Temperature and Salinity Structure in the Weddell Sea

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ABSTRACT

The general circulation of water in the Weddell Sea is part of a large cyclonic gyre. A section taken across this gyre from the Scotia Ridge to Cape Norvegia shows that the Warm Deep Water forms an asymmetric lens-like structure with the thickest portion south of the center of the sea. This large-scale feature of the Weddell Sea is evidently due to a divergent Ekman flux driven by the general atmospheric circulation and upwelling in the center of the gyre. Vertical profiles of temperature and salinity in the center of the gyre show small step-like structures in the upper part of the transition from colder, less salty Winter Water to the warmer, saltier Warm Deep Water below and large step-like structures in the lower part of the transition region. Double-diffusive convection can take place in both regions. Circumstantial evidence leads one to believe that the cabbeling instability is effective in the large-step region. Internal waves and shear instabilities may also be mechanisms that contribute to the formation of the step-like structures.

1. Introduction

The Weddell Sea is an area of the ocean which is of great interest to the physical oceanographer as it is probably the major source region for Antarctic Bottom Water (Deacon, 1937). It is also a region with a very interesting temperature and salinity structure.

The gross features of the Weddell Sea were first described by Brennecke (1921) using the data he obtained during the cruise of the *Deutschland* in 1911–12. Subsequent observers (Deacon, 1933, 1937; Mosby, 1934; Elder and Seabrooke, 1970) have confirmed these features. Below a seasonally variable surface layer, which extends to depths between about 20 and 50 m, three water masses make up the great bulk of the water in the oceanic portion of the Weddell Sea: Winter Water characterized by potential temperatures $\theta=-1.8$ to -1.6°C and salinities $S=34.36$ to 34.52‰, Warm Deep Water $\theta=0$ to 0.8°C and $S=34.64$ to 34.72‰, and Antarctic Bottom Water $\theta=-0.8$ to 0°C and $S=34.64$ to 34.68‰. The Warm Deep Water lies between the Winter Water and Antarctic Bottom Water in an asymmetric lens-like structure with the thickest portion lying south of the center of the sea. The drift of the *Deutschland*, while she was beset in the ice pack, showed that the surface flow was clockwise around the sea. The general circulation has since been shown by Carmack and Foster (1975a, b) to consist of a fairly narrow current to the west flowing along the Antarctic coast in the south and a broad outflow to the east in the north just south of the Scotia Ridge. In the northern part, at least, the flow is largely barotropic with velocities varying only a few centimeters per second from top to bottom. The total flow out of the Weddell Sea is estimated to be about 97×10^6 m^3 s^{-1}. The circulation within the Weddell Sea is part of a large gyre, which extends over the Atlantic Antarctic Basin possibly as far east as about 40°E (Treshnikov, 1964). The oceanic circulation is probably the result of the cyclonic circulation of the atmosphere around a nearly permanent, elongated low with its center at about 60 to 65°S between 10 and 20°E and with a secondary minimum sometimes at about 65°S between 20 and 30°W (Taljaard *et al.*, 1969).

With the advent of continuous profiling temperature and salinity sensors (STD's), layered structures were found at numerous places in the ocean (e.g., Stommel and Federov, 1967; Tait and Howe, 1968; Neal *et al.*, 1969). Similar step-like temperature and salinity structures were observed by the *General San Martin* in the transition between the Winter Water and Warm Deep Water layers in the central Weddell Sea (see Foster, 1972). During January and February 1973 as part of the International Weddell Sea Oceanographic Expedition (IWSOE) a hydrographic section across the Weddell Sea from the Scotia Ridge to Cape Norvegia was made by Scripps Institution of Oceanography aboard the USCGC *Glacier* (Fig. 1). Continuous vertical profiles of temperature and salinity were obtained using a Plessey Model 9040 STD. At each station the STD was lowered at 20–30 m min^{-1} over the shelf and in the upper part of the

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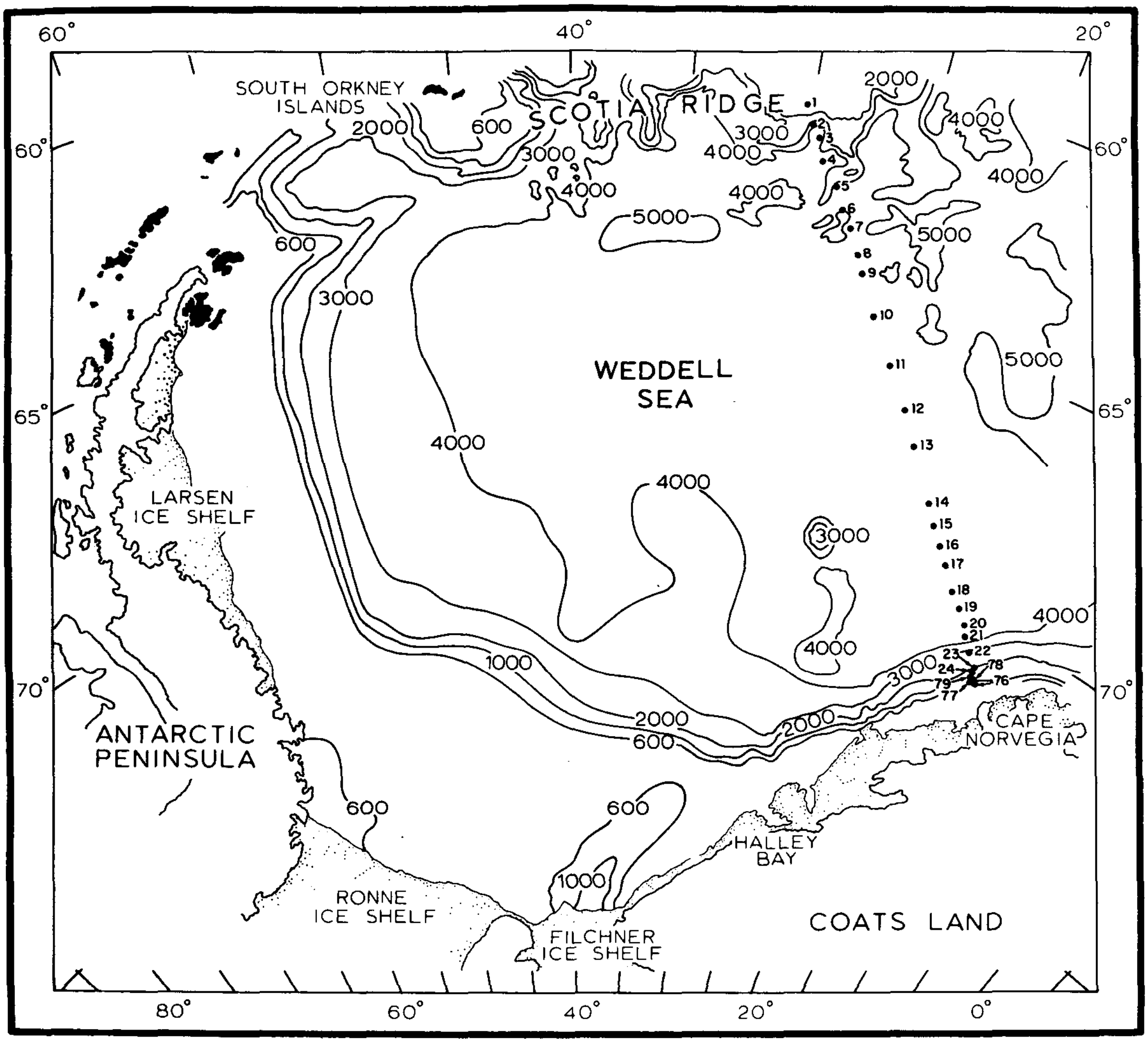


FIG. 1. Bathymetry of the Weddell Sea and location of stations taken by the USCGC *Glacier* during the International Weddell Sea Oceanographic Expedition in 1973 along the section from the Scotia Ridge to Cape Norvegia.

ocean where rapid variations of temperature and salinity occurred in order to minimize salinity errors due to the different response times of the temperature and conductivity sensors. Below the core of the Warm Deep Water in the open ocean (200–600 m) variations were slow and the STD was lowered at faster speeds up to a maximum of 60 m min^{-1}. The STD was standardized against special low-range (-3 to 2°C) reversing thermometers, water salinities were measured on a Hytech Model 6230 inductive salinometer, and depths were measured with the ship's echo sounder and the pinger. The salinity data were corrected for depth using a second-order polynomial fitted to the calibration data which resulted in a standard deviation of $\pm 0.005‰$. The temperature data agreed with the calibration thermometers with a standard deviation of ± 0.006°C. The pressure data from the STD were corrected for density variation with depth and fitted to a second-order polynomial. These depths agreed well within the experimental accuracy of the acoustic data. The data were recorded as analog traces and on some stations also digitally on magnetic tape. Discounting obvious salinity spikes the data recorded by both methods compared well with each other. The analog records were digitized by hand with the aid of an Oscar Model S-2 digitizer. The procedure followed was to digitize the record at every major change of slope, omitting the spikes in the salinity trace. This eliminated both instrument noise and features with a scale less than about 1 m. Thus the final records appear very clean. We believe that, although the microstructure has been smoothed by the digitization process, the small-scale features are preserved. The slow lowering speeds used and the low level of ship motion contributed to the success of the STD in recording these features.

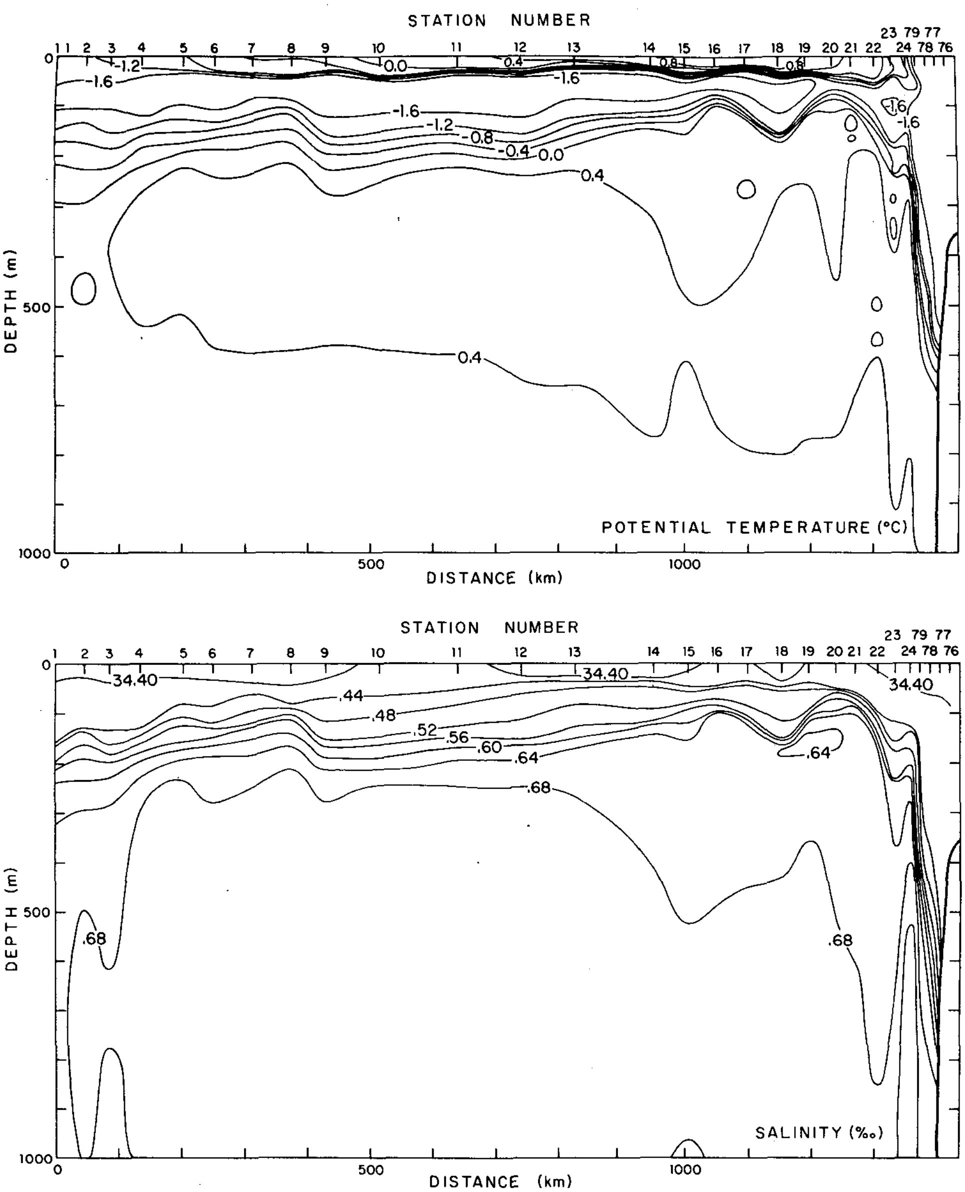
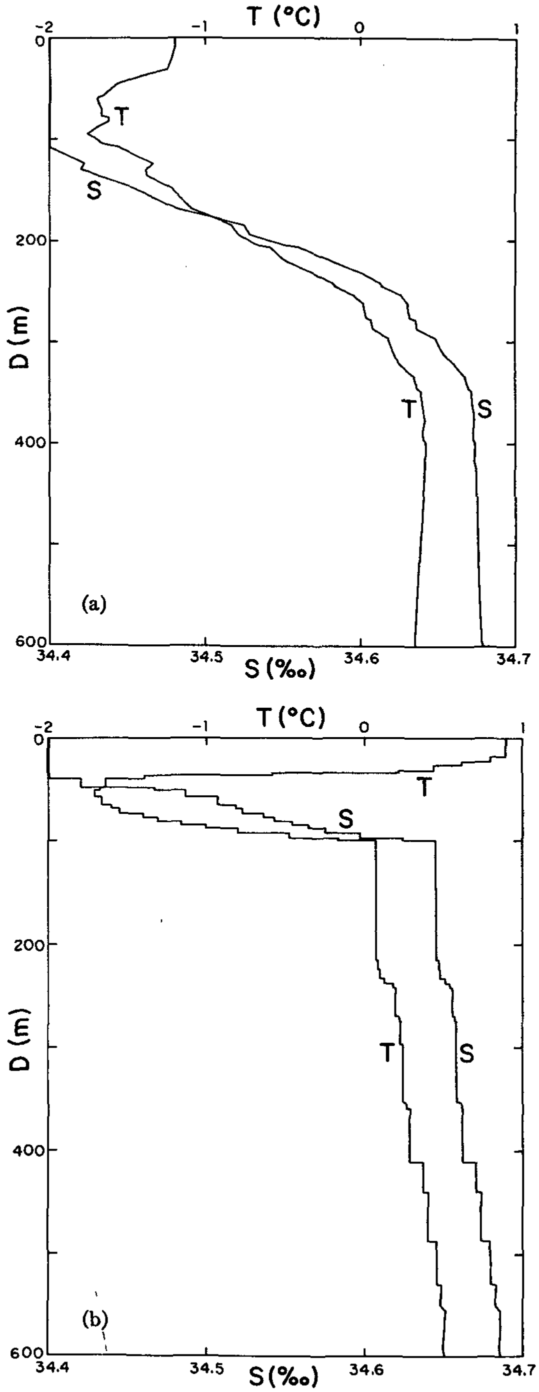


FIG. 2. Hydrographic sections of potential temperature (a) and salinity (b) from the Scotia Ridge to Cape Norvegia.

2. Large-scale features

Figs. 2a and 2b show the potential temperature and salinity for the top 1000 m of the section from the Scotia Ridge to Cape Norvegia. The high temperatures and low salinities in the top 20–50 m reflect summer solar heating and sea ice melting. The 0°C potential temperature and the 34.6‰ salinity isopleths clearly show the dome-like structure of the Warm Deep Water with its main vertex at Station 16 and a secondary vertex at Station 20. The asymmetry of the structure probably reflects the narrow inflow in the south and the broad outflow in the northern part of the Weddell Sea gyre. Thus the coldest surface temperatures are found at the southern end of the section (Stations 24, 79, 78, 77 and 76 were taken through leads and polynyas in the ice pack) and at the northern end (Station 1 was taken in open pack ice), and the warmest from Stations 15 to 17 (in open water).

The general structure of our section appears to be in good agreement with Deacon's (1933) interpretation that there is a divergence region between the currents flowing west and east in the central Weddell Sea and that here the subsurface waters upwell and tend to flow outward in, or just below, the surface layer. Our section is nearly orthogonal to the general flow, westward in the south and eastward in the north (Carmack and Foster, 1975a, b). Due to the low atmospheric pressure generally found over the east central Weddell Sea, the wind stress changes sign along our section. The wind speed is on the order of about 5 m s^{-1} at the ends of the section (Jenne *et al.*, 1971). The wind stress relation for low wind speeds (Sverdrup *et al.*, 1942) then gives a value of about 0.3 dyn cm^{-2}. The transport orthogonal to the wind stress gradient along our section can be very crudely estimated by Sverdrup's (1947) method by assuming



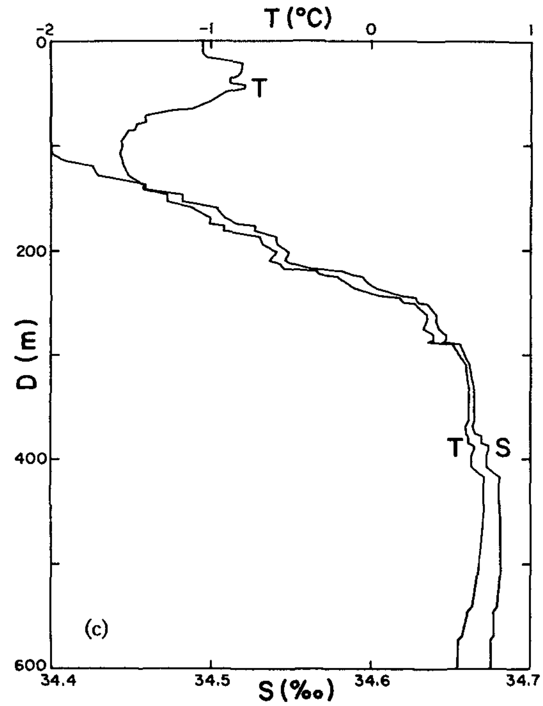


FIG. 3. Vertical profiles from processed STD records of temperature and salinity for Station 3 near the northern end of the section (a), Station 16 at about the turning point of the gyre (b), and Station 24 near the southern end of the section (c).

that the wind stress is zero at the center and increases linearly toward the edges. This gives a flux of 6 m^3 s^{-1} m^{-1} along the section, and if we integrate this flux around the roughly semi-circular periphery of the sea, the total divergent flux becomes 10×10^6 m^3 s^{-1}. Uniform upwelling over the sea to provide this flux would have vertical velocities on the order of 10^{-3} cm s^{-1}. The general structure of our section agrees with this simple model of divergent Ekman flux and upwelling. The details of the thermohaline structure, however, show that other processes may be operative; some of these processes are discussed below.

3. Small-scale features

The individual STD profiles of temperature and salinity show considerable variation in small-scale features along the section. The upper 600 m of the profiles for Station 3 near the northern end of the section (Fig. 3a), Station 16 at about the turning point of the gyre (Carmack and Foster, 1975b) (Fig. 3b), and Station 24 near the southern end of the section (Fig. 3c) are typical of profiles for stations in these regions. The similarity of the two profiles near the ends of the section is striking, but the profile at the turning point is quite different; especially to be noted is the sudden change in the general trend of the gradients near 100 m depth at Station 16. While step-like structures are seen in all the profiles they

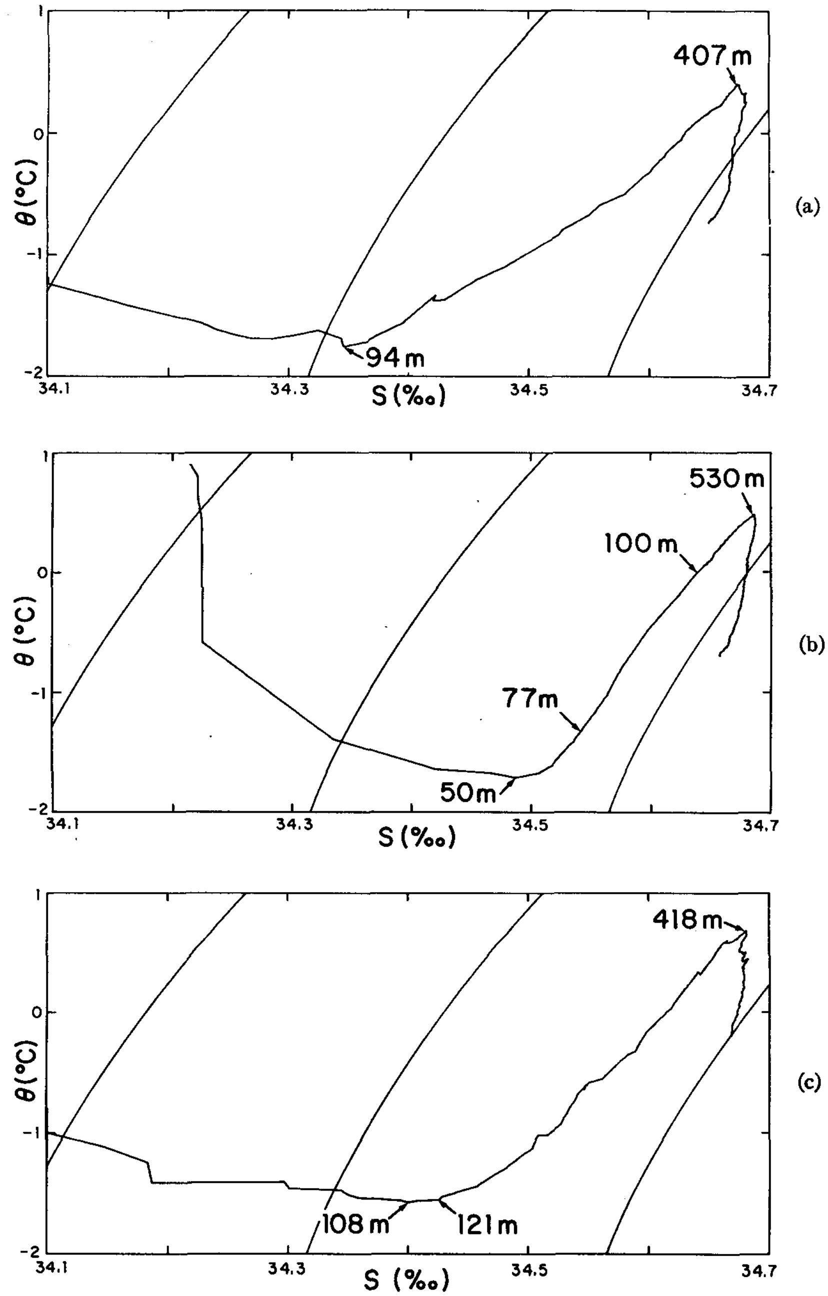


FIG. 4. Potential temperature-salinity (θ-S) diagrams for the same stations as in Figs. 3a–c. Lines of constant density at a pressure of 300 decibars are shown.

are much more prevalent at the stations near the turning point. Here there also seem to be two distinct types of steps: small changes in depth and large changes in temperature and salinity in the upper layers and generally large changes in depth and small changes in temperature and salinity in the lower layers. Station 24 has some small inversions, and for Stations 79 and 78 further to the south these become more marked probably as a result of interleaving of shelf water with the waters offshore.

At Stations 16 and 17 we obtained four separate STD profiles of the upper waters (two down and up traces at each station). The changes between profiles an hour or two apart were considerable, indicating that the large-step structures either had limited horizontal extent (the ship probably drifted a kilometer or two between traces) or were short-lived. Internal waves no doubt contributed to the variability of the large steps as they did for the small steps investigated by Neshyba *et al.* (1972) in the Arctic. The small-step structures found in the upper layers of the central Weddell Sea seem very similar to those found in the Arctic, but the large steps are apparently quite different (Neshyba, personal communication).

Figs. 4a–c show the potential temperature-salinity (θ-S) diagrams for the same three stations used in Figs. 3a–c. With the exception of the warm surface water found at Station 16 the main difference between this station and Stations 3 and 24 is the slope of the θ-S curves in the transition region between Winter Water and Warm Deep Water. Station 16 has a much steeper slope than Stations 3 or 24 and in the region of the large steps is almost parallel to the lines of constant density.

4. Mechanisms

a. Double-diffusive convection

The idea that the turbulent diffusion of salt in the ocean should be slower than that for heat is an old idea in oceanography and was used by Drygalski (1926) to explain the formation of Antarctic Bottom Water. More recently, a mechanism, double-diffusive convection (sometimes called the "diffusive" or "anti-salt-fingering" instability), has been proposed (Turner and Stommel, 1964) which can account for the different diffusion rates when cold, less salty water overlies warmer, saltier water. Since this is the situation in the central Weddell Sea where Winter Water overlies Warm Deep Water, the diffusive instability probably plays some role in the mixing that takes place between the two water masses.

We have calculated the density ratio $R_\rho=\beta\Delta S/\alpha\Delta T$ (where α is the coefficient of thermal expansion, β the coefficient of expansion for salinity changes, and ΔT and ΔS are the differences in temperature and salinity between two adjacent layers) using Ekman's (1908) equation of state at the *in situ* pressure at each interface between layers for Station 16. For the steps from 77 to 100 m depth the average value of the density ratio was 1.39, and for steps from 100 to 530 m the average was 1.03. Huppert (1971) has shown that a series of double-diffusive layers tends to remain in equilibrium if the density ratio is greater than 2. Since for both regimes of layers the density ratio is less than 2, it would appear that the steps at Station 16 are unstable and should break down in time. Perhaps the large changes observed in repeated STD profiles at the same station are due to this instability. The density ratio in the Arctic is on the order of 7 (Neshyba *et al.*, 1971), and this may explain the great permanence found in the Arctic step structure compared to that for the Weddell Sea steps.

b. Cabbeling instability

Witte (1902) was probably the first to point out that the nonlinearity of the equation of state of sea water could lead to vertical motion in the ocean driven by the density increase resulting from mixing water parcels with different temperature and salinity but of the same density. Fofonoff (1956) has shown that the formation of Antarctic Bottom Water is probably influenced by this process. Foster (1972) has analyzed the stability of superposed layers of sea water considering the nonlinearity of the equation of state and has demonstrated that a convective instability can take place. The cabbeling instability will occur whenever the mixture of two adjacent layers is denser than the lower layer and will result in convective flow in the lower layer that gradually erodes the layer above. The convection will continue until the properties of the two layers are such that mixtures are no longer denser than the lower layer.

The stability of any two layers to the cabbeling instability can be examined by means of a temperature-salinity plot. The line of constant density calculated at the depth of the interface and passing through the temperature and salinity of the lower layer is plotted; if the point for the upper layer lies to the higher density side of the tangent to this line at the point for the lower layer, but to the lower density side of the line itself, the cabbeling instability will take place. In the transition region between Winter Water and Warm Deep Water we have a weakly stable density profile with colder, fresher water overlying warmer, saltier water. In winter the salinity of the Winter Water will be increased by salt rejection during sea ice formation while its temperature will remain nearly constant since it is already nearly at the freezing point. Thus the proper way to examine the stability of the layers would seem to be to calculate what the salinity of the upper layer would need to be to just initiate the cabbeling instability. We wrote a computer program that does this by calculating the tangent to the isopycnal through the temperature and salinity of the lower layer at the pressure of the interface and finds the salinity at the intersection of the tangent with the temperature of the upper layer. If the salinity of the upper layer is greater than this salinity, the interface is unstable. For Station 16 the layers between 77 and 100 m were slightly stable (an average increase in the salinity of the top layer at each interface by 0.003‰ would be required for instability) whereas almost all the layers between 100 and 530 m had nearly neutral stability at each interface with an average salinity of 0.001‰ greater than would be required for instability. Thus it seems very likely that cabbeling plays a role in the convection in the large-step region (100–530 m depth). In the small-step region (77–100 m depth) cabbeling instability was not operative at the time of our observations; however, if the salinity of the Winter Water were increased by haline convection induced by sea ice formation in winter by about 0.024‰, then

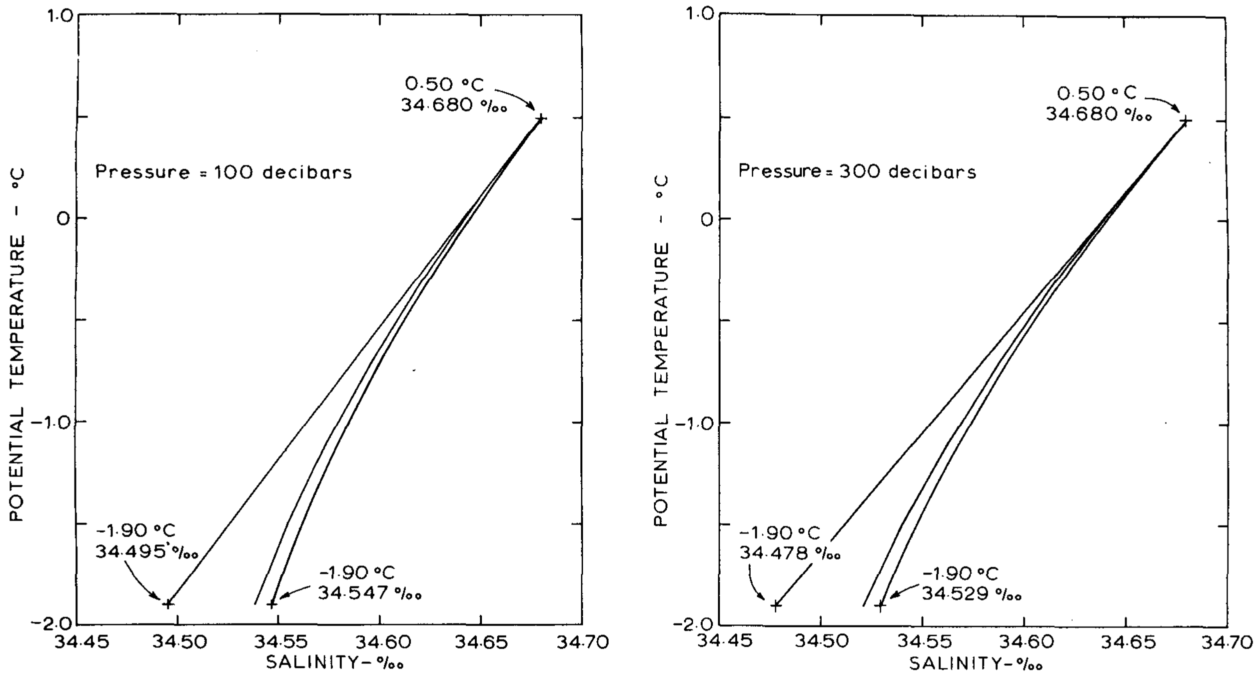


FIG. 5. Temperature-salinity (*T-S*) diagrams at pressures of 100 (a) and 300 db (b).

the small steps would be unstable (assuming the number of steps does not change).

The degree to which the cabbeling instability can change the slope of the mixing curve on a *T-S* diagram depends not only on the pressure but also the number of separate interfaces in a *T-S* interval. Figs. 5a and 5b show these effects. We examine here the hypothetical mixing of Warm Deep Water (0.5°C, 34.68‰) with Winter Water (at the freezing point −1.9°C) above. The curved line to the right in each figure is the isopycnal passing through the Warm Deep Water *T-S* point. The straight line to the left is the tangent to the isopycnal. We see that for a single interface the salinity that the Winter Water must exceed for the cabbeling instability to be operative is 34.495‰ at a pressure of 100 db and 34.478‰ at 300 db. The segmented line to the right of the tangent represents the critical salinity for a series of six interfaces separated by temperature steps of 0.4°C each. The salinity difference for a series of layers rapidly approaches that for a constant density curve as the temperature difference across each interface decreases. It is interesting to note that at Stations 3 and 24 the break in the θ-*S* curves (Figs. 4a and 4c) where Winter Water starts to mix with Warm Deep Water occurs at a lower salinity than for Station 16 (Fig. 4b). Evidently this is due to the pressure effect shown in Fig. 5 as the depth of the Winter Water–Warm Deep Water transition is greater for Stations 3 and 24 than for Station 16 (Fig. 2).

The *Deutschland* data (Brennecke, 1921) were taken by bottle sampling at discrete depths so we cannot see what the small-scale vertical structure was like in the only winter data from the Weddell Sea. We can, however, examine the gross stability of the profiles. It was found that as winter progressed the salinity of the Winter Water showed a general increasing trend. Three stations in late winter (68, 69, 71) showed a possible cabbeling instability between the lowest sample in the Winter Water layer and the next deeper sample. Furthermore, the maximum salinity reached by the Winter Water in the *Deutschland* winter data or any more recent data is less than 34.52‰ which is the minimum salinity required for the cabbeling instability to be operative for extreme properties (bottom layer: 0.5°C, 34.7‰; top layer: −1.8°C, interface at 100 db).

c. *Finite-amplitude pressure instability*

Gill (1973) has shown that instabilities may arise when parcels of water are given finite vertical displacements due to the differences in compressibility at different temperatures. Thus, since the compressibility of cold water is generally greater than that of warmer water, if a parcel of cold water overlies warmer water and is given a large displacement downward, it might remain at the lower depth. We tested this idea by making virtual vertical displacements of water parcels downward and upward various distances and then comparing their densities at the new pressure level with the *in situ* densities. The temperature of the parcel was assumed to vary adiabatically. This variation was calculated using the Ekman equation of state and fitting a 27 term (selected from 108 terms) polynomial to the resulting adiabatic temperature changes; this polynomial was accurate to about 10^{-4} °C for the range of temperature, salinity and pressure in the Weddell Sea.

It was found that small displacements did not result in finite-amplitude instabilities except for layers that were based on suspect data (slightly unstable to infinitesimal displacements). Large displacements also did not result in instabilities except for the cold layer of water just below the temperature minimum (from about 50 to 75 m at Station 16) which, when moved to the bottom, proved to be denser than the bottom water. This cold layer of water is Winter Water that has mixed slightly with Warm Deep Water. The layer is evidently in a metastable state, but since it would require a displacement on the order of thousands of meters to become unstable, it seems doubtful that this would occur. The possibility of a finite-amplitude instability becomes even more unlikely when we consider that the central Weddell Sea is a region of upwelling. We therefore conclude that while pressure instability may contribute to the processes leading to the formation of step-like structure, it probably is not a primary process.

d. Internal waves and shear instabilities

Breaking internal waves have been proposed as one mechanism by which small-scale vertical structure may be generated (Orlanski and Bryan, 1969). Since we did not make any time-series measurements beyond a very few repeated STD profiles, we have no direct evidence concerning internal waves. The Brunt-Väisälä frequency was calculated for the general profile, and at Station 16 we found in the small-step region from 77 to 100 m that the frequency was 3.1×10^{-3} s^{-1}, and in the large-step region from 100 to 550 m that it was 5.6×10^{-4} s^{-1}. The very low Brunt-Väisälä frequency in the large-step region would suggest that internal wave breaking may be a factor that influences the vertical structure in this region. It should be noted that in the central Weddell Sea along our section the bottom topography is remarkably flat and deep (about 4500 m), and thus topographically induced internal waves should not be produced in this region.

Shear instabilities of the Kelvin-Helmholtz type (see Turner, 1973) may also influence the vertical structure especially in water of low stability, such as is found at Station 16 in the large-step region. Unfortunately, we do not have any direct measurements of shear. Calculations by Carmack and Foster (1975a, b) show that only in the southern part of our section (near Station 24) are the geostrophic velocities over a few centimeters per second. In fact, it is mainly in the southern part of the section where considerable small inversions of temperature and salinity were found (see Fig. 3c). However, it is to be noted that these inversions were nearly all density-compensated indicating that they probably formed by interleaving of the Warm Deep Water with the adjacent shelf water.

5. Discussion

On the basis of our rather crude measurements of temperature and salinity it does not seem possible to make a definitive evaluation of the various processes that may affect the development of the small-scale structure. The proper conditions for double-diffusive convection and the cabbeling instability to be operative were found (double-diffusive in the small-step regime, both in the large-step regime), but it was not possible to determine the relative importance of the two mechanisms. Circumstantial evidence, particularly with regards to the salinity maxima of Winter Water and its decrease with increasing depth, leads us to believe that the cabbeling instability is an effective mechanism for promoting vertical mixing. We cannot rule out the possibility that mechanical mixing induced by internal waves or shear is also effective. High-resolution temperature and salinity profiles, time series and velocity measurements would be of great assistance in sorting out the various mechanisms that may contribute to the formation of the temperature and salinity structure of the Weddell Sea.

The upwelling of Warm Deep Water in the center of the Weddell Sea gyre and subsequent mixing with Winter Water results in the formation of a new water mass, Modified Warm Deep Water. This water mass exists only as a transition between the two water masses except when it intrudes onto the continental shelf where it can be identified by its core properties (Carmack and Foster, 1975a). Since it is Modified Warm Deep Water and not Warm Deep Water itself that intrudes onto the shelf and mixes with shelf water to form bottom water (Foster and Carmack, 1975), the importance of understanding the mixing processes in the step-like structures is apparent. Since the low atmospheric pressure region is located just east of the center of our section all year around, the Ekman flux and upwelling should be nearly continuous. Thus we do not see how bottom water formation could take place in the central Weddell Sea.

The gross structure of the Arctic gyre bears a resemblance to that of the Weddell Sea gyre. There is a cold, low-salinity surface layer, a warmer, saltier intermediate layer (Atlantic Water), and a gradual decrease in temperature to the bottom. However, the resemblance is largely superficial. In the Arctic the great influx of fresh water from rivers causes the surface water to be very much less saline than Atlantic Water. Thus the salinity difference from surface to intermediate water varies from about 4 to 6‰ in the Arctic while in the Weddell Sea, where the main influx of fresh water is glacial ice, the salinity difference is only about 0.5 to 1‰. The Arctic and Antarctic regions are antithetical to each other, the Arctic being ocean and the Antarctic being land. Thus the Arctic ice is largely trapped while in the Weddell Sea the ice floats out to the north before it can all melt and mix fresh water down into the surface layer. Thus the step-structure in the Arctic is in a strong halocline, which prevents the cabbeling instability from taking place. Furthermore, the Arctic gyre is anticyclonic with the isopycnals bending down in the middle (Treshnikov and Baranov, 1972). Thus, there is neither an Ekman layer pumping surface water to the edges nor any general upwelling. Finally, we should note that bottom water does not seem to be formed in the Arctic ocean but flows in through the Fram Strait from the Greenland Sea.

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