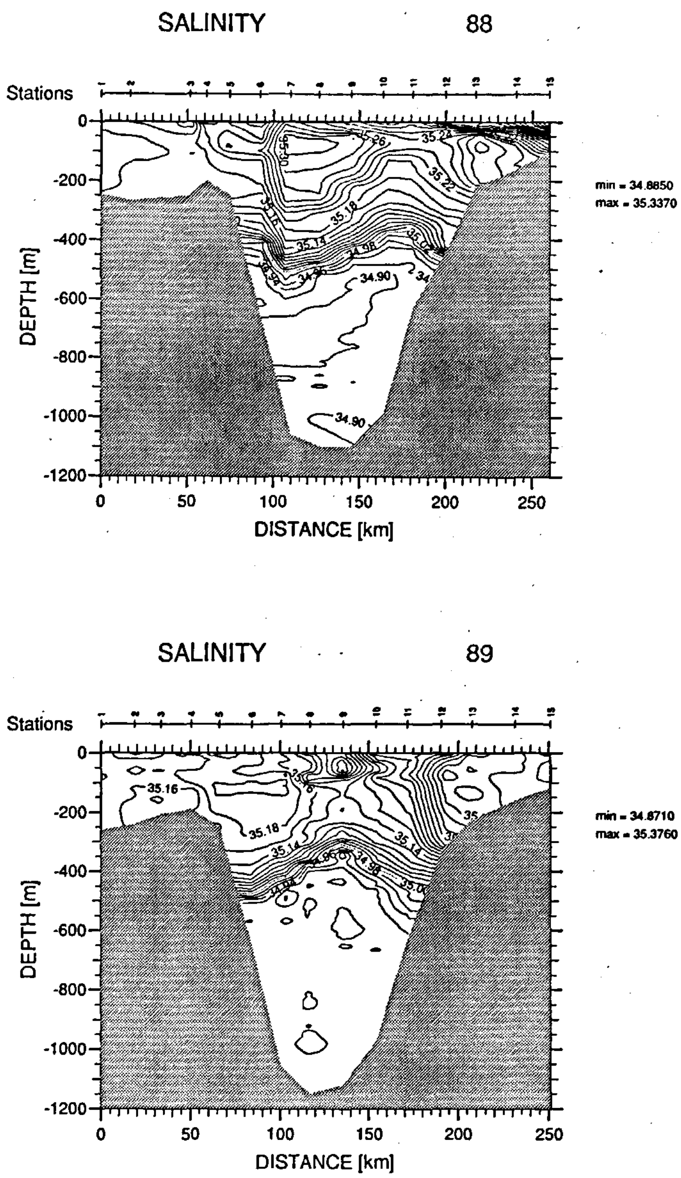


Figure 5

Distribution of salinity versus *depth (m) along the cross-section in the Faeroe-Shetland Channel in summer of 1988 and in summer of 1989.*

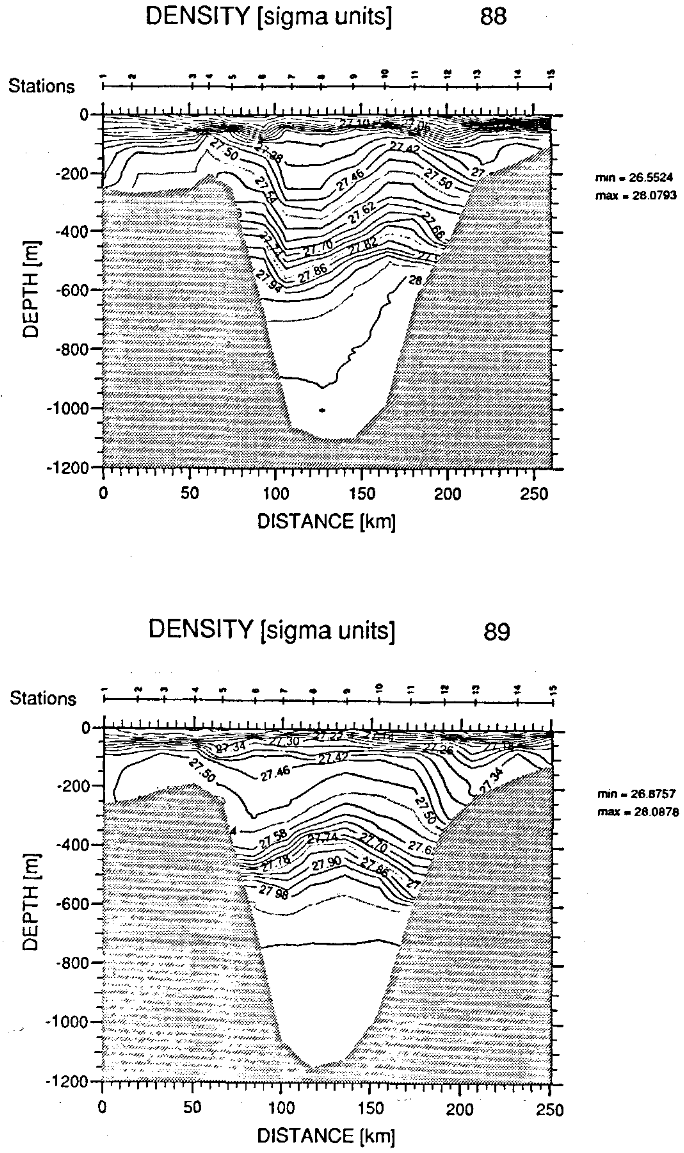


Figure 6

Distribution of potential density (σ units) versus *depth (m) along the cross-section in the Faeroe-Shetland Channel in summer of 1988 and in summer of 1989.*

Surface waters

- North Atlantic Water (NAW) with θ > 9°C and S > 35.3;

- Modified North Atlantic Water (MNAW) with θ > 8°C and S > 35.1;

- Lower Atlantic Water (LAW) with θ ≈ 7.8°C and S ≈ 35.18;

Intermediate waters

A mixture of:

- Arctic Surface Water (ASW) with θ > 2°C and 34.7 < S < 34.9;

- Upper Arctic Intermediate Water (UAIW) with θ < 2°C and 34.7 < S < 34.9;

- Lower Arctic Intermediate Water (LAIW) with 0°C< θ < 3°C and S > 34.9;

with overlying and/or underlying waters.

Deep waters

- Norwegian Sea Deep Water (NSDW) with θ < 0°C and S ≈ 34.92

as well as fresher surface water types. One should bear in mind that these characteristics are only a crude indication of the range of temperature and salinity of the waters in the channel.

The surface waters are generally of Atlantic origin. They seem to be of two types. Saltier NAW (S_{max} = 35.34 in 1988 and S_{max} = 35.38 in 1989) probably enters the channel directly from the south. From Figures 4 and 5 it follows that, in 1988, there are two branches of NAW, with centres at stations 7-8 and 13. In 1989 only one jet of NAW, with the centre at station 13, is encountered. More diluted MNAW is of more western extraction, probably coming from frontal regions in the Norwegian Sea. The θ-S diagrams clearly show that there are two best salinity fits for MNAW in our data set. The fresher one (S ≈ 35.18), found in both years, suits MNAW in the western stations of the section. The saltier one (S ≈ 35.24), encountered mainly in 1988, is characteristic for stations of more east-

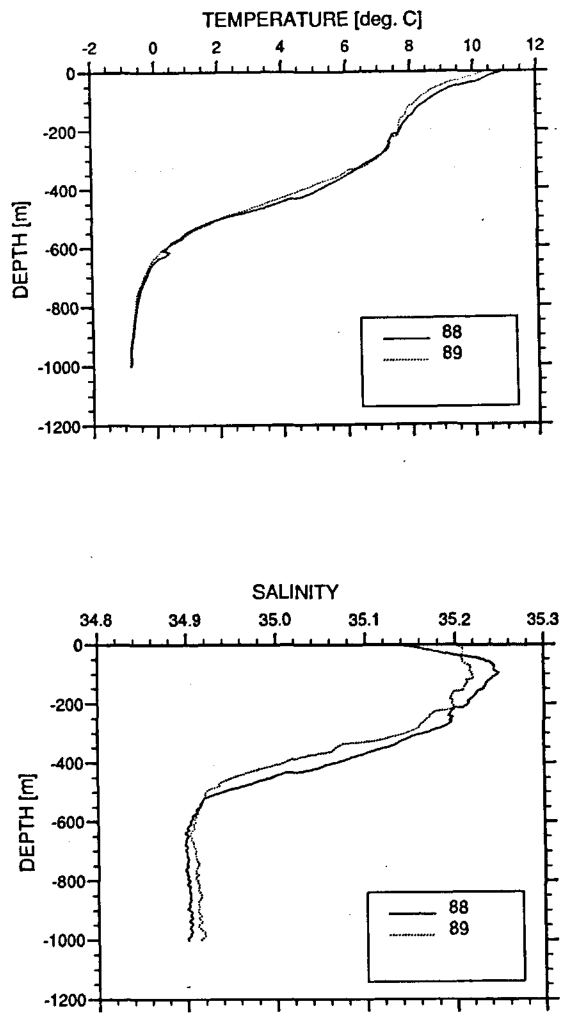


Figure 7

Mean profiles of potential temperature (a) and salinity (b) for the section in the Faeroe-Shetland Channel in summer of 1988 and in summer of 1989.

ern location, *i. e.*, between two cores of NAW. Above the permanent thermocline where NAW and MNAW converge, LAW can be found. It is supposed to be a mixture of colder parts of Atlantic surface waters with other water types as, *e. g.*, the Labrador Sea water (van Aken and Eisma, 1987). As noticed before, in 1988, a surface water with the salinity reduced below 35.0 (S_{min} = 34.89) was observed near the Shetland Islands. In 1989, a core of relatively fresh water (S_{min} = 35.01) occurred in the central part of the channel. The intrusion of water with characteristics comparable to those of ASW is noticed only in 1989. This water, found in that year at the intermediate depth levels (a peak at θ = 4°C in Fig. 8), probably comes from the surface layer of the Iceland Sea (Swift, 1986). Neither the low salinity UAIW of arctic or polar origin (Blindheim, 1986) nor the high salinity LAIW, being a result of winter cooling of Atlantic waters, possibly in the Greenland Sea (Swift, 1986), are observed in their original form. Nevertheless, their mixture with LAW and NSDW is clearly present in the channel. The description of the coincident existence of low and high salinity intermediate waters in the channel was given, *e. g.*, for the summer of 1983, by van Aken and Eisma (1987). In 1989, this structure, however highly modified (S_{min} = 34.87 at station 7), is common for all deep stations (7-10) while in 1988 it is found only at station 10. The change in the structure of intermediate waters in the channel from year to year can be due to much weaker occurence, in 1988, of the double extremum configuration of these waters in the south-eastern NS from where they are supposed to outflow. To support this hypothesis the θ-S diagrams for stations 16 and 17 are plotted in Figure 9. As one may expect, the layer of the intermediate waters is at these stations shallower than in FSS. In 1988, the min-max salinity structure of the intermediate waters is found only at station 16 and is not so pronounced as in 1989. In the latter year, the salinity maximum is very strong at station 17 at the depth of 300 m (S_{max} = 34.98), and is diluted before reaching the channel (S_{max} = 34.93), but still it is there at the depth of about 550 m. From Figure 9 it follows also that the surface waters were much more saline and warmer to the north of the channel in 1988, contributing to the concept of a farther northern extent of Atlantic waters in that year.

In the deep part of the cross-section one can observe exclusively NSDW. It is found below the salinity minimum that results from the intrusion of cold arctic intermediate waters typical for the Norwegian Sea (compare Figures 8 and 9). It is worth noting that waters at 1 000 m are cooler (about 0.3°C) in the channel than at stations 16 and 17. It must be due to an "upwelling" of NSDW encountering the ridge on its way to the south.

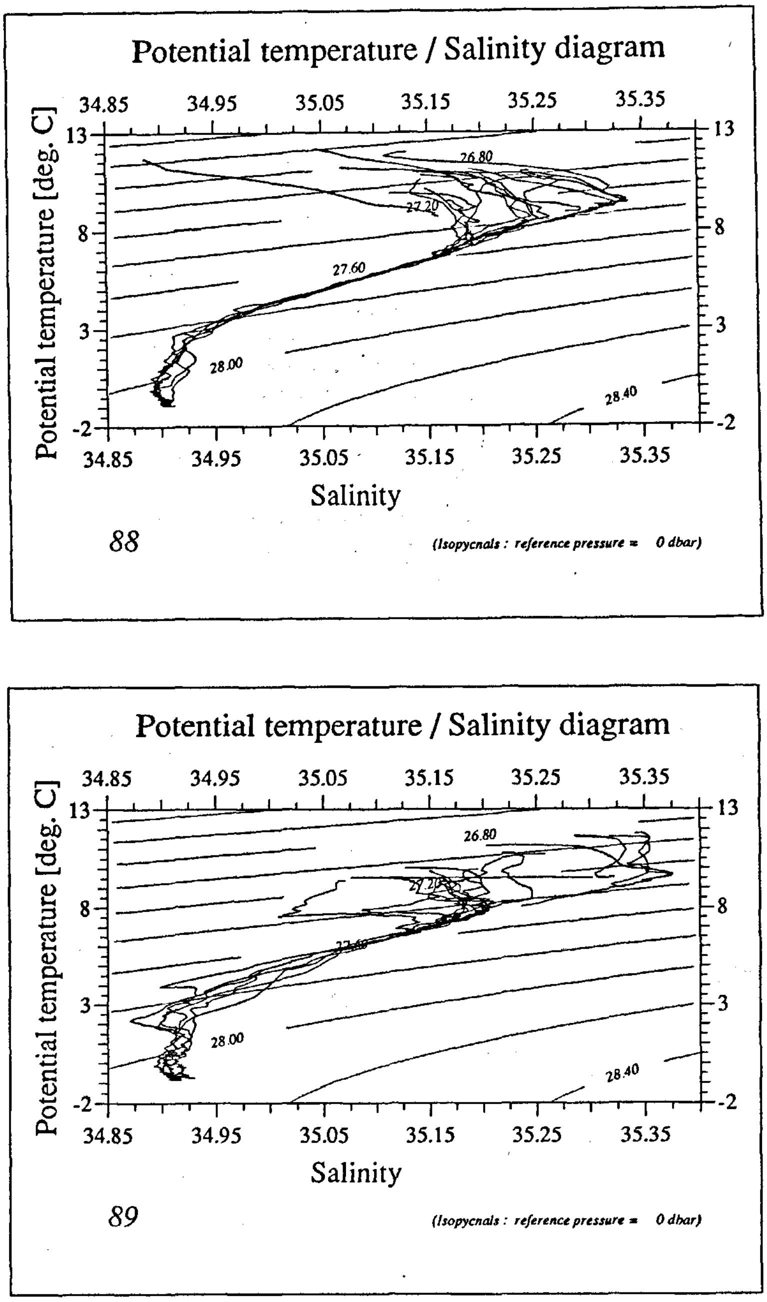


Figure 8

Summary θ-S diagram for the section in the Faeroe-Shetland Channel in summer of 1988 (a) and in summer of 1989 (b).

In the next section we will consider the water column in the channel in terms of four basic layers. The mean densities at the mean depth of certain isotherms, usually taken as the bounds on different water mases (Borenäs and Lundberg, 1988; Saunders, 1990), were selected as the boundaries of the layers. The temperature and density ranges of each layer are marked in Figures 4 and 6 by dashed lines. The upper layer (layer I) with $\theta \geq 7$°C, which is formed by the very thin mixed layer and mainly by the seasonal thermocline, contains NAW, MNAW and LAW as well as fresher costal water types. The lower density boundary of this surface layer is σ_θ = 27.54 and reaches the point (the depth of about 300 m) where the cross-channel θ-S diagrams for all stations converge (Fig. 8). The second layer (layer II),

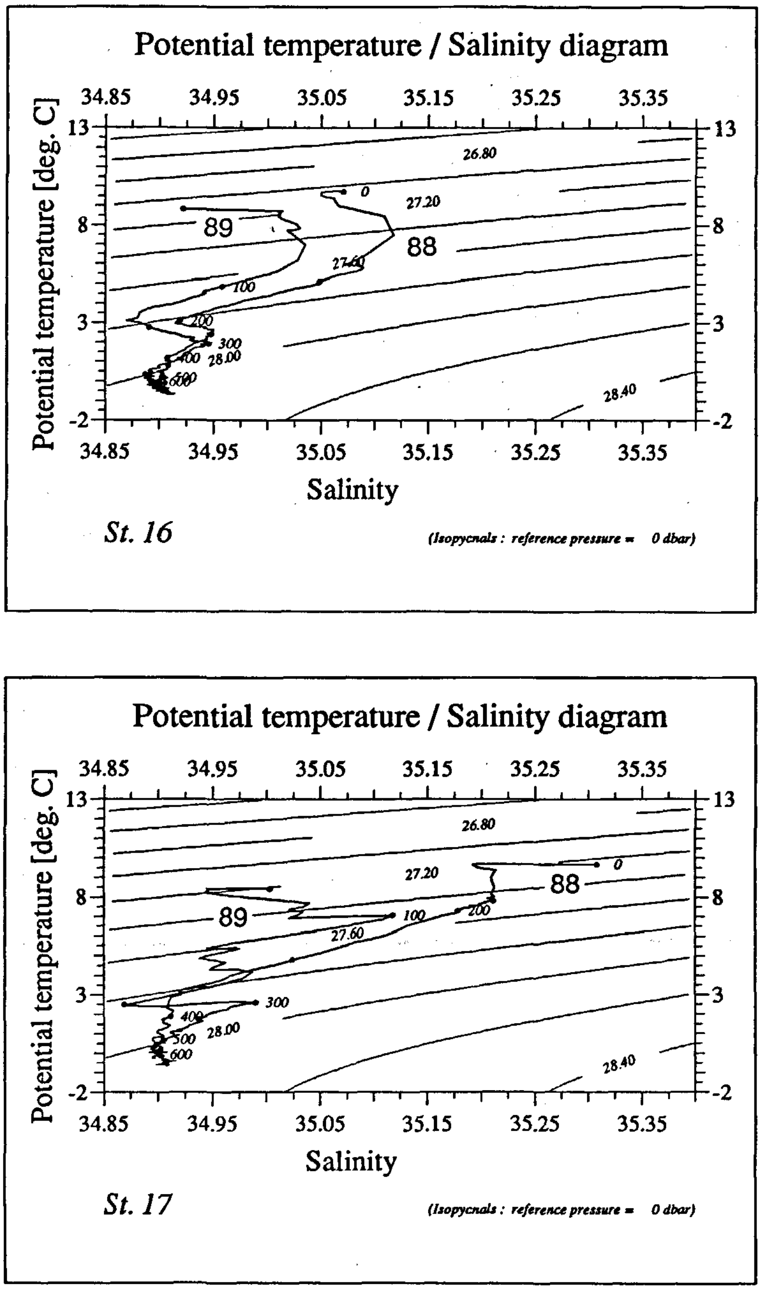


Figure 9

θ-S diagrams for stations 16 and 17 in the Norwegian Sea in summer of 1988 (a) and in summer of 1989 (b).

bounded by θ = 7°C and θ = 3°C or by σ_θ = 27.54 and σ_θ = 27.82, contains waters found between the seasonal and the main pycnocline and, first of all, in the main halocline. From Figure 8 it turns out that these waters are a mixture of Atlantic waters with intermediate waters. The lower density limit of layer II, found at the mean depth of about 480 m, converges, in 1988, with the point where θ-S diagrams for different stations begin to diverge. The third layer (layer III), with the boundaries taken to be θ = 3°C (σ_θ = 27.82) and θ = 0°C (σ_θ = 28.02), comprises the intermediate waters with their salinity minimum and maximum. On the average it reaches the depth of 640 m where a second salinity minimum is encountered. The deepest layer (layer IV) begins at σ_θ = 28.02 and ends at the bottom. It consists mainly of waters with characteristics comparable to those of NSDW (θ < 0°C).

GEOSTROPHIC VELOCITIES AND TRANSPORT

On the basis of the data set of sea water temperature and salinity (and hence density) the geostrophic current velocities and the volume transport of water masses through FSS were calculated. To estimate current velocity, the dynamic method (*e. g.*, Fomin, 1964) was applied. To extrapolate the geostrophic shear in the interval between the different bottom depths, z_A-z_B, of each CTD pair (A,B), four different approaches have been studied. They are: 1) suppose that the vertical geostrophic shear is zero from z_B towards z_A (classical dynamic method); 2) decrease the geostrophic shear linear to zero from z_B to z_A; 3) maintain the geostrophic shear constant from z_B to z_A; and 4) calculate the geostrophic shear between stations A and B from the density structure, obtained by maintaining the slope of the isopycnals from z_B downwards to the bottom while below the bottom an imaginary no-shear water mass with horizontal isopycnals is assumed. In all four methods the bottom shallower than the reference level is thought to be the level of no motion. Our experience allows us to support the conclusion of van Aken (1988) that the fourth method, basing upon the most hydrological information, gives the best transport estimate.

A level of no motion (z_o) was chosen with the aid of a procedure similar to that of Fiadeiro and Veronis (1982). Two versions were applied. First, varying the reference level, the volume transport imbalances were calculated for the three upper layers (layers I, II and III from the previous section) separatly in the following polygons:

POL8 = DS6 + station 17 + station 16
POL7 = DS6 + station 17
POL6 = DS4 + station 17 + station 16
POL5 = DS4 + station 17.

Then, for each polygon and each reference level, the mean (over layers) square imbalance T was found. Finally, z_o (= const) was chosen at the level for which T (z) was minimum. In the second version (R-experiments: POL8R, POL7R, POL6R, POL5R) the z-th coordinate was replaced by σ_θ and the whole procedure was repeated. It gave a variable z_o [= z (σ_θ =const)] and a function T (σ_θ) instead of T (z). The functions T (z) and T(σ_θ) were computed with the intervals 10 m and 0.03 sigma units and normalized to vary between 0 and 1 in such a way that 0 represents no mass imbalance and 1 stands for a maximum departure from the mass conservation for all layers.

Next, the mass conservation was imposed on the three upper layers. The correction velocities to bring the layers into mass balance were found applying a least square principle. For a description of the method one may consult the pioneer work by Wunsch (1978) or, in the context of other inverse schemes to calculate velocities from CTD data, the reviewing paper by Schlichtholz (1991).

No unique reference level was obtained. However, in each experiment one (POL6, POL7, POL5R and POL6R in 1988, and POL5, POL6, POL8, POL5R and POL6R in 1989) or two (POL5, POL8, POL7R and POL8R in 1988, and POL7, POL7R and POL8R in 1989) distinct minima were undoubtly found on the imbalance functions T (z) and T (σ_θ). As an example T (z) for POL6 an POL8 and T (σ_θ) for POL6R and POL8R are plotted in Figures 10 and 11. The depths (densities) of the minima varied in all experiments from 200 (27.65) to 610 m (27.95 sigma units). Let us recall here that van Aken (1988) estimated z_0 as 400 m for his much larger data set from the summer 1983. Saunders' (1990) CTD casts combined with ADCP measurements in 1987 located the reference level at z_0 = 300 m. Dooley and Meincke (1981), for their "Overflow 73" data, used a deeper level z_0 = 550 m. Our analysis gives, on the average, z_0 = 400 m (27.80 sigma units) for 1988 and z_0 = 290 m (27.75 sigma units) for 1989. Therefore, the reference level is found in the lower part of layer II in 1988 and in its upper part in 1989.

In Figure 12, the depth distribution of the geostrophic current velocity component perpendicular to the plane of the cross-section, estimated using $z_0 = z\ (\sigma_\theta = 27.80)$ in 1988 and $z_0 = z\ (\sigma_\theta = 27.75)$, is presented. As one may expect from the analysis of the thermohaline structure of water (Fig. 4, 5 and 6), there are some differences in the distribution of the geostrophic velocity in the channel in summer of 1988 and 1989. In 1988, one can distinguish two flows into NS, with a return flow between them in the central part of the section. The maximum current velocity does not

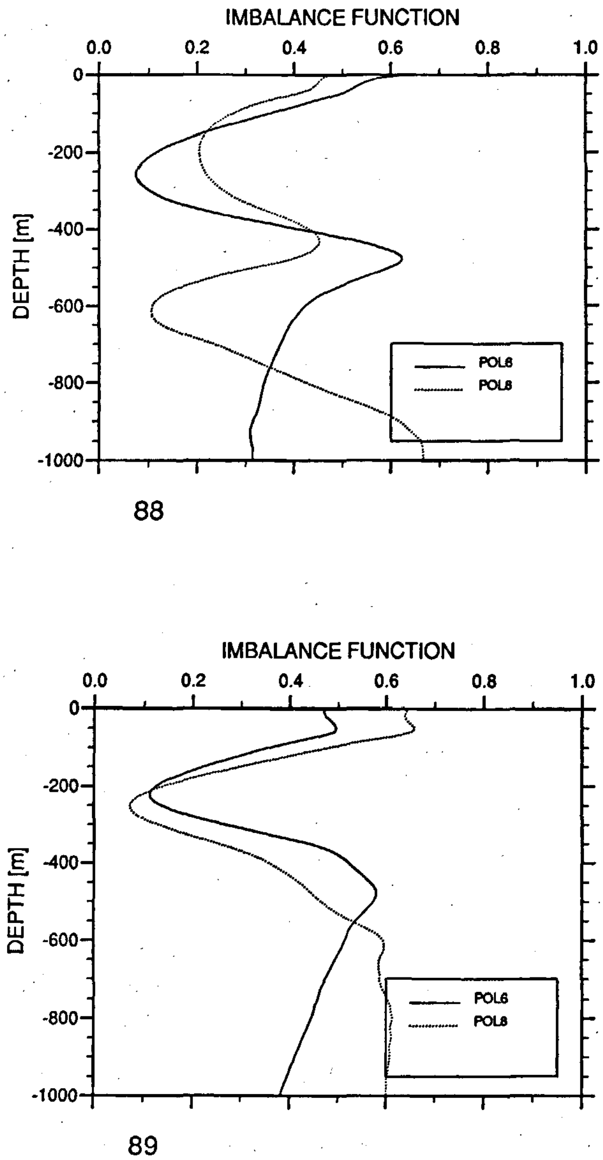


Figure 10

Imbalance function versus *depth for POL6 and POL8 experiments. For explanation* see *the text.*

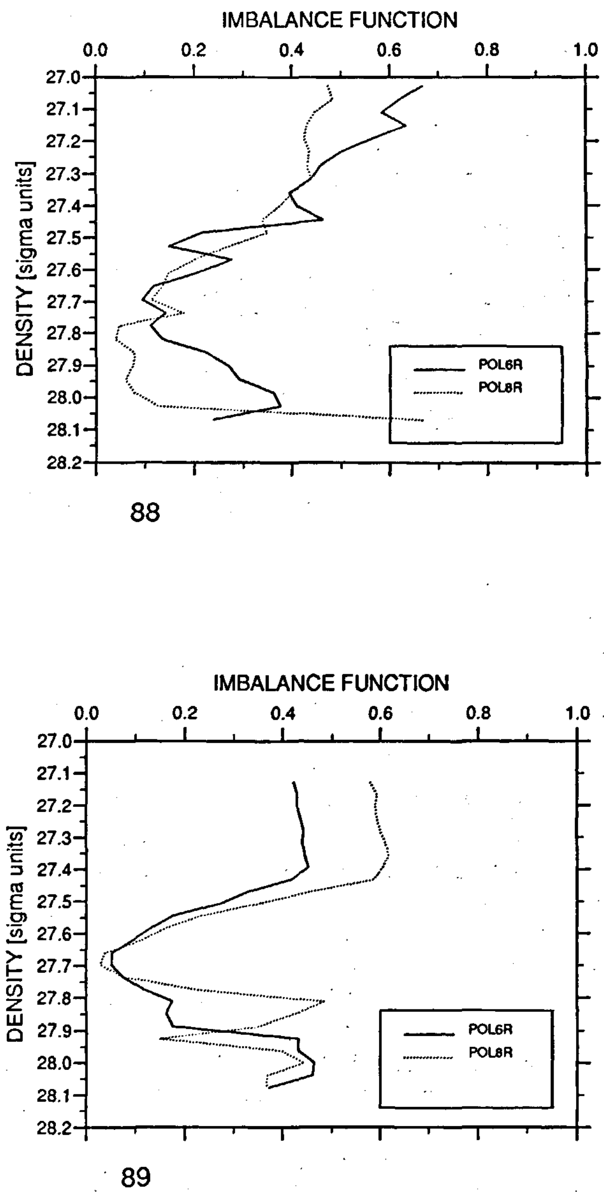
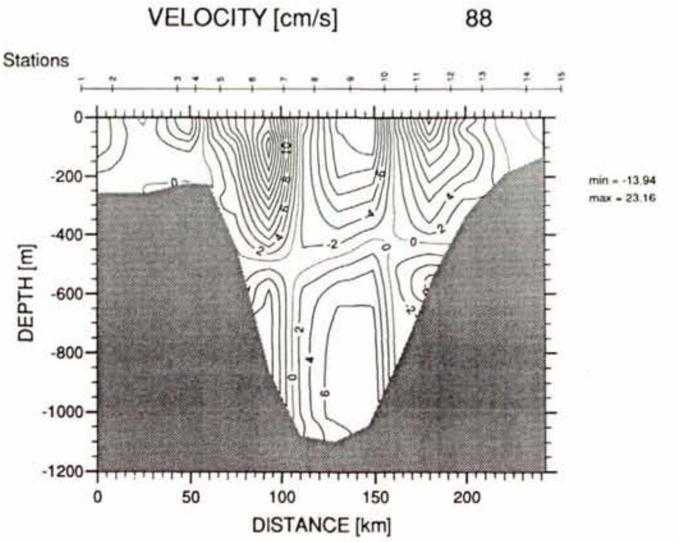


Figure 11

Imbalance function versus *potential density for POL6R and POL8R experiments. For explanation* see *the text.*



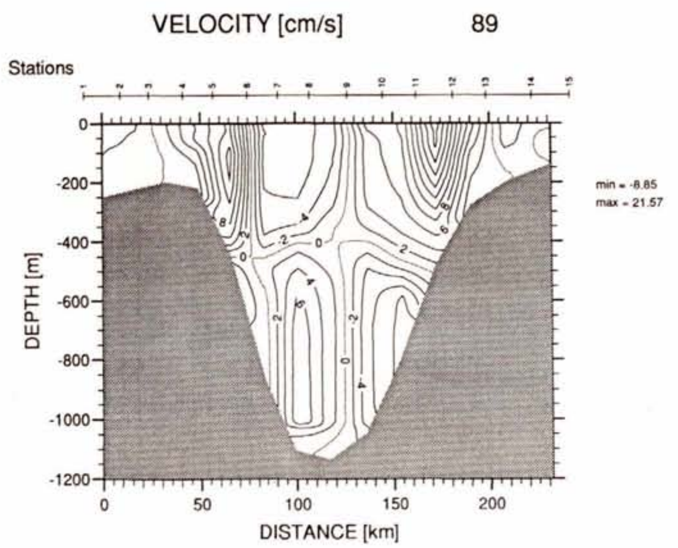


Figure 12

Distribution of geostrophic current velocity component (cm/s) perpendicular to the plane of the cross-section in the Faeroe-Shetland Channel in summer of 1988 and in summer of 1989 (positive values - towards the Norwegian Sea). Level of no motion at $z_o = z(\sigma_\theta) = 27.80$ in 1988 and $z_o = z(\sigma_\theta) = 27.75$ in 1989.

exceed 25 cm/s [compare with a maximum > 1 m/s (Hopkins, 1988)], reaching the value of 23 cm/s in the flows into NS and the value of 12 cm/s in the return flow. Analyzing Figure 12 together with Figures 4 and 5, one can notice that the surface part of the eastern flow into NS consists of NAW and MNAW with S ≈ 35.24. The presence of this MNAW in the probably recirculating flow agrees with van Aken's (1988) findings. The occurrence of NAW in both the western flow into NS and in the recirculating flow can be thought of as a result of the existence of a baroclinic eddy in the channel, split from the main branch of Atlantic waters going to the north. In the western part of the section, near the Faeroe Islands, one can see two small return flows with maximum velocity not being greater than 8 cm/s, containing MNAW (S ≈ 35.18), and partially recirculating into NS.

The picture of the current velocity distribution in summer of 1989 is slightly different. The maximum velocities of the flows into NS and the returning ones are about 22 and 9 cm/s, accordingly. The western flow into NS is very weak, comparing to the one in 1988, and is moved to the west. The main return flow is weaker, and its axis is also moved towards west. In the vicinity of the Faeroe Islands, only one return flow can be observed, and it is rather weaker than in 1988. In all these flows MNAW with S ≈ 35.18 is found. The eastern flow into the Norwegian Sea is also weaker than in 1988 but wider - it covers almost the whole eastern part of the cross-section. In its easternmost part, it contains NAW, which appears to recirculate over the shelf of the Shetland Islands, as in 1988 and in van Aken (1988). The cyclonic feature, with low salinity at the surface, found in DS4, seems to be advected from NS where a core of low salinity surface water occurred at station 17 (Fig. 9).

The deeper part of the cross-section is marked by a recirculation pattern of the deep and intermediate waters flow, once again moved in 1989 to the West. The presence of topographically trapped eddies in this part of the channel was reported by van Aken (1988). It is worth noting here that, in 1988, the northward branch of the deep flow is as strong as the southward one (> 6 cm/s) but it is broader than the latter one.

To balance the mass in the three upper layers, in any of the experiments carried out, correction velocities as small as v_o < 1 cm/s were needed. Such small velocities (on the average 0.5 cm/s in 1988 and 0.25 cm/s in 1989) would not add too much to the description of Figure 12. They will be treated as a contribution to the uncertainty of our transport estimates. It should be noted, however, that in all experiments they were directed into NA for all stations along DS4 or DS6.

The values of the volume geostrophic transport through FSS, calculated for $z_o = z$ ($\sigma_\theta = 27.8$) in 1988 and $z_o = z$ ($\sigma_\theta = 27.75$) in 1989, are presented in Table 1. Layers I and II as well as III and IV were combined there into one layer. The total water flow into NS attains about 4.2 (3.9) Sv in 1988 (89). It is not balanced by the flow into NA which

Table 1

Water volume transport in layers (10^6 m^3/s) through the cross-section in the Faeroe-Shetland Channel in summer of 1988 and 1989. Positive values are assumed for the transport directed towards the Norwegian Sea. Abbreviations: z_o - level of no motion (at the depth of given σ_θ), A.O. - towards Atlantic Ocean, N.S. - towards Norwegian Sea.

z_o	Layer	N.S.	A.O.	N.S. + A.0.
		YEAR 1988		
$\sigma_\theta = 27.80$	I + II	2.79	- 1.80	0.99
	III + IV	1.45	- 0.81	0.64
	I-IV	4.24	- 2.61	1.63
		YEAR 1989		
$\sigma_\theta = 27.75$	I + II	3.03	- 1.21	1.82
	III + IV	0.84	- 1.47	- 0.63
	I-IV	3.87	- 2.68	1.19

Table 2

Net water volume transport in layers (10^6 m^3/s) through the cross-section in the Faeroe-Shetland Channel in summer of 1988 and 1989 for three experiments discribed in the text. Positive values are assumed for the transport directed towards the Norwegian Sea. Abbreviations: z_o - level of no motion, Tr - transport, (1) - the first method of vertical shear extrapolation.

z_0	Layer	Tr	z_0	Layer	Tr
YEAR 1988			YEAR 1989		
$\sigma_\theta = 27.95$	I + II	1.20	$\sigma_\theta = 27.92$	I + II	2.77
	III + IV	0.61		III + IV	- 0.26
	I-IV	1.81		I-IV	2.51
$\sigma_\theta = 27.95^{(1)}$	I + II	0.45	$\sigma_\theta = 27.92^{(1)}$	I + II	0.63
	III + IV	0.62		III + IV	- 0.23
	I-IV	1.07		I-IV	0.40
400 m	I + II	0.72	290 m	I + II	0.04
	III + IV	0.53		III + IV	- 0.75
	I-IV	1.25		I-IV	- 0.71

amounts to about 2.7 Sv in both years. Thus, the net transport of 1.6 (1.2) Sv in 1988 (89) is directed into NS. Its distribution among layers differs from year to year. In 1989 we have a rather usual case with a net inflow (1.8 Sv) of warm waters with $\theta \geq 3°C$ (layers I + II) and a net outflow (0.6 Sv) of cold waters (layers III + IV). The year 1988 seems to be anomalous. A net input occurs, in that year, in both the warm (1.0 Sv) and cold (0.6 Sv) layer.

The water volume transport in the warm temperature range between stations 6-17-16, which take the same value (1.3 Sv) in both years is directed mainly into NS. Thus, recalling that the transport through FSS in that temperature range was greater in 1989, we see that the presence of warmer waters in 1988 does not imply that the Atlantic waters penetrate farther to the North in that year. A stronger recirculation in 1988 suggests that there must be a mechanism blocking the transport of Atlantic water masses somewhere to the north or west of the channel.

Let us dwell for a moment on the uncertainty of the above estimates. We cannot eliminate the presence of tides in our data, probably of the order of 10 cm/s (Saunders, 1990). We can try, nevertheless, to put on our results some error bars due to the choice of the level of no motion and to neglecting the correction velocities. The net transport through FSS is given for three cases in Table 2. First (row 1), we changed $(\sigma_\theta)_o$ from the averaged over experiments value of 27.80 (27.75) to the maximum one - 27.95 (27.92). Next (row 2), the method of extrapolation of the geostrophic shear was switched from 4 to 1. Finally (row 3), method 4 of calculating the shear was restored and z_0 was taken as constant (400 m in 1988 and 290 m in 1989).

Comparing Table 2 to Table 1, one may notice that the warm layer transport is more sensitive to the choice of the reference level than the cold one. Another feature following from the comparison is that the estimates in 1988 are more stable. The maximum differences between the net transport given in Table 2 and those in Table 1 are 0.54 Sv (I + II, 1988), 0.11 Sv (III + IV, 1988), 1.78 Sv (I + II, 1989), and 0.4 Sv (III + IV, 1989). If we add to these figures the transports calculated for the mean correction velocities obtained in the inversions for DS6 (DS4) and streached over the whole FSS, we find the following net transport estimates: 1.0 ± 1.0 Sv (warm waters in 1988), 1.8 ± 2.0 Sv (warm waters in 1989), 0.65 ± 0.35 Sv (cold waters in 1988), and - 0.65 ± 0.55 Sv (cold waters in 1989).

SUMMARY AND CONCLUSIONS

The results of this paper, based on the CTD data measured down to 1 000 m in the cross-section of the Faeroe-Shetland Channel and at two stations in the south-eastern Norwegian Sea, have shown that oceanographic conditions in the channel in July of 1989 differed slightly from those in July of 1988. The sea water in the channel and north of it was warmer and more saline in 1988 down to the bottom of the main pycnoline.

In both years, the water column in the channel could be divided into three groups of water masses, *i. e.*, surface, intermediate and deep. In the surface layer, one could find NAW, MNAW and LAW as well as fresher water types. NAW entered the channel from the south over the shelf slope near the Shetland Islands. A part of this water recirculated back over the shelf, especially in 1989. In 1988, a baroclinic eddy separated from the main branch of waters carrying NAW. It was encountered in the centre of the channel. MNAW of more western origin, found in the western side of the section, flowed from the Norwegian Sea. It was meandering in the western side of the section, and to a large extent, recirculated to the Norwegian Sea in both years. In 1988, a low salinity water type influenced the hydrological regime of the surface waters only very close to the Shetland islands. In 1989, a core of highly diluted surface waters was present in the central part of the section. It was transported from the Norwegian Sea, probably from the Arctic front extended farther to the east and/or south in that year. Below NAW and MNAW, LAW - a result of mixing of the Atlantic waters with underlying waters outside the channel, was present. The mixture of this type of water with the intermediate waters was found in the main halocline. In 1989, it contained a fraction of ASW which seems to confirm our hypothesis concerning the position of the Arctic front in that year.

In the intermediate waters observed in the channel, a mixture of low salinity UAIW of arctic or polar extraction and of LAIW, originated from winter cooling of the Atlantic waters, with underlying and overlying types was recognizable. The simultaneous presence of both intermediate waters, connected with their transport from the Norwegian Sea, was more strongly pronounced in 1989. The intermediate water salinity minimum was not as fresh as to fit the trend, reported from the literature, of extremely low salinity of intermediate waters in the channel since 1972 (van Aken and Eisma, 1987). The intermediate waters were squeezed on the shelf slopes were the Atlantic waters could come into direct contact with the deep water. Below the salinity maximum connected with LAIW, the second salinity minimum occurred. It resulted from the advection of the intermediate waters from the deep regions of the Norwegian Sea. The deep water contained exclusively NSDW. In 1988, it tilted up at the eastern ("wrong") side of the channel, a fact probably due to eddy activities.

Dynamic calculations obviously revealed a strong dependence of the results on the choice of the reference level and the method of extrapolating the geostrophic shear along the uneven topography. A varying level of no motion was chosen at the depths of an isopycnal for which the mass imbalance in three boxes was minimum. The mass conservation in the bottom layer could not be imposed because of the lack of data between the depth of 1 000 m and the bottom, much deeper than 1 000 m at the two Norwegian Sea stations. The average over several experiments values were taken, *i. e.*, $\sigma_\theta = 27.80$ for 1988 and $\sigma_\theta = 27.75$ in 1989. The vertical geostrophic shear was extrapolated in a way which uses as much information on the varying vertical structure as possible. Computations with other configurations of parameters served as an estimate of errors to which the results of some inversions were added up. The correction velocities were too small (< 1 cm/s) to analyse them. They were, however, permanently directed into the Atlantic Ocean. In both years the total volume transport was directed into the Norwegian Sea (1.6 Sv in 1988 and 1.2 Sv in 1989). The vertical structure of this net flux differed much between the two years. The transport of warm waters ($\theta \geq 3$°C) into the Norwegian Sea in 1989 (1.8 ± 2.0 Sv) rather exceeded that in 1988 (1.0 ± 1.0 Sv). Thus, the presence of warmer waters in the channel in 1988 did not imply the farther northern extent of Atlantic waters in the Nordic Seas in that year. Our data set seems to be, however, too poor to give a complete picture of the circulation scheme even in the closest vicinity of the channel.

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