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Observations of overflow on the Iceland Faeroe Ridge*

by

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With 8 figures and 1 table

Beobachtungen des Overflow auf dem Island-Färöer-Rücken

Zusammenfassung

Während der internationalen ICES-Expedition „Overflow '73“ wurden von „F. S. Meteor“ aus Vertikalprofile von Temperatur T, Salzgehalt S, Licht-Attenuation A und Strömung gemessen. Das relativ kleine Meßgebiet befand sich südlich und westlich eines Einschnittes von 485 m Satteltiefe im Island-Färöer-Rücken (Abb. 1 u. 3). Es zeigte sich, daß Overflowwasser ($T \leqslant 2$ °C, $S \leqslant 35‰$, $A \geqslant 30\%$) und atlantisches Wasser ($T \geqslant 5$ °C, $S \geqslant 35{,}05‰$, $A \leqslant 28\%$) durch starke vertikale und horizontale Gradienten dieser Größen getrennt waren. Die Strömungen in der Overflowschicht waren südlich und westlich des Einschnittes nahezu isobathenparallel und nach Westen gerichtet. In der Nähe der Satteltiefe zeigten nördliche Strömungen mit starken Schwankungen in Richtung und Stärke einen Rückstrom des kalten Wassers in Richtung auf die Norwegische See an. Es werden einige Aspekte diskutiert, die die Vermischung durch die Schicht der starken vertikalen Gradienten betreffen. Richardson-Zahlen von 13 Stationen deuten an, daß die Strömung in dieser Schicht „fast turbulent“ ist.

Summary

During the international ICES-expedition "Overflow '73" profiles of temperature T, salinity S, light attenuation A and currents were measured by R. V. "Meteor" in a small area south and west of a channel across the Iceland-Faeroe ridge with 485 m sill depth (figs. 1 and 3). Sharp gradients, both vertical and horizontal, in T, S and A separated a band of overflow water ($T \leqslant 2$ °C, $S \leqslant 35‰$, $A \geqslant 30\%$) from the surrounding Atlantic water ($T \geqslant 5$ °C, $S \geqslant 35{,}05‰$, $A \leqslant 28\%$). Current measurements within the overflow layer indicated an outflow parallel to the isobaths south and west of the channel, and a back flow of cold water towards the Norwegian Sea in the northern part of the area. This back flow appeared to be highly variable in direction and intensity. Some aspects of mixing across the sharp vertical gradient layer are discussed. Richardson-numbers in this layer have been calculated for 13 stations. They seem to indicate marginal stability in the interface.

1 Introduction

The chain of submarine ridges between Greenland, Iceland, the Faeroe Islands and Northern Scotland divides the Norwegian Basin with more than 1500 m depth from the North Atlantic Ocean with more than 4000 m depth. The water balance of the Norwegian Sea requires an outflow of 6 Sverdrups in the bottom layers (Worthington 1970). From numerous investigations (e. g. Dietrich 1967, Meincke 1972) on the ridge it seems to be established that this outflow is highly variable and much weaker in summer than in winter.

In August/September 1973 the international expedition "Overflow '73" with 13 ships of 8 nations was aimed at studying the kinematics and dynamics of the overflow processes. The observations presented in this paper were taken from R. V. "Meteor" on her first leg (August 7 to August 28) in the region

*ICES-Expedition Overflow '73, contribution No. 2

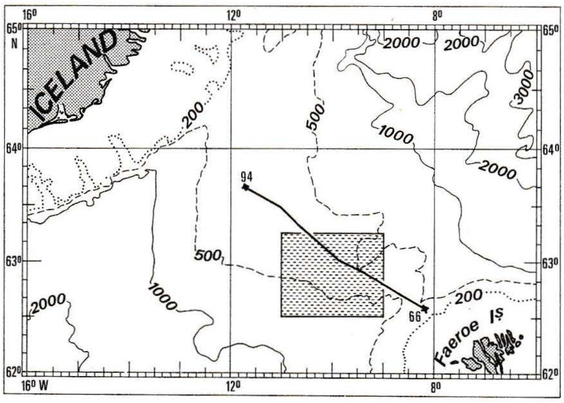


Fig. 1. Area of investigation. The line indicates the section parallel to the crest (Stat. 66—94) of the ridge, the area of small scale investigations has been shaded. Depths in m.

Das Untersuchungsgebiet mit der Lage des kammparallelen Schnitts (Stat. 66—94). Das Gebiet der kleinräumigen Untersuchungen ist schattiert. Tiefen in m.

between Iceland and the Faeroe Islands (fig. 1). The greatest depth of 485 m on the ridge is found in a channel across its southern part, the highest parts of the ridge rise up to 250 m below the surface.

The equipment used aboard the "Meteor" for the shipborne measurements consisted of the newly developed "Kieler Multimeeressonde" (Kroebel 1973), a profiling current meter and expendable bathythermographs (henceforth XBT). The parameters measured by the Multimeeressonde (henceforth MMS) are temperature, salinity, pressure, light attenuation and some other properties not referred to here. This instrument proved in fact to be a very successfull tool which operated without any failure during the expedition. The errors according to Kroebel (1973) are $\pm$ 0.01 °C for temperature, $\pm$ 0,01‰ for salinity, $\pm$ 2.5‰ of the full range for pressure and $\pm$ 0.1% of full range for light attenuation. It turned out that especially the light attenuation data were of high value for the tracing of the overflow.

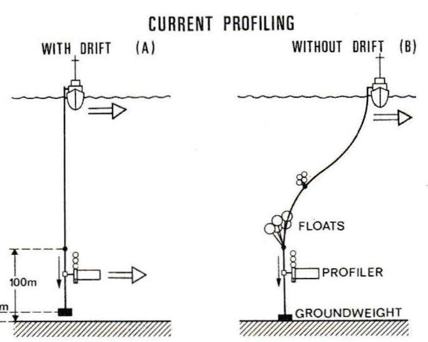


Fig. 2. Methods A (left) and B (right) of taking profiles with the PCM.

Methoden A (links) und B (rechts) zur Profilmessung vom driftenden Schiff.

The profiling current meter (henceforth PCM) was simply a Bergen type meter switched to continous recording which means a recording cycle every 32 seconds. It measures pressure, horizontal current speed and direction and temperature with errors of $\pm$ 4 dbar, $\pm$ 2 cm/sec, $\pm$ 5° and $\pm$ 0.1 °C, respectively. Electrical conductivity is also measured. Small negative buoyancy of the instrument was achieved by attached floats similar to the procedure used by Düing & Johnson (1972). To measure the near bottom current profile, two different methods, A and B in fig. 2, were tried. Both methods start with the same procedure. The hydrographic wire, with groundweight, PCM, stopper at the wire (and floats attached in method B) is lowered with a velocity exceeding the free fall velocity of the PCM. Therefore the PCM is caught by the stopper 100 m above the groundweight. In method A the wire is stopped, when the groundweight arrives at a depth about 3 m above the bottom. The PCM then sinks freely (about 10 m/min) while horizontally drifting with the ship.

In order to calculate absolute currents from the measurements, a drift correction is necessary. This is difficult to achieve because position finding by satellite navigation is not frequent enough within the duration (10 to 15 minutes) of the current meter descent, and Loran C was too inaccurate at least during part of the time. Radar buoys were moored, but the drift velocities calculated from distance measurements to these buoys also carried errors too large to be acceptable. In the case of strong currents (above 30 to 40 cm/sec), however, gross features of speed and direction can be obtained by method A.

Method B avoids the difficulties caused by the ship's drift. The groundweight is put onto the bottom. While the PCM is sinking, more wire has to be paid out from the drifting ship, with the floats keeping the wire vertically upwards to 100 m above the bottom and about neutral above the larger main floats. In this way a mooring is established during the time of the measurement which takes 15 minutes.

2 The Data

During the first leg four MMS sections were carried out, two across the ridge and two along it's crest. The temperature distribution in the first section along the crest is presented in fig. 3. It shows that bottom temperatures less than 1 °C only occured at the deepest part of the section in the channel mentioned above. In order to investigate the southward extension of this cold water mass "Meteor" took profiles on the stations shown in fig. 4, working from northeast to southwest with a time lag of 86 hours between the first station (Stat. 96) and the last station (Stat. 158). Wherever cold bottom water ($T \leqslant 2$ °C) was found, it was separated from the

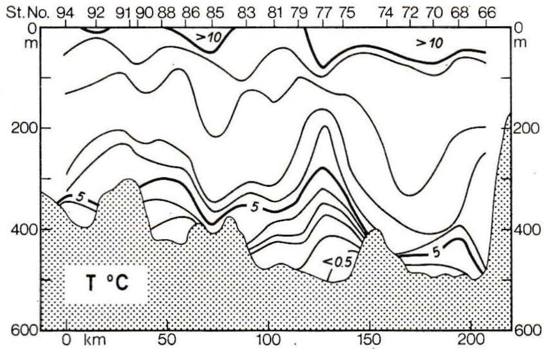


Fig. 3. Temperature section parallel to the crest (18 to 20 Aug. 73). The position of the section is given in fig. 1.

Temperatur-Schnitt (18.—20. Aug. 73) parallel zum Kamm des Rückens. Lage des Schnitts siehe Abb. 1.

warmer Atlantic water (T ⩾ 5 °C) by sharp gradients in temperature, salinity, attenuation and, in most cases, current speed and direction. As an example, the profiles at Stat. 154 are given in fig. 5. Sharp horizontal gradients in T, S and A are indicated by the maps of bottom distances of special values of these quantities (fig. 6b). The pattern of the vertically averaged currents in the cold bottom layer (T ⩽ 2 °C) is given in fig. 7.

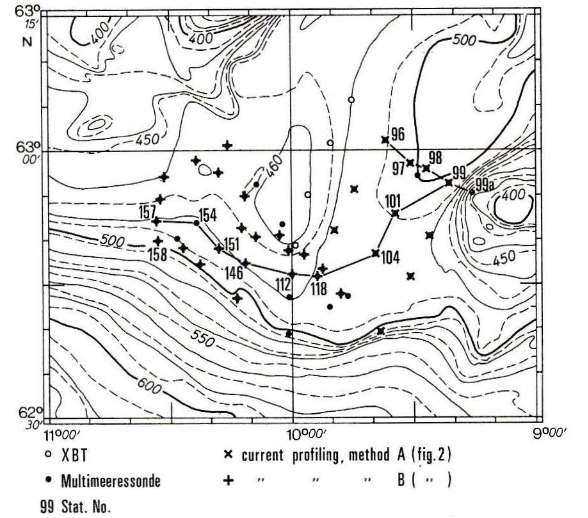


Fig. 4. Area of small scale investigations (shaded in fig. 1) with depth contours and the position of σ_t-sections (shown in fig. 8). Types of measurements are indicated by the station symbols. Depths in m.

Das kleinräumige Untersuchungsgebiet (schattiert in Abb. 1) mit Tiefenlinien und Lage der σ_t-Schnitte in Abb. 8. Die Art der Messungen ist durch die Stationssymbole gekennzeichnet. Tiefe in m.

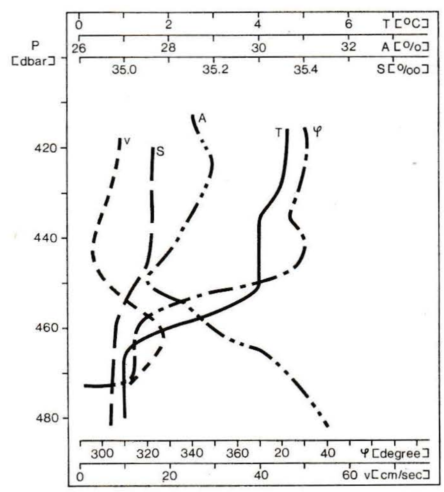


Fig. 5. Near bottom profiles of temperature T, salinity S, light attenuation A, current speed V and direction φ at Station 154.

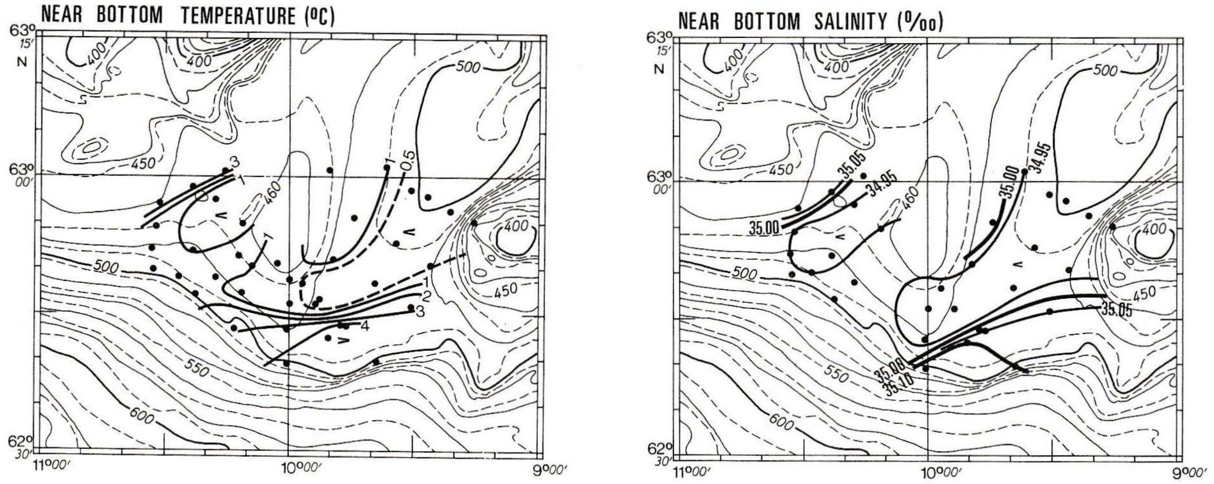
Bodennahe Profile von Temperatur T, Salzgehalt S, Attenuation A, Strömungsgeschwindigkeit V und -richtung φ auf Station 154.

3 Discussion

As already mentioned before, the overflow at the Iceland Faeroe Ridge appears to be highly variable. Further support for this statement is provided by several sections (not presented here) which were obtained close to a line along stations 96—99a (fig. 4) during the second leg of the cruise[1]. According to these data the maximum height of the 1 °C isotherm above the bottom varies from 100 m to 30 m and it's location varies from the eastern to the western part of the channel. In the case of a quasistationary outflow one would expect the isopycnals to slope down towards the east in the channel. In fig. 8a, however, the σ_t-section from station 96 to 99a shows the maximum height of the isopycnal $\sigma_t = 28.0$ between stations 98 and 99 just east of the channel center. This isopycnal corresponds to the interface between the Atlantic water and the bottom water.

The flow pattern in fig. 7 indicates an unsteadiness in the overflow. Northward flow with speeds of 30 to 40 cm/sec was observed in the sill region (St. 97—98, 102 and 103). Perhaps there was some contribution to this backflow on stations 144 and 145 where the water depth is some meters less than the sill depth, but the measured velocities are within the noise range. South of these stations outflow with speeds of approximately 20 cm/sec and westward direction occured. One should expect tidal effects to be mainly responsible for the variability of the current field in this region. This is supported by observations of Steele (1967) who tracked parachute drogues in the area north of our stations 104 and 144. He reports tidal (or inertial) oscillations. However in the western part of our area of observations (St. 146—158), the flow pattern in fig. 7 looks rather regular although the time lag between St. 146 and 158 covers 2.2 semidiurnal tidal cycles and the stations in between

[1] W. Horn and J. Meincke, personal communication.



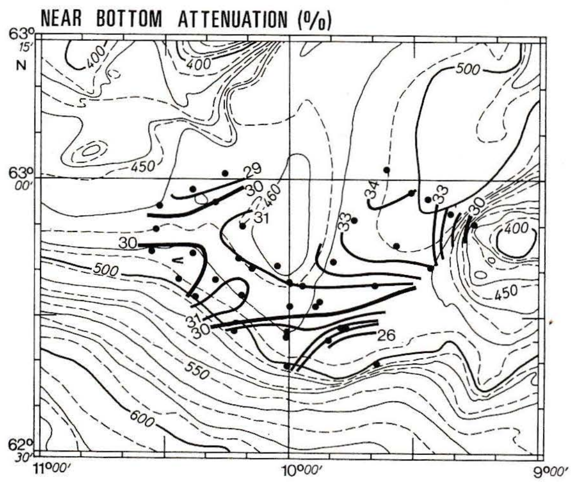


Fig. 6a. Near bottom values of temperature, salinity and light attenuation in the area of small scale investigations. — Depths and bottom distances in m.

Temperatur, Salzgehalt und Licht-Attenuation in Bodennähe. Tiefen und Bodenabstände in m.

were distributed uniformly over this time interval. Thus the flow field in this part of the area with velocities of about 20 cm/sec does not seem to be influenced very much by tidal disturbances.

As a possible interpretation of the flow pattern in fig. 7 one may expect that an internal disturbance propagates from the Norwegian Sea towards the ridge, piles up at the crest and results in a cold water spill-over through the channel. While the disturbance decreases on the side of the Norwegian Sea and the cold water is flowing back, the overflow water southwest of the sill depth becomes independent of the initial perturbation. This isolated water mass with larger density than that of the environment will flow almost parallel to the isobaths driven by the slope-parallel component of gravity and influenced by friction and Coriolis force. In a similiar way such a highly unsteady single overflow event was generated and observed qualitatively in a rotating tank by one of the authors[2]. Instead of one single perturbation, several consecutive perturbations with a time scale of a few hours (e. g. tidal oscillations) could set on an overflow with a current field similar to that in fig. 7 for the time of generation, i. e. alternating currents in the sill region with the time scale of the perturbation and approximately steady flow south and west of the sill depth during this time.

The maps in fig. 6a and b also indicate that the cold and low salinity water in this region on the southwest side of the ridge originates from the northern part of the channel. By both the near bottom temperatures and salinities (fig. 6a) the overflow can be traced flowing to the right and following the depth contours after having passed the sill. To the west a second, possibly earlier event of overflow seems to be indicated. The near bottom attenuation (fig. 6a) has its maximum in the northern part of this area. It is not necessarily correlated with the current speed since it also depends upon the type of the bottom material. Towards the south bottom attenuation values are already almost as low as in the intermediate minima in fig. 5. The pattern of bottom distances (fig. 6b) of the 1 °C level-decreasing from more than 75 m to zero — as the pattern of near bottom

[2] T. J. MÜLLER, unpublished notes.

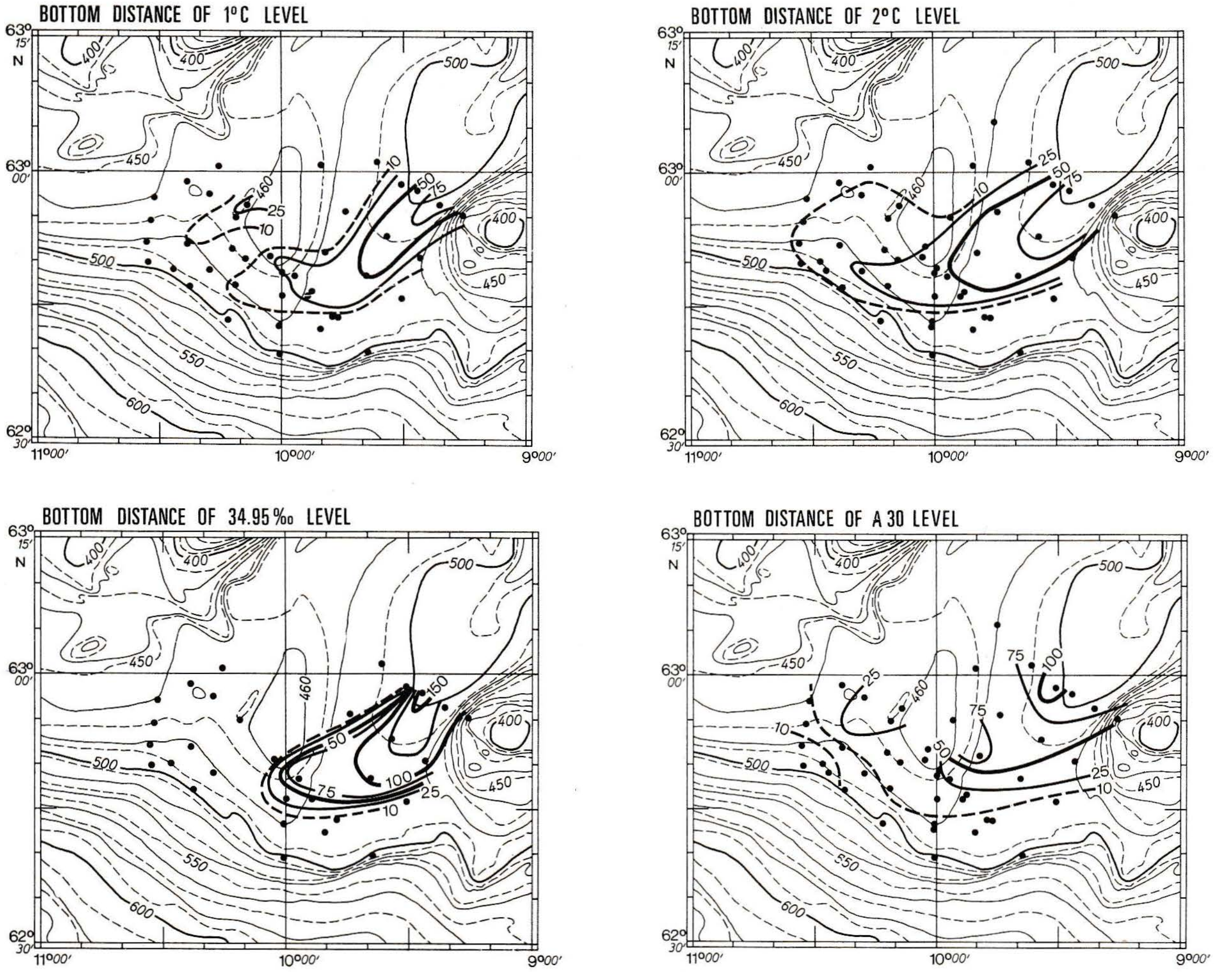


Fig. 6b. Bottom distances of $T = 1$ °C, $S = 34.95$ ‰, $A = 30$ % and $T = 2$ °C levels.

Bodenabstände der $T = 1$ °C, $S = 34{,}95$ ‰, $A = 30$ % und $T = 2$ °C Flächen.

temperatures and near bottom salinities in fig. 6a gives evidence for an overflow event and a second lower temperature water body in the west.

The bottom distance of the 34.95 % level decreases from more than 150 m to zero in the westward extension of the tongue. The water mass in the west cannot be distinguished from that in the east by the salinities. The same features emerge from the bottom distance map of the 2 °C level (fig. 6b) and (not shown here) from both the maps of the 35‰ bottom distances and of the thickness of the homogeneous bottom layer.

The increase of near bottom temperatures and salinities with increasing distance from the "source" of overflow water at Stat. 96—99a indicates entrainment of Atlantic water into the overflow water body. On the other hand the bottom charts show that there is almost no increase of the width of the overflow layer. Therefore the decrease in the bottom distances of certain temperature and salinity levels corresponds to a flux of mixed overflow water into the lighter

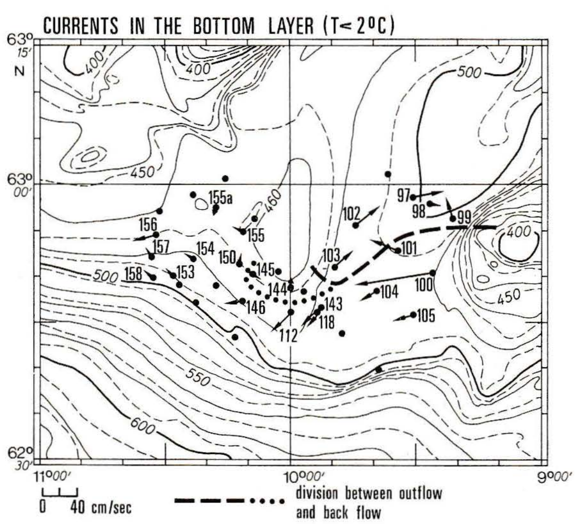


Fig. 7. Currents in the bottom layer with $T \leqslant 2$ °C. Depths in m.

Strömungen in der Bodenschicht mit $T \leqslant 2$ °C. Tiefen in m.

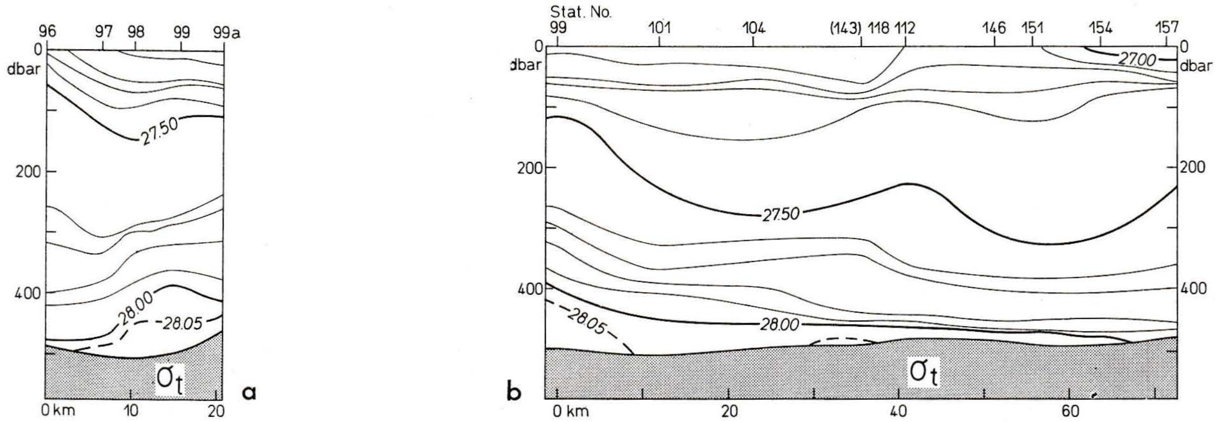


Fig. 8. a) σ_t-section from Stat. 96 to 99a. b) σ_t-section along the axis of the overflow from Stat. 99 to 157. The positions of stations are given in fig. 4.

a) σ_t-Schnitt von Stat. 96 bis 99a. b) σ_t-Schnitt entlang des Overflowwasserkörpers von Stat. 99 bis 157. Lage der Stationen s. Abb. 4.

stably stratified water above which should result in an increase of potential energy[3]. Indeed the σ_t-section in fig. 8b does not show this effect, but it may simply be lost due to the small scale of the event compared with the water masses above.

The vertical profiles (fig. 5) show sharp gradients, but there is no evidence for an Ekman spiral at the interface since both right and left turning of the flow in this gradient layer was observed at several stations. Yet all gradients are finite and thus models with penetrative mixing as used by Pollard, Rhines & Thompson (1973) for the wind mixed surface layer, Thompson (1973) for stratified Ekman boundary layers, or Kato & Phillips (1966) in their tank experiments cannot be used as an approach to the problem. Moore & Long (1971) carried out tank experiments with a constant gradient layer separating two almost homogeneous layers with opposite flow which resembles the condition in our case more appropriately. They computed Richardson-numbers Ri_L based on a vertical scale L within the gradient layer and found Ri_L to be of order one. We have calculated such Richardson-numbers

$$Ri_L = g\,\Delta\,\rho\,L/(\rho_B\,|\Delta\,u|^2)$$

from our PCM measurements within the gradient layer above the bottom layer. Here Δu is the vector difference of velocity and $\Delta\rho$ the density difference in a thin layer of thickness L (5 to 10 m) where the sharp and approximately constant gradients in temperature and density occur; g is gravity and ρ_B is the density in the near bottom layer. The results for 13 Stations (where method B was used for current profiling) are listed in table 1. The relatively high values of Ri_L at Stat. 143 and 158 are due to the fact that the depth of the maximum velocity shear on these station does not exactly coincide with the reference depth, defined by the maximum temperature gradient. In general the numbers Ri_L do not differ very much from 1, thus indicating marginally stable flow. One therefore is encouraged to use $Ri_L = 1$ for further theoretical investigations of such flows with mixing across continous gradient layers.

4 Conclusions

The observations in an overflow event showed that near bottom temperatures and salinities increased with increasing distance from the source. This is due to the entrainment of Atlantic water into the overflow water. At the same time the decrease of the cross section area of the cold water with increasing distance from the source indicates upward flux of mixed overflow water. Overflow theories will have to take into account both these facts as well as vertically continous density gradients separating a vertically homogeneous layer below from a weakly stratified layer above.

Table 1 Richardson-number

$$Ri_L = g\,\Delta\,\varrho\,L/(\varrho_B\,|\,\Delta\,u\,|^2)$$

based on a vertical scale L within the gradient layer. g, ϱ_B, $\Delta\varrho$ and $\Delta\,u$ are gravity, density in the near bottom layer, density difference and current vector difference across L, respectively. L is approximately 5 to 10 m. The distribution of this small number of Ri_L values is not Gaussian.

St. No.	112	118	143	144	145	146	150
Ri_L	0.7	0.3	5.3	1.5	0.6	1	1
St. No.	153	154	155	156	157	158	
Ri_L	1	1.7	0.5	1.2	1.8	7	

[3] J. S. Turner in a personal communication pointed out that the thickness of a plume or a thermal can only be judged from profiles, not from the level of particular isotherms (see also Turner 1973, p. 169). But note that in our case such levels correspond to the thickness of the bottom layer because of the sharp vertical gradients.

Acknowledgement

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