Deep Antarctic Convection West of Maud Rise[1]

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(Manuscript received 17 August 1977, in final form 28 December 1977)

ABSTRACT

In February 1977 a column of water (14 km radius), within the central region of the Weddell Gyre west of Maud Rise, was observed in which the normal Antarctic stratification sequence of temperature-minimum to temperature-maximum was absent. The column appeared as a cold, low-salinity, high-oxygen, cyclonic flowing (surface velocity above 50 cm s^{-1}) eddy extending to at least 4000 m. It is hypothesized that similar eddies were common in this region (at least in Austral summer 1977) and represent winter structures which have survived into the summer period. Eddy formation is explained as a product of winter period static instability, similar to the MEDOC observations in the Mediterranean, but without the subsequent sinking and spreading phase. Winter period static instability in the Weddell Gyre is shown to be a likely condition and may be related to the frequent occurrence of a large polynya within the central region of the Weddell Gyre. Deep penetration of winter surface water within the eddy supplies the characteristics of a deep, low-salinity, high-oxygen intrusion near sigma-2 stratum 37.21 to 37.23 (between 1500 and 2000 m). This intrusion may represent a distinct water type formed within the Weddell Gyre. It would represent a variety of Antarctic Bottom Water [or Antarctic Deep Water (Wüst 1933)] with an open ocean origin. It may spread by isopycnal processes north of the Argentine Basin over the Rio Grande Ridge. The continental-margin-produced Antarctic Bottom Water may be partly topographically confined within the Weddell-Argentine-Crozet Basin trio.

1. Introduction

Formation of Antarctic Bottom Water is generally considered to result from processes at specific sites of Antarctica's continental margins. The site which is apparently responsible for the coldest variety of bottom water is the Weddell Sea (Gordon, 1974; Foster and Carmack, 1976a). Formation of abyssal waters of the Labrador, Norwegian, Greenland and Mediterranean Seas are believed to be an open ocean process, with deep convection primarily associated with the centers of the geostrophically balanced cyclonic gyres, where the pycnocline attains its shallowest positions (Nansen, 1906; Helland-Hansen and Nansen, 1909; MEDOC Group, 1970; Stommel, 1972; Lazier, 1973; Lacombe and Tchernia, 1974). Deep convection within the Norwegian-Greenland Seas has not been directly observed, but inferred from water mass characteristics, and more recently from the GEOSECS tritium data set (Petersen and Rooth, 1976).

There are two major cyclonic gyres within the Southern Ocean: the Weddell Gyre and Ross Sea Gyre; a third weaker one may exist east of Kerguelen Plateau (Mosby, 1934; Deacon, 1937). An obvious question is "Does deep convection and water mass formation take place within the centers of these gyres?"

Wüst (1928) questioned the ability of continental margin processes producing all of the deep Anarctic influence in the Atlantic Ocean. He speculates that deep convection occurs in the center of the Weddell Gyre in analogy to deep convection of the Northern Hemisphere. However, Mosby (1934) believes the available summer period data do not support Wüst's speculations. In 1933 Wüst changes his speculation as to the formation area of the open ocean component, but continues to stress that two Antarctic water masses are formed in the Weddell—Antarctic deep water ($\theta = -0.2$ to -0.7°C; $S = 34.62-34.66$‰) formed in the open ocean at the fringes of the winter sea ice in accordance with Nansen's concept, and Antarctic bottom water ($\theta = -0.8$ to -0.9°C; $S = 34.67-34.69$‰) at the continental margins in accordance to Brennecke's (1921) concept.

Observations within the Weddell Gyre in early 1977 present a distinct possibility that open ocean convection and deep water mass formation does indeed occur, most likely within the central region of the Gyre.

Two oceanographic cruises in the Atlantic sector of the Southern Ocean aboard the Argentine Research Ship *ARA Islas Orcadas* were accomplished during the austral summer 1976–77. The second

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[1] LDGO Contribution No. 2688.

0022-3670/78/0600-0612$06.50

cruise, numbered 12–77 (Gordon and LaBrecque, 1977), obtained extensive measurements within the Atlantic-Indian-Antarctic Basin, which will be referred to as the Weddell Basin.

The physical oceanographic program consisted of CTD-O$_2$ hydrographic stations (near continuous record of conductivity, temperature and pressure coupled with an oxygen sensor) equipped with a 24-bottle General Oceanics rosette for chemistry and CTD-O$_2$ standardization. In addition, expendable bathythermograph (XBT) observations were obtained between hydrographic stations. The CTD-O$_2$ data used in this study have been standardized to the results of the rosette sampling for each station. In the case of depth, temperature and salinity these corrections are considered to be final and accurate to 2 m, 0.015°C and 0.005‰, respectively; albeit the final editing of the total *Islas Ochadas* data set (1976–77 and 1977–78 cruises) for CTD noise will be carried out in late 1978. Oxygen correction is considered to have absolute accuracy of $\sim$0.1 ml l^{-1}; improved accuracy is expected upon final data processing.

A number of the CTD hydrographic stations and XBT observations in the vicinity of Maud Rise (Fig. 1) were of particular interest in that they reveal the presence of deep-reaching convective exchange in an area remote from the continental margins of Antarctica. These observations are the subject of this paper.

2. Observations

a. Upper layers

On 7 February 1977, at CTD-O$_2$ station 115, near 67°S, 7°40′W, anomalous thermohaline stratification was encountered (Fig. 2a): the sequence of temperature-minimum/temperature-maximum stratification normally found in Antarctic waters was absent. Instead, there was a warm, low-salinity surface layer occupying approximately the upper 190 m above a cold nearly homogeneous water column. CTD 118, a more typical Weddell Basin station, at 66°S, is shown (Fig. 2b) to demonstrate the uniqueness of CTD 115. The stratification revealed by the CTD 115 downtrace is confirmed by the uptrace, water bottle data obtained from a rosette sampler during the uptrace, and XBT observations using an expanded scale recorded (Georgi, 1977).

After CTD 115 a brief survey pattern (Fig. 3) was followed for approximately 12 h during which XBT and two additional CTD-O$_2$ observations were obtained. The limits of the survey grid were based on reentering the normal stratification of surrounding waters. The ship drift vectors determined from the navigation program, using the onboard IBM 1130 computer, indicates clockwise (cyclonic) surface currents with magnitude of 1.0–1.5 kt (50–75 cm s^{-1}). Since wind was calm and seas smooth, the drift vectors are believed to be a reasonable representation of surface current.

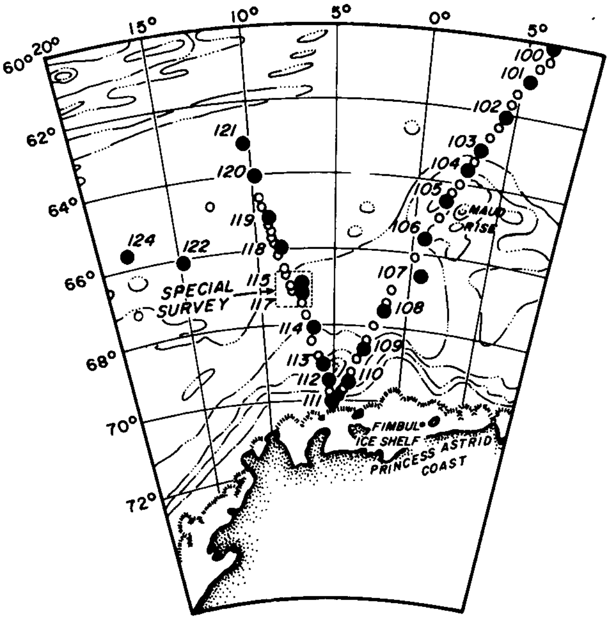

FIG. 1. Positions of CTD hydrographic stations and XBT observations in the vicinity of Maud Rise obtained during cruise 12–77 of the *ARA Islas Orcadas*. The special survey is shown in Fig. 3.

The thermal structure of the upper 500 m along the A–B section of the special 12 h survey (Fig. 4) depicts a vertically elongated (relative to the vertical exaggeration) cold cell of water (<0°C), which is not continuous with the main stratum of winter water. The cold cell is embedded in the warmer Weddell deep water layer (>0.25°C). The distance between the 0.25°C isotherms (from 125–400 m) along section A–B, which traverses the feature, is $\sim$27 km, giving an eddy radius of 13 or 14 km. There is no sea surface temperature expression of the cold cell, making remote sensing of such features unlikely, though there is a surface salinity and silicate expression.

Surface salinity and silicate have high gradients near points A and D, at the southern boundary of the cell, but no significant surface water changes at the northern end, even though normal subsurface stratification resumes. This suggests the eddy circulation may be a bit more complex than revealed in the rather limited present data set.

The large-scale thermal structure with surface salinity and silicate (Fig. 5) further brings out the anomalous character of the cold-core cyclonic eddy. The *T*-min layer becomes colder and thicker on progressing both to the north and to the south of the eddy. The thickness of the colder than 0°C water is nearly 200 m at 63°S and 360 m near 70°S, but

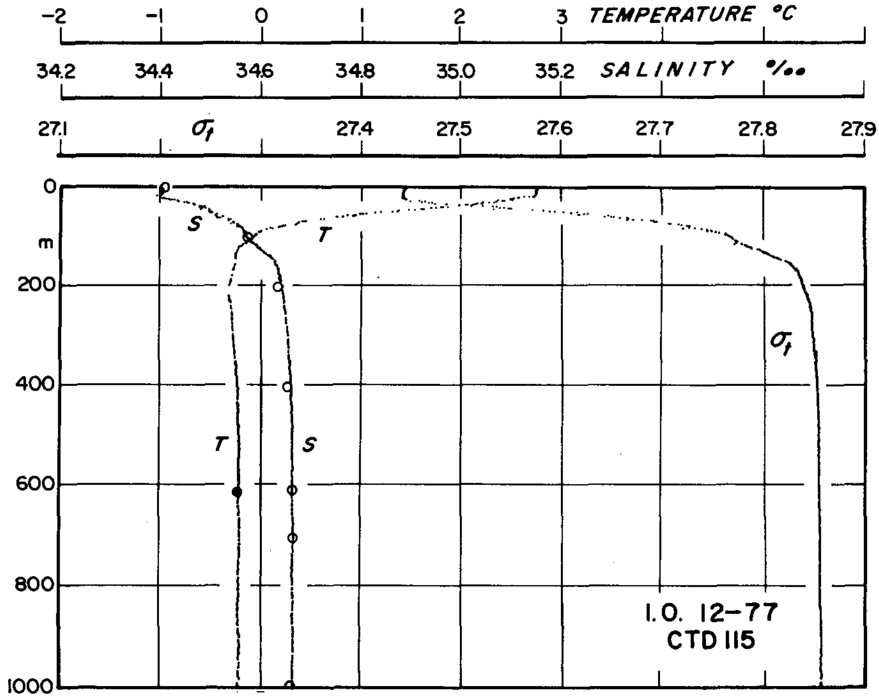

FIG. 2a. Vertical temperature, salinity and sigma-*t* observed at CTD station 115 downtrace (67°02′S, 7°40′W) at 0100 GMT 7 February 1977. The water bottle salinity and temperatures are shown. The water bottle data are obtained by rosette sample during the CTD uptrace, hence some differences may occur from the time of the downtrace. CTD calibration is based only on uptrace values.

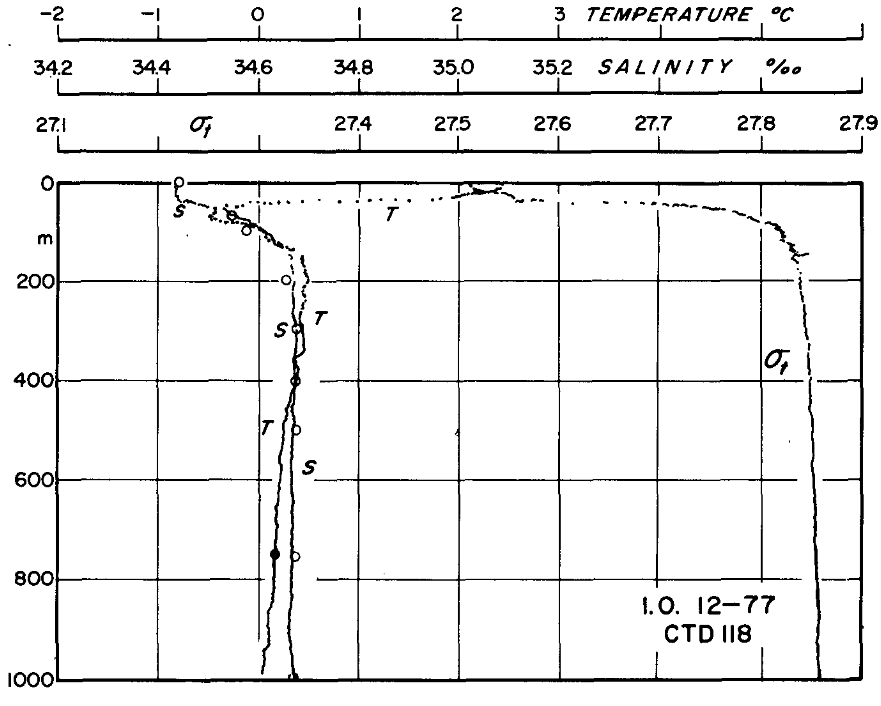

FIG. 2b. CTD station 118 (66°01′S, 8°43′W) at 2200 GMT 7 February 1977.

only 70–80 m in the vicinity of the eddy. At both ends of the section the *T*-min drops to below −1.5°C. Thus the overall characteristics of the winter water are rather attenuated near the cold eddy. The domelike appearance of the warm deep water (e.g., structure of the lower 0°C isotherm) suggests that the section crosses the central part of the large-scale Weddell cyclonic gyre between 66.5 and 68.5°S. This latitudinal position is in agreement with the gyre's central position found to the west during the IWSOE-73 transect from Scotia Ridge to Cape Norvegia (Foster and Carmack, 1976b), and in latitude and longitude of that reported by Treshnikov (1964), though Meyer (1923) and Mosby (1934), using a very sparse data set, position the gyre's center to the northwest (60–65°S, 30–35°W). It is interesting to point out that Treshnikov's surface current map also shows small cyclonic eddies embedded within the larger scale Weddell Gyre.

The surface salinity is highest near 66°S, 110 km north of the eddy. Surface silicate also is higher to the north, reaching 80 ml l^{-1} between 64 and 65°S. The salinity of the *T*-min (station 115's *T*-min salinity is taken at the very weak *T*-min expression at 210 m) increases toward the eddy, but is relatively high throughout the area, attaining values above 34.50‰ north of CTD 114 (68°S). Characteristic winter water salinity in the Weddell Sea is 34.36–34.52‰ (Foster and Carmack, 1976b); however, CTD stations 115, 118 and 119 *T*-min salinities are above this upper limit: 34.65, 34.56 and 34.55‰, in order of the station number.

The salinity of the *T*-min is an interesting parameter since it is believed to represent the winter period surface water salinity. Between the *T*-min and the *T*-max (an interval of about 100–300) salinity is the stabilizing factor. Therefore, the higher the salinity of the winter freezing point surface water (represented in summer by the remnant *T*-min), the lower the static stability of the surface winter layer. The salinity range of the *T*-max layer is only 0.01‰ in the Weddell Basin data of cruise 12-77, thus making the *T*-min salinity with a range of 0.4‰ the dominant factor in determining static stability of the surface water relative to the upper deep water. Vertical flux of heat and salt, accomplished by a variety of possible processes (Foster and Carmack, 1976b), increases as the vertical stability of the water column is reduced. Hence the salinity of the winter water may be taken as a measure of possible vertical interchange with the underlying reservoir of warm, saline deep water. The *T*-min salinity measured during the *ARA Islas Orcadas* cruises 11–76 and 12–77 in the Weddell Basin (Fig. 6) shows significantly higher values to the west of Maud Rise. This may be taken as evidence that the Maud Rise is of some significance to the characteristics of the surrounding ocean and possibly to the extent of the Weddell Gyre. East-west gradients over Maud Rise are also present in temperature of the slightly deeper temperature-maximum and in salinity of the still deeper salinity-maximum.

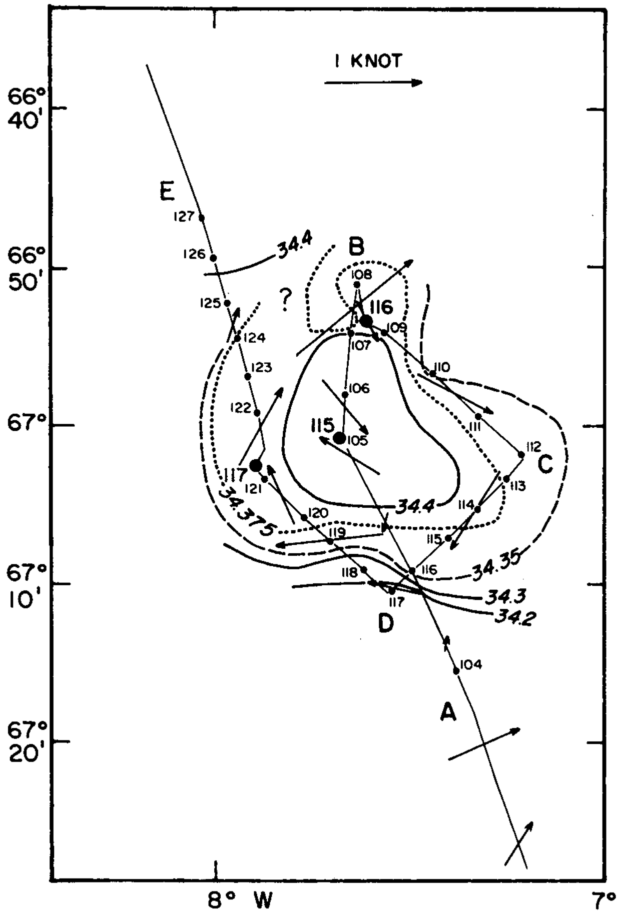

FIG. 3. Survey pattern followed during a 12 h period after CTD station 115 from 0330 to 1515 GMT 7 February 1977. Position of CTD (large dots) and XBT (small dots) observations are shown. The arrows denote ship drift encountered between satellite position fixes. The isopleths are of sea surface salinity. The large letters (A–E) are turning points used in construction of thermal sections for Fig. 4.

b. Full water column

The full depth section of temperature, salinity, oxygen and density (Fig. 7) indicates that the isopleths of all parameters shoal to a remarkable extent within the eddy. All indicators of the deep water (high temperature and salinity, and low oxygen) are breached by the eddy structure. The density section shows the "disturbance" extends to 4000 m.

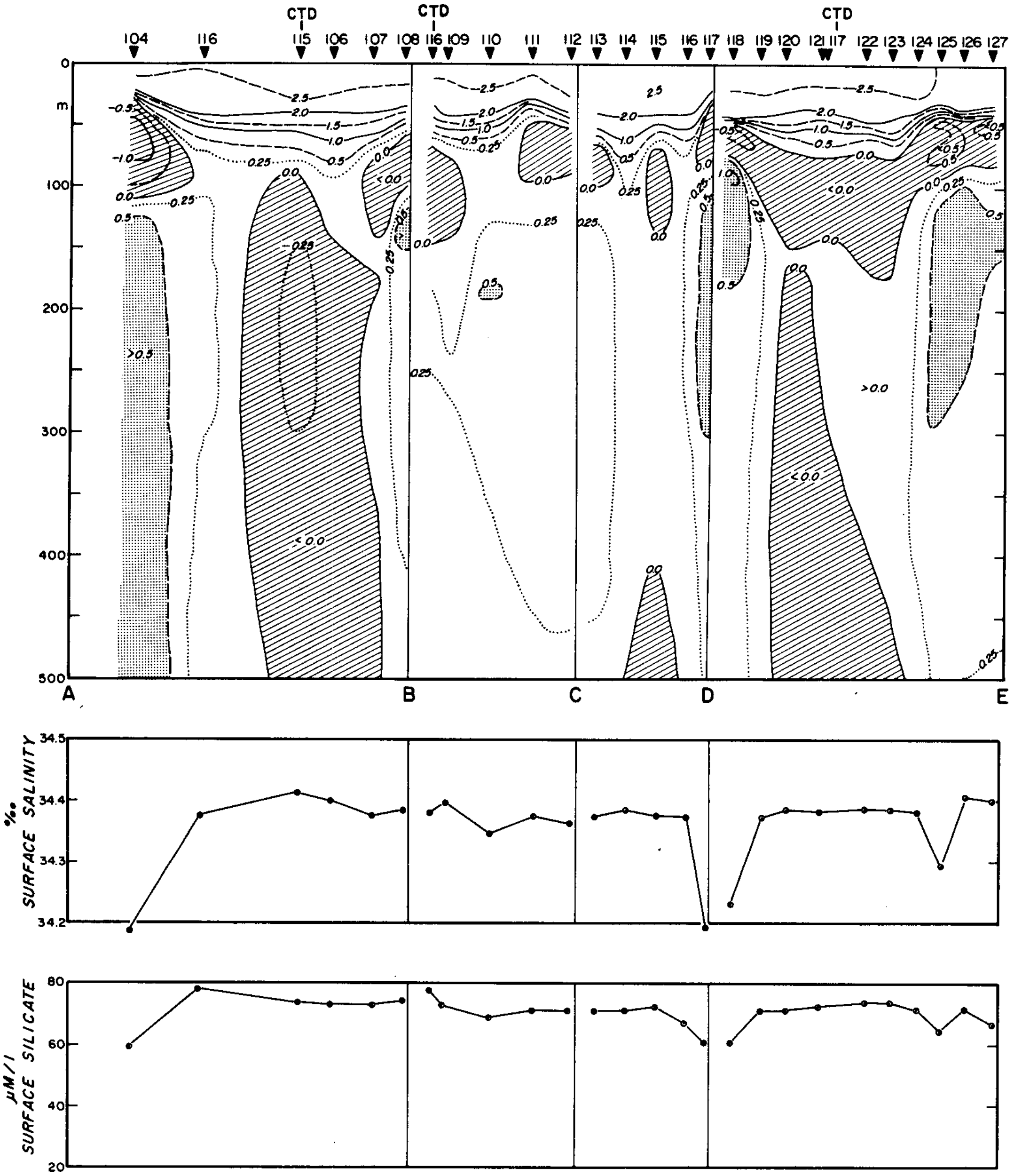

FIG. 4. Thermal structure for upper 500 m, surface salinity and silicate segments of track of special survey shown in Fig. 3.

The deep water potential temperature-salinity correlation (the θ/S curve below the S-max) of Weddell Basin water revealed by the February 1977 *Islas Orcadas* cruise data has much structure (Fig. 8). Discontinuities, or offsets, in the slope of the θ/S curve are common, particularly near −0.1°C (see station 104 θ/S in Fig. 8). Jacobs and Georgi (1977) also report deep water θ/S slope changes in the eastern parts of the Weddell Basin. The structures are believed to be associated with differing advective-mixing histories of segments of the deep water column.

West of Maud Rise a particular deep water structure is present in the 1977 observations: the group of θ/S and θ/O_2 curves for CTD-O_2 stations 104–120 (Fig. 8) reveal a low-salinity, high-oxygen feature between sigma-2 levels 37.21–37.23, (usually lying between 1500 and 2000 m). This feature is well developed at stations 107, 108 and 109. In other stations the salinity signature is less pronounced, but

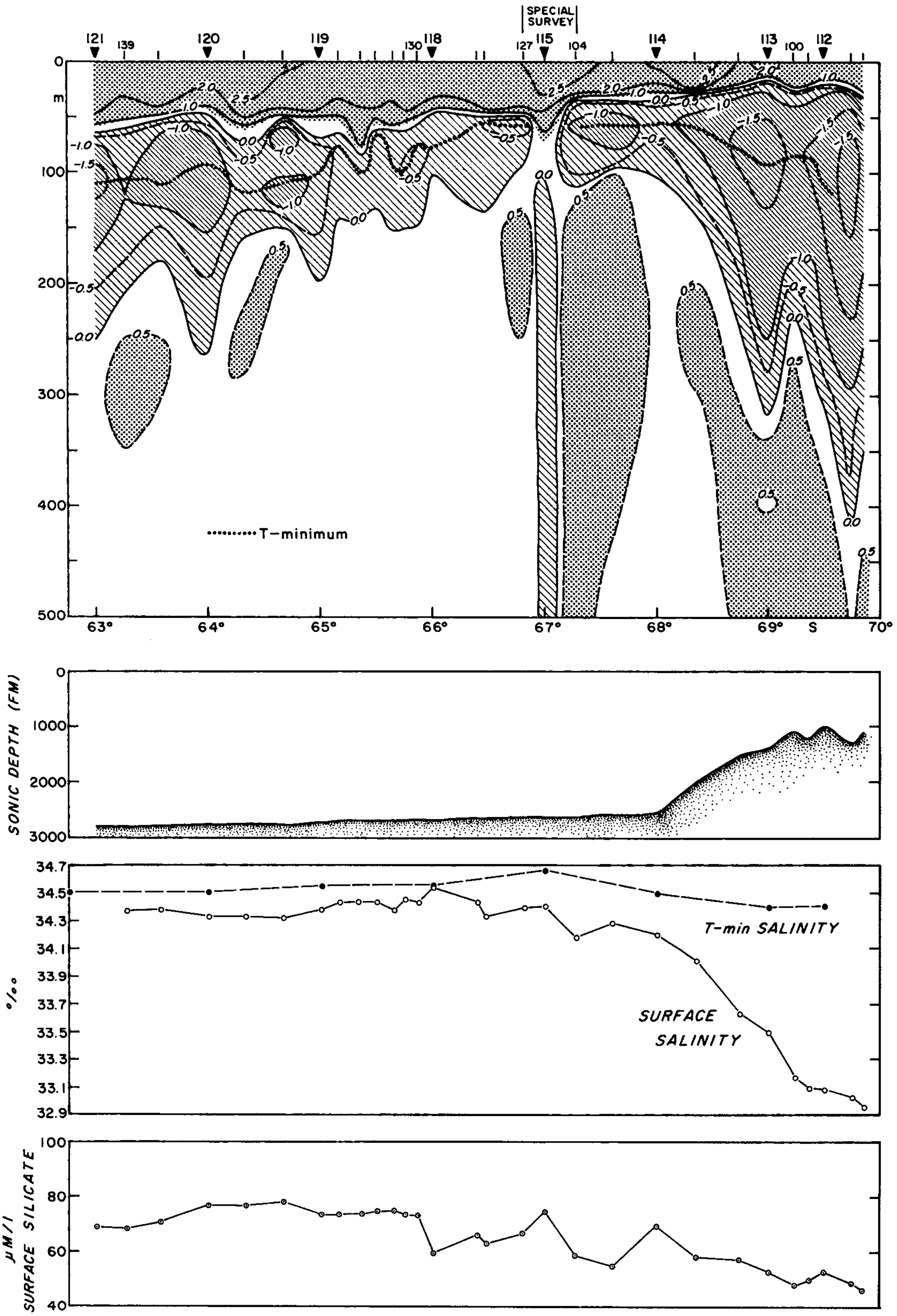

FIG. 5. Thermal structure for upper 500 m, surface salinity and silicate, salinity of the T-min layer, and bottom topography along ship track between 70 and 63°S (see Fig. 1).

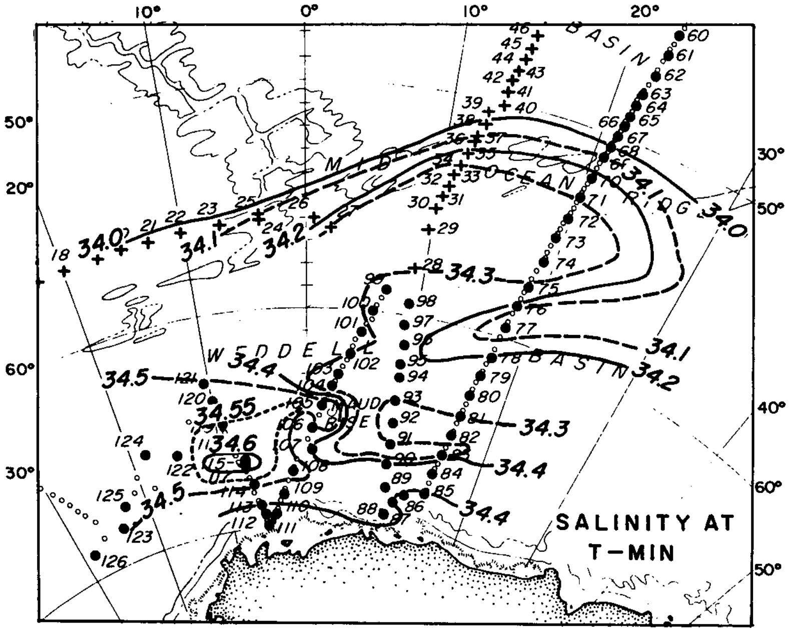

FIG. 6. Salinity of the temperature minimum layer measured by the *ARA Islas Orcadas'* cruises 11–76 and 12–77 CTD hydrographic stations (large solid dots) within the Weddell Basin. The open circles are XBT observations which do not contribute to the salinity data set.

the high-oxygen feature is generally present. In any case, the salinity in this sigma-2 range is low when compared to a straight line reference between the S-max and bottom θ/S point, and hence represents a distinct water type.

3. Discussion

a. Eddy characteristics

The cold-core eddy was found without previous knowledge of its existence, following a line of hydrographic stations at 1° latitude spacing. In view of the small size of the eddy, either we were extremely lucky in finding a rare eddy in the oceanographic zoo [following the term used by Stommel *et al.*, (1977)] of eddies or, as is more likely the case, such eddies are relatively common within the central regions of the Weddell Gyre. It is significant to note that the Norwegian research program aboard the *Polarsirkel*, following a 1° latitude station spacing schedule, encountered a similar, if not the same, feature about 30 n mi southwest of CTD 115 two weeks later (Arne Foldvik, personal communication, 1977).

Inspection of the historical hydrographic data in the NODC files for the Weddell Basin west of 10°E and south of 60°S, exclusive of the continental margins where the sonic depth is less than 2000 m reveals only one station without the warm deep layer. The criteria used to denote deep water absence or significant weakness in stratification is lack of $T > 0°\mathrm{C}$ below the surface layers. The station is from the Argentine R.V. *San Martin* station number 6 at 66.333°S, 11.250°W, taken on 13 January 1961. The T-max of −0.07°C occurs at 1942 m (the deepest observation of the station) with a salinity of 34.70‰. The salinity is constant at 34.70 below 187 m and sigma-t attains a value of 27.88 at 141 m. The shallowness of the 34.70‰ isohaline and 27.88 sigma-t isopleth, coupled with the lack of greater than 0°C water (above 1942 m) is quite similar to the CTD 115 observation. The major difference between the *San Martin* and *Islas Orcadas* observations is that the 1961 station has a strong T-min at 75 m (−1.77°C).

The rarity of observations of eddies prior to 1977 does not rule out more common occurrence, since the historical data set is very sparse and represents only a few years of the last five decades. It is noted that the historical data shows a great range of the salinity at the 200 m horizon within the central region of the Weddell Gyre, suggesting significant variability of the vertical stratification.

The eddy observed in 1977 is in the area in which nearly every winter-spring the sea ice cover thins or disappears, resulting in a polynya (Fig. 9). Streten (1973), reporting on satellite observations, mentions the frequent occurrence of open water in the Weddell Sea near Kapp Norvegia in early spring. The Antarctic ice charts for 1974, 1975 and 1976 (Fleet Weather Facility), show polynya development in late winter, or weak ice throughout the

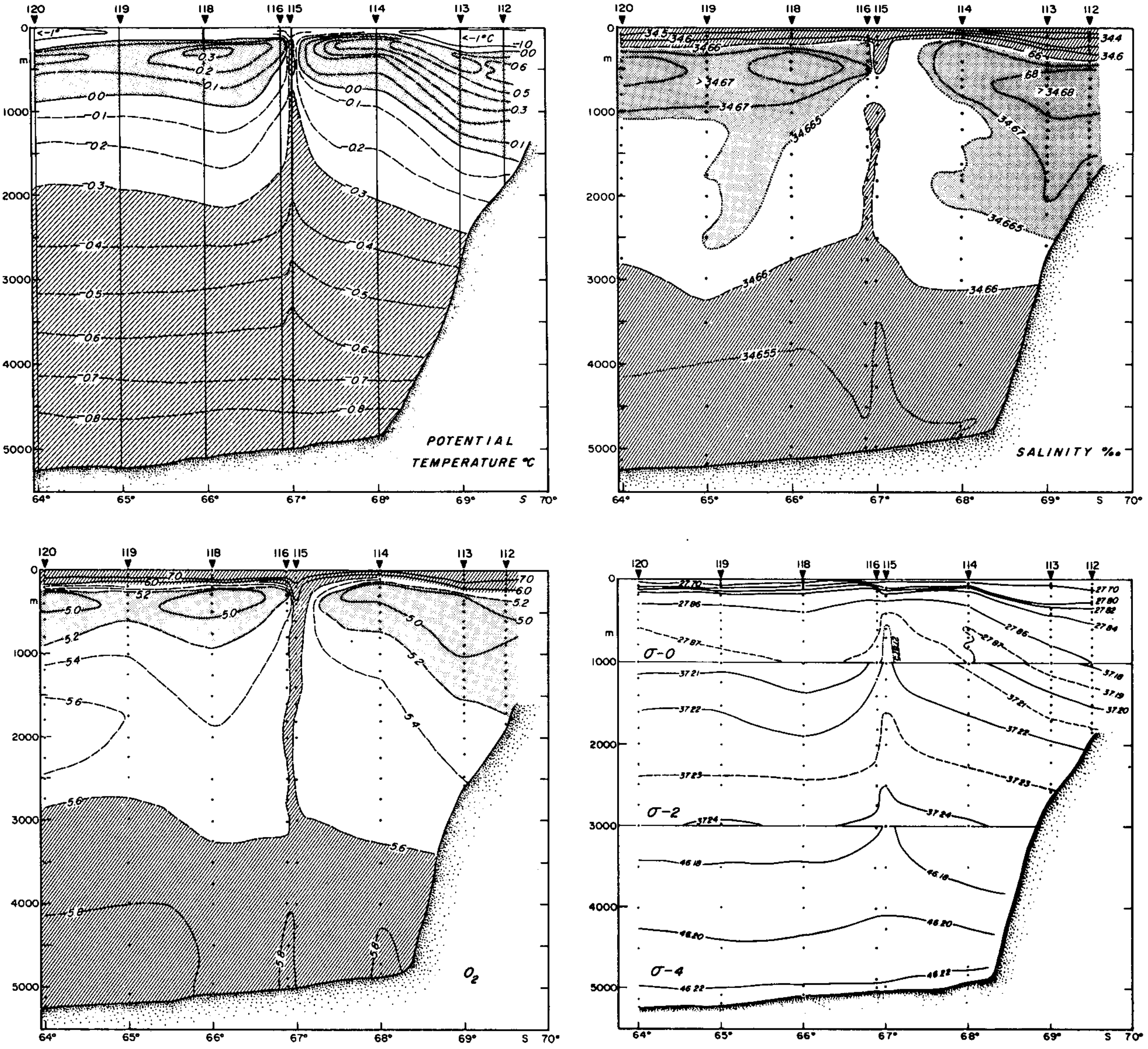

FIG. 7. Characteristics along the track between CTD stations 112 to 120 (see Fig. 1): temperature (a) from CTD (vertical lines denote CTD trace position), salinity (b) from rosette bottles, oxygen (c) from rosette bottles, and sigma-p (d). Dots on Figs. 7b, 7c and 7d are water bottle observations.

winter. The 1977 ice charts indicate no polynya development. The polynyas for the 1974–1976 period were situated mainly west of Maud Rise, although in 1974 a substantial part of the polynyas extended to the east of Maud Rise.

b. Rossby parameters

The Rossby Number Ro = V/Lf (where V is the characteristic surface velocity, taken as 50 cm s^{-1} based on ship drift, L is the characteristic radius, 14 km; and f the Coriolis parameter) of the eddy observed by *Islas Orcadas* is 0.27. Therefore, one-fourth of the sea surface pressure gradient (8.5 $\times 10^{-3}$) is compensated by inertial terms and the remaining three-fourths by the Coriolis term.

The Rossby radius of deformation is $\lambda = NH/f$, where N is the Brunt-Vaïsälä frequency given by $N^2 = (g/\rho_0)(\partial\rho/\partial z)$, H is the thickness of the feature (about 4000 m), and f the coriolis parameter. In view of the high compressibility of the low-temperature antarctic waters, it is appropriate to determine the density difference between 200 to 4000 m (the seasonally warmed surface layer is not included in determination of the Rossby deformation radius), using the sequence of σ-p for p of 0, 1, 2, 3, 4 (p refers to the pressure surface used for reference in units of 10^3 db). In this case the density difference in sigma

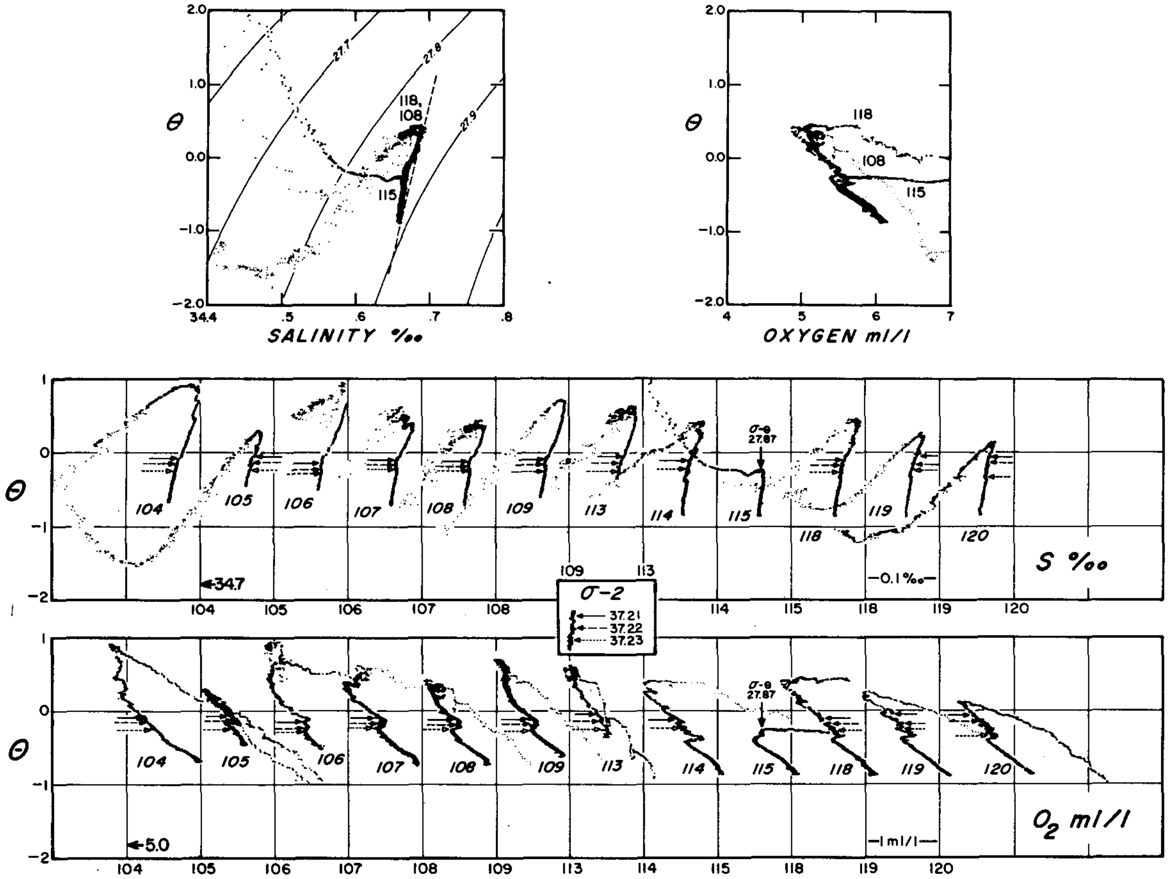

FIG. 8. Sequence of θ/S and θ/O_2 curves for deep water stations west of Maud Rise. The salinity and oxygen corrections applied to the CTD-O_2 are preliminary, since the final processing stage will not be reached until the 1977/78 cruises are completed. The corrections are very close to expected final values.

units is 0.06, yielding a Rossby radius of deformation of 11.0 km. This is only slightly below the eddy radius. Due to uncertainties in precisely determining the eddy radius with the available data, and in establishing the applicable density difference and depth for Rossby radius of deformation calculations, it is concluded that the observed eddy is approximately the same scale as the deformation radius or possibly slightly larger.

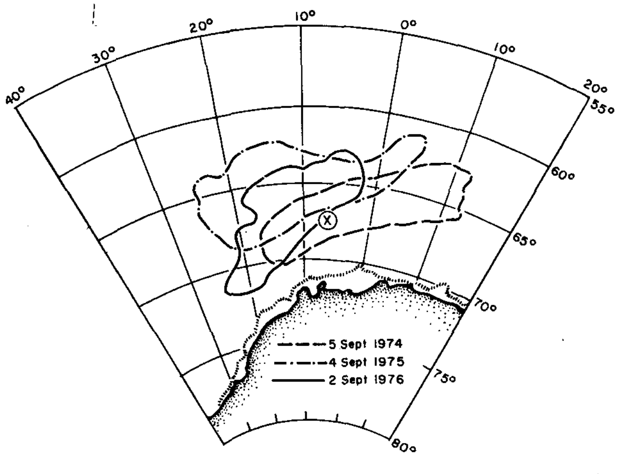

FIG. 9. Limits of Weddell polynya revealed in the 1974, 1975 and 1976 sea ice maps of the Fleet Weather Facility for the first week of September. In 1973 and 1977 no late winter polynya is shown on the sea ice map.

c. The hypothesis

The hypothesis offered to explain the *Islas Orcadas* observations is as follows: the central region of the Weddell Gyre is one of generally low stability due to the Ekman-induced upwelling and the geostrophically balanced large-scale cyclonic flow of the Weddell Gyre, leading to a relatively thin layer of stable water. Winter period sea-air flux would induce large thermohaline alterations of this thin surface water layer. Both of these factors (upwelling and low-volume surface layer) combine to produce a relatively dense and saline surface water, as substantiated by the high salinity T-min winter water west of Maud Rise. Thus the central region of the Weddell Gyre may, in winter, be the site of a neutrally stable or unstable water column, giving rise to convective events, the remnants of which may be cold core eddies.

It is possible that the eddy does not have a convective origin; it may have been the product of baroclinic instability [in a manner in which the MEDOC eddy may have formed (Gascard, 1977)] near the margins of the Weddell Gyre, or perhaps near Maud Rise [where a strong east-west thermohaline gradient in the upper kilometer was observed in the *Islas Orcadas* data set (see Fig. 6)]. The eddy would then migrate toward the center of the Gyre where winter conditions may induce convection (Gascard, personal communication, 1977).

The hypothesis suggests that the Weddell polynya is a response not only to mechanical removal of the sea ice by Ekman-induced divergence, but also to attenuation, or melting, of sea ice by upward heat (and salt) flux associated with vertical instability of the surface water within the center of the Weddell Gyre. Were this not the case, Ekman divergence might not induce a polynya, since it does not do so elsewhere, though such divergences are widespread over the Southern Ocean (Gordon and Taylor, 1975).

d. *Probable number of eddies in 1977*

By assuming that the late winter polynya characteristic of the Weddell region delineates the area affected by the instability, it is possible to make a rough guess at the maximum number of cold-core eddies that may exist in the austral summer of 1977 within the hub of the Weddell Gyre. The polynya, in the beginning of September, is approximately 5° latitude by 20° longitude, or 5×10^5 km^2. Since a randomly positioned north-south line of CTD and XBT observations across this area (with XBT observations spaced no more than 30 km apart) located one eddy, it is possible that a series of north-south lines 30 km apart (the eddy diameter being a little less than 30 km) would each find one eddy, making a total of 33 eddies. Therefore, the number of eddies that may have been in existence in early 1977 is more than 10 less than 100.

e. *Vertical processes*

When the static stability of the surface layer decreases to zero within the gyre center, cabaling may become important (Foster and Carmack 1976b; Neshyba, 1976), or—in some years, if not all years—a region of deeply penetrative convection may occur similar to the "violent mixing" phase of the sequence of events observed by the MEDOC Group in 1969 and 1970 (MEDOC Group, 1970; Stommel, 1972; Sankey, 1973; Killworth, 1976). These factors would tend to further increase surface water salinity and hence have a positive feedback influence on the Weddell Gyre circulation.

In the latter case the observed cold-core eddy may be a remnant of a patch of anomalous deep winter water. Some "chimneys" (to borrow a term used in MEDOC discussions) may not be preserved into the summer period but may disappear relatively quickly during a "sinking and spreading" phase, as apparently was the case in the MEDOC studies. The MEDOC observations show that the chimney broke up quickly after the violent mixing phase due to shearing of deep water relative to the shallow layers (Sankey, 1973) or to baroclinic instability (Killworth, 1976). These factors are apparently not as active on the 14 km radius spatial scale within the Weddell Gyre, since the eddy integrity is preserved for a longer period.

The θ/S and θ/O_2 relation of CTD station 115 (Fig. 8), within the cold-core eddy, is highly anomalous in comparison to the surrounding stations, in view of the absence of a well defined T-min and T-max. An abrupt change in the θ/S curve of CTD 115 occurs near 27.87 σ-θ at 680 m. Below that level the θ/S correlation enters into the deep water θ/S form (decreasing salinity with decreasing temperature). The sigma-θ isopleth of 27.875 (above 1000 m) forms a nearly continuous isopycnal with the sigma-2 37.22 (in the 1000–3000 m layer), which marks the central density surface within the low-salinity, high-oxygen stratum of surrounding stations. Hence the 27.875 isopycnal can serve as the pathway for isopycnal spreading of near-surface water into the deeper layers.

The similarity of density of the surrounding low-salinity, high-oxygen stratum with the base of the surface layer within the cold-core eddy is probably more than merely a coincidence. Should the surface layer, revealed by CTD station 115, represent seasonal heating and freshing, then the previous winter period surface mixed layer within the cold-core eddy would allow exposure of the sigma-2 37.21 to 37.23 density stratum to atmospheric effects which could account for the anomalous low salinity, high oxygen of that layer.

The volume of dense surface water sinking within the chimney would be limited by geostrophic effects, which would develop in response to the initial net horizontal convergence of surface water toward the chimney (Solomon, 1974). Therefore, each chimney might not contribute much "new" water to the deep ocean, especially if a MEDOC-style sinking and spreading phase does not occur, but would permit the water within the deep density stratum to have access to the winter period surface layer within the chimney. In this way, exchange of colder, fresher surface water with warmer, saltier deep water may occur on the strongly sloping isopycnals, resulting in a net upward flux of heat and salt into the winter surface mixed layer. The surface water characteristics may infiltrate by diffusion or stirring into the deep water, without a wholesale flux of a water mass by advection.

If all the water of the upper kilometer of each of the 33 eddies which may have existed in early 1977

TABLE 1. Ocean nonradiative heat loss (cal cm^{-2} day^{-1}).

		Autumn	Winter
Evaporative	Australian sector 60–70°S (Zillman, 1972)	50	100
	South Atlantic 67°S (Viebrock, 1962)	120	240*
Sensible	Australian sector 60–70°S (Zillman, 1972)	30	50
	South Atlantic 67°S (Viebrock, 1962)	150	250*

* Assuming the ratio of autumn to winter nonradiative heat loss, found in the 60–70°S Australian sector, is also valid in the South Atlantic at 67°S.

(as discussed above) were to ultimately contribute to renewal of water characteristics at the sigma-2 37.22 level, and each eddy was 14 km in radius, a total of 2×10^{13} m^3 of surface water would be transferred to a deep layer. Evenly distributed over one year, this yields a rate of 0.6×10^6 m^3 s^{-1}. It is probable that this estimate is not very meaningful; perhaps it is low since eddies which do not endure into the summer are not included, and thus are not explicitly part of the "estimated" statistics of one eddy per north-south line of stations.

f. One-dimensional model

Not knowing the sequence of winter events, nor the initial size and stratification characteristics of the cold-core eddy, we can only speculate as to its origin and similarities to the MEDOC process. Should the eddy exist into the autumn/winter period, the seasonally warmed surface layer may be removed to again expose the deep water to the winter atmosphere's effects.

The time required to produce unstable stratification between the winter water and deep water within the eddy and in surrounding waters may be estimated from a one-dimensional heat flux model. The following is an attempt to determine whether reasonable assumptions of sea-air heat flux values would remove vertical static stability observed in the summer, and if so in what month might we expect neutral stability to be attained.

The radiation balance for April and July in the 60–70°S belt is −36 and −69 cal cm^{-2} day^{-1}, respectively (Sasamori *et al.*, 1972; Tables 2.7, 2.8 and 2.11). The evaporative and sensible heat loss are less certain, but estimates can be made for the Weddell region. For the autumn (April–June) and winter (July–September) periods Zillman (1972) gives the nonradiative ocean heat loss for the Australian sector between 60 and 70°S (Table 1). Viebrock (1962) determined autumn sensible heat loss near 67°S in the South Atlantic region (his Fig. 8), and using his evaporative rate values (Viebrock's Fig. 1), the evaporative heat loss for the autumnal South Atlantic at 67°S is determined (Table 1). If the ratio of Australian to South Atlantic nonradiative heat flux found in autumn is valid in winter, then South Atlantic winter values would be slightly less than 500 cal cm^{-2} day^{-1} (Table 1) for non-ice cover conditions.

The total autumn and winter heat loss (radiative and nonradiative) to the atmosphere is about 110 and 200 cal cm^{-2} day^{-1}, respectively, using Zillman's data (60–70°S, south of Australia). Using Viebrock's autumn data, a Weddell heat loss of 300 cal cm^{-2} day^{-1} is likely, with a winter value of ~570 cal cm^{-2} day^{-1} (near 67°S in the South Atlantic). As a basis for an estimation of the time necessary to produce unstable conditions in the observed eddy, a heat loss of 300 cal cm^{-2} day^{-1} is used for the autumn and 570 cal cm^{-2} day^{-1} for the winter period (ice-free conditions).

The stratification of CTD 115 can be viewed in a simple manner as a halocline structure with relatively warm water to 190 m, below which the water column is essentially homogeneous in a layer to 680 m and in another layer to abyssal depths. The water column would be isothermal with removal of 1.49×10^4 cal, requiring a little over 1.7 autumn months (mid-March to early May) to accomplish. The surface layer salinity would be 34.58‰, assuming evaporation is balanced by precipitation. The rest of the autumn months' and winter months' heat loss (early May and June to late July) are required to reduce the upper 190 m to the freezing point and begin ice formation. At this time, the density of the surface layer is sigma-*t* 27.86, which is only 0.01 below that of the deeper water.

Formation of approximately 10–15 cm of sea ice would introduce enough salt into the surface water to increase the surface water density to that of the deep water and hence induce vertical overturning. The time required to form the necessary amount of ice to reach unstable conditions depends on the ratio of ice cover to open water [since sea-air heat exchange is strongly inhibited by sea ice (Fletcher 1969)], and so it is not possible to calculate the winter ice-cover heat budget with any assurance of accuracy. Only a few days of winter open ocean heat loss would be needed to produce 10–15 cm of ice. It is reasonable to conclude that by the beginning of August static stability within the observed eddy would be zero or negative and the enhanced vertical exchange of water mass properties would melt, or at least attenuate further sea ice formation.

The calculation can be repeated for a normally stratified condition within the hub of the Weddell Gyre, as revealed by the neighboring CTD station 118 (Figs. 2b and 5). The surface layer would attain freezing temperature conditions to 160 m (marking the extent of the surface water seasonal halocline) by mid-July. At this time only 5–10 cm of ice are required to introduce enough salt to develop neutral stability. Therefore, unstable conditions could again occur by the beginning of August, as is the case of the water within the eddy.

These calculations reveal the susceptibility of the entire area to winter period instability and deep convection with possible upward flux of heat and salt into the surface layer which would act to severely limit sea ice growth.

4. Significance of observations

Open ocean convection within the central Weddell cyclonic gyre to sigma-2 levels of 37.21–37.23 provide another locale and mechanism for Southern Ocean deep vertical exchange. This is in addition to the more traditional view of Antarctic Bottom Water formation near the continental margins and Antarctic Intermediate Water formation in the vicinity of the polar front zone. The water mass formed within the Weddell Gyre may be called Weddell Deep Water (Reid *et al.*, 1977) or by the older term of Antarctic Deep Water, as suggested by Wüst (1933). The characteristics of the water within the deep low-salinity, high-oxygen feature at sigma-2 37.21–37.22 is close to the warm saline end of the range of Antarctic Deep Water as defined by Wüst.

Piercing of the pycnocline separating the surface and deep water masses need not be a yearly event. Perhaps convection into the main body of the pycnocline, but not through it, occurs more frequently, or as a precursor to deeper convection. Note that pycnocline water is the Modified Warm Deep Water of Foster and Carmack (1976a), which they believe contributes to formation of Antarctic Bottom Water in the vicinity of the continental shelf frontal zone. Therefore, processes within the hub of the Weddell Gyre may also be significant by contributing the Modified Warm Deep Water involved in formation of bottom water at the continental margins.

Reid *et al.* (1977) associated water mass boundaries with layers of increased vertical stability. They were able to associate each layer with a boundary of a known water mass except one, stable layer number 8 in the Weddell Sea. Layer 8 falls near sigma-2 value 37.23 (Reid *et al.*, 1977; Fig. 6b), which is also the lower boundary of the observed low-salinity, high-oxygen intrusion. Reid *et al.* (1977) believe the water between stable layers 7 and 8 may be ". . . the denser water of Atlantic (Weddell Sea) origin". The data of *Islas Orcadas* (December 1977) support this speculation.

The water below stable layer 8 must be the Antarctic Bottom Water formed along the continental margins, which may have been partly diluted during a circuit or two into the Weddell Gyre (Carmack and Foster, 1977).

Assuming water mass characteristics spread most efficiently on density surfaces, it is significant to point out that the northward continuation of the sigma-2 37.2 surface and its sigma-4 continuation at 46.1 (Reid and Lynn, 1971; Fig. 3a) is the deepest isopycnal to clear the crest of the Rio Grande Ridge, separating the Argentine Basin from the Brazil Basin. Reid *et al.* (1977, Fig. 6b) also show the sigma-2 37.208 and its continuation of sigma-4 46.13 clears the ridge. Therefore, the Weddell Deep Water characteristics which form in response to processes within the center of this gyre may, by isopycnal spreading, contribute to the bottom water characteristics north of the Argentine Basin. This implies the denser continental-margin-produced Antarctic Bottom Water may be limited by isopycnal spreading to within the Weddell, Argentine and possibly the Crozet Basin grouping, though some may escape to the north through the Vema Gap.

It is interesting to note that Wüst (1933, Fig. 5) uses his θ/S definition of Antarctic Deep Water (at this point he calls it "mainly deep water") as the end point for determining percentage mixtures of Antarctic Bottom Water and North Atlantic Deep Water. Hence he implies it is the open ocean produced Antarctic water mass that supplies the Atlantic rather than the continental margins produced Antarctic Bottom Water. The θ/S break at 680 m on the CTD-115 curve falls directly on Wüst bottom water and deep water mixing curve at the 93% bottom water concentration. Perhaps the other 7% is the continental margin water component.

5. Conclusion

The *ARA Islas Orcadas* cruise 12–77 observations west of Maud Rise in February 1977 are thought-provoking. They indicate that deep vertical exchange of water within the Weddell Gyre, remote from the continental margin, occurs and may be responsible for formation of at least some component of Antarctic Bottom Water found in the more northern basins.

Hopefully, these very preliminary observations will encourage further field and theoretical work within the central regions of large-scale cyclonic gyres, in particular the Weddell Gyre, with regard to water mass formation.

A line of closely spaced hydrographic stations across a cold-core eddy in the Weddell Gyre would allow more quantitative determination of the geostrophy, available potential energy and detailed structure of the eddy. Because of the remoteness of the Weddell Gyre from ports (i.e., transit time) and the time necessary to search for and survey an eddy, such a study would have to be the primary goal of the cruise. As pointed out by Mosby (1934) the sea ice conditions within the Weddell Gyre vary significantly from year to year and corresponding variability in the structure of the surface water is expected. Hence the observed eddy in 1977 need not occur every year, just as the polynya does not form every year (1977 austral winter being the case in point).

Acknowledgments. Herr Prof. Dr. Georg Wüst died on 8 November 1977. I dedicate this paper, in which many of his early ideas are supported, to his memory.

The collection of the circumpolar survey data set aboard the *Ara Islas Orcadas* in the austral summer of 1976–77 was supported by Grant DPP 74-12838 and its successor DPP 76-81240 from the Division of Polar Programs of the National Science Foundation. The scientific success of the *Islas Orcadas* results from the hard work of a number of people from Lamont who "put together" the CTD-O_2/computer system and engaged in careful collection of the data: D. Georgi, E. Molinelli, E. Draganovic, and T. Baker. The manuscript has been critically reviewed by S. Jacobs, D. Georgi and E. Molinelli. Comments from P. Killworth and J. Gascard are appreciated, as well as those of the anonymous reviewers.

I am thankful for the aid given by the Servicio de Hidrografia Naval of Argentina in making the *Islas Orcadas* cruises productive and comfortable. Special thanks are given to the science officers R. Parodi, E. Rodriguez and A. Concela, and to the ship's commandante, Don Eduardo G. Lestrade.

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