Weddell Deep Water variability

by Arnold L. Gordon[1]

ABSTRACT

Temperature of the Weddell Deep Water west of the Greenwich Meridian in 1977 is dramatically lower than that observed in the same region in 1973. The most intense thermal alteration extends from approximately 200 m to 2700 m. Cooling, which averages 0.2°C, accompanies a nearly density compensating salinity decrease, averaging 0.02‰. The most intensive cooling and freshening of 0.4°C and 0.03‰ respectively, occurs within a region of about the same size and position of the Weddell Polynya, as observed by satellite images in the mid 1970's. It is suggested that the alterations of Weddell Deep Water are due to convective introduction of winter surface water. This hypothesis is consistent with the convective model for the Weddell Polynya.

The rate of convective sinking of winter freezing point surface water required to account for the observed cooling is estimated as 1.6 to 3.2×10^6 m^3/sec during the three years of a fully developed polynya (1974-1976). If all the convection occurs during five months of regional sea ice cover (July to November) in each of the three years, the rate is 3.8 to 7.7×10^6 m^3/sec. This is a substantial amount of surface water contribution to abyssal levels, relative to the estimated surface water contribution to bottom water formation over the continental margin.

1. Introduction

In January-February 1973 a series of STD (Plessey 9040 Salinity-Temperature-Depth recorder) hydrographic stations were obtained from *Glacier* on a line from the Scotia Ridge near 60S, 30W across the Weddell Basin to Cape Norvegia, Antarctica near 71S, 12W (Fig. 1; Carmack and Foster, 1973). The *Glacier* 1973 data have been used extensively in studies by Carmack and Foster (1975); Foster and Carmack (1976a, b). Temperature and salinity distribution along the section (Fig. 2) shows the relatively warm saline Weddell Deep Water (WDW) to be in excess of 0.4°C potential temperature and 34.68‰ salinity. The depth and magnitude of the temperature maximum measured by *Glacier* is in agreement with the regional temperature maximum core layer distribution given by Deacon (1979, Fig. 3). The colder WDW measured at the northern *Glacier* stations is part of a zonal band of relatively cold water extending well to the east along 60S and can be associated with the outflow from western margins of the Weddell Basin (Deacon, 1976, 1979). A volu-

1. Lamont-Doherty Geological Observatory of Columbia University, Palisades, New York, 10964, U.S.A.

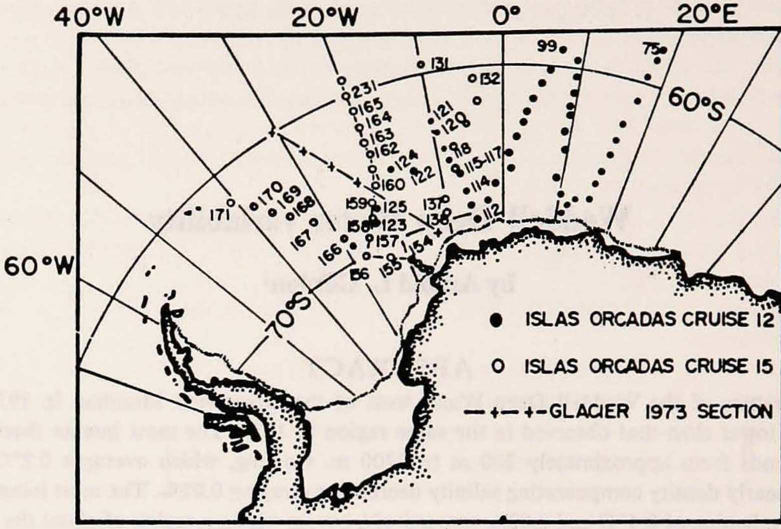

Figure 1. Position of hydrographic stations from cruises 12 and 15 of the *Islas Orcadas* in the Weddell Basin. The position of the *Glacier* 1973 section from the Scotia Sea to Cape Norvegia is shown. Table 1 in text lists coordinates of stations used in Figures 5-8. Stations 230 and 231 (northernmost stations along 20W) are from *Islas Orcadas* cruise 16. (See Huber *et al.* (1981a,b,c) for the complete *Islas Orcadas* stations map for the South Atlantic as part of the circumpolar survey.)

metric θ/S study of the Oceanic Domain of the western Weddell Basin waters shows a total of 566×10^3 km^3 of WDW with potential temperature above 0.4°C and 76 $\times 10^3$ km^3 above 0.6°C (Carmack and Foster, 1977).

In 1976 to 1978 the *Islas Orcadas* (formerly the USNS *Eltanin*), was used to obtain an array of CTD (Neil Brown Conductivity-Temperature-Depth recorder) hydrographic stations in the Atlantic sector of the Southern Ocean (Gordon, 1978a). The stations within the western hemisphere segment of the Weddell Basin are shown in Figure 1. A section nominally along 20W during cruise 15 (January-February 1978; Fig. 4) reveals a much colder WDW layer than measured in 1973. The WDW potential temperature maximum attains values above 0.4°C only near the Antarctic continental margin and north of the Weddell Basin. The temperature maximum within the Weddell Basin ranges between −0.2 to +0.2°C. From the 950 to 1400 km marks, the deep water temperature maximum does not rise above 0°C. This feature is referred to as the "cold spot." The salinity within the deep water is near or below 34.68‰. Within the "cold spot" the salinity is near 34.66‰. Oxygen concentration associated with the "cold spot" is above 5.0 ml/l.

Temperature distribution at the potential temperature maximum core layer constructed from the 1976-1978 *Islas Orcadas* data set (Fig. 5) reveals significantly

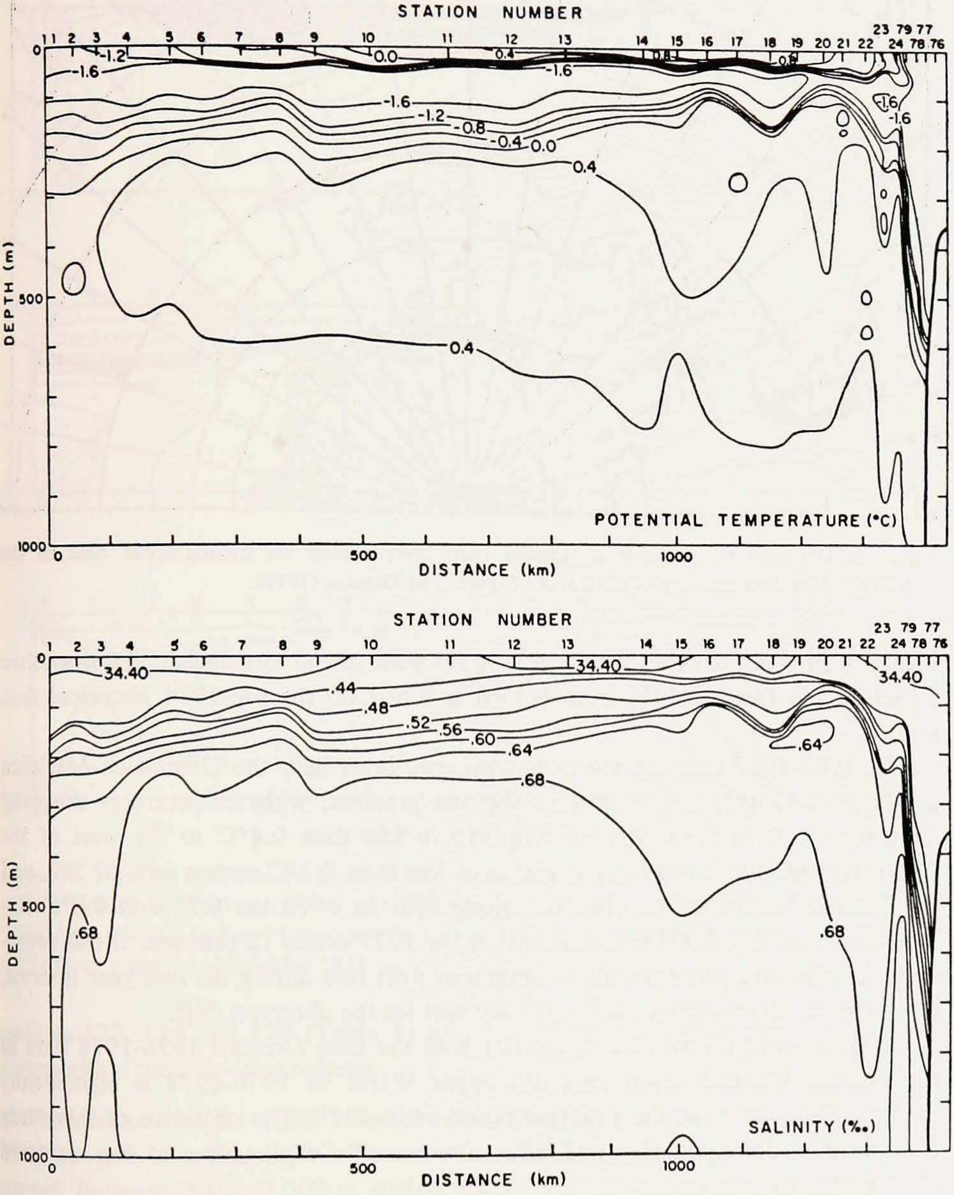

Figure 2. Potential temperature and salinity section across the Weddell Basin based on *Glacier* data January-February 1973. Reproduced from Figure 2 of Foster and Carmack (1976a) publication in the Journal of Physical Oceanography. Station 1 defines the northern limit of the section.

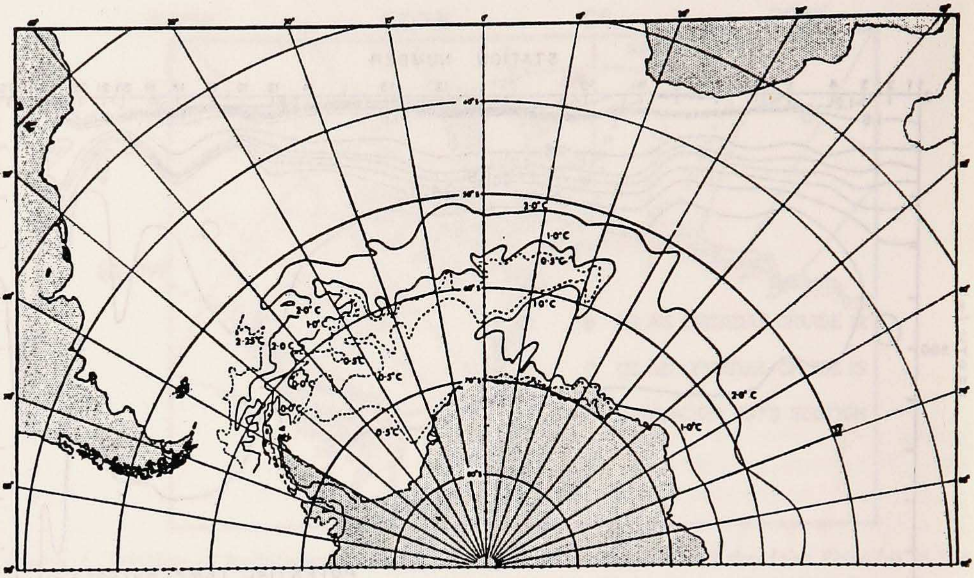

Figure 3. Potential temperature maximum core layer, based on hydrographic data in the NODC 1974 data set. Reproduced from Figure 2 of Deacon (1979).

colder deep water within the Weddell Basin west of the Greenwich Meridian than shown in the Deacon 1979 map (which is based on the pre-1974 historical data set).

The 1976-1977 temperature maximum core layer near the Greenwich Meridian displays a relatively large east-west thermal gradient, with temperatures dropping from near 1°C in the eastern hemisphere to less than 0.4°C to the west of the Greenwich Meridian. An extensive area of less than 0.2°C occurs west of 5W, and the subzero "cold spot" is observed along 67S. In 1978 the 0.2° and 0.0°C isotherms occur about 400 km to the west of the 1977 cruise 12 position. If this represents an advective process with a continuous drift rate during the one year interval, a velocity of 1.3 cm/sec is required to account for the observed shift.

Comparison of the *Glacier* 1973 data with the *Islas Orcadas* 1976-1978 data in the western Weddell Basin indicates upper WDW in 1976-1978 is significantly cooler and fresher than the situation observed in 1973. The objective of this study is to further document this and offer a reasonable explanation of the observed differences.

2. Water column comparison 1973 and 1977

Comparison of profiles of potential temperature, salinity and sigma-θ versus depth of *Glacier*, 1973 stations 11, 13 and 16 with *Islas Orcadas*, cruise 12 1977

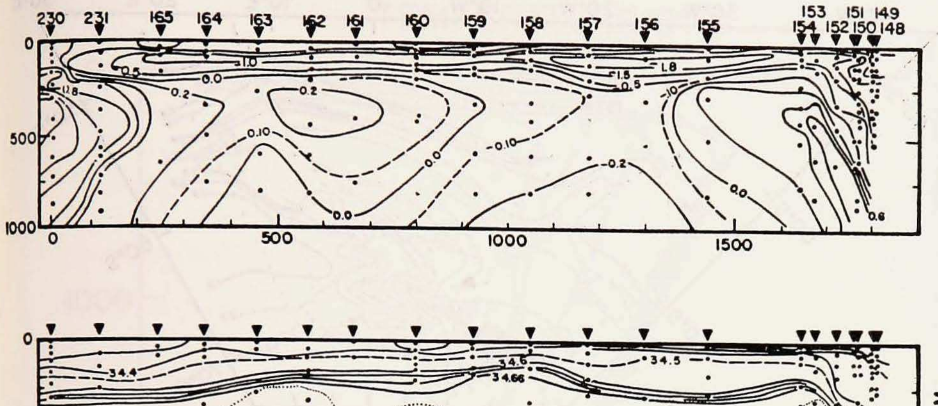

Figure 4. Potential temperature (a), salinity (b), and oxygen (c) approximately along 20W (see Fig. 1) based on CTD temperature and rosette bottle salinity and oxygen data obtained during *Islas Orcadas* cruise 15 stations 148-165, January-February 1978 and stations 230 and 231 from cruise 16, April-May 1978.

stations 123, 124 and 125 (Table 1) reveals the extent of the thermohaline alterations within the full water column.

a. Temperature versus depth (Fig. 6). The three *Glacier* stations from 500 m to 4000 m agree to within 0.1°C. Below 4000 m differences of up to 0.15°C are observed. Between 200 and 500 m station 16 is colder by an average of 0.25°C and yields a heat deficit in this depth interval of 7250 cal/cm^2 relative to stations 11 and 13. Foster and Carmack (1976a, their Fig. 3) indicate the upper 600 m of the station 16 STD profiles have well developed step structure in temperature and salinity versus depth. They offer four possible explanations, including the effects of cabelling and double diffusion.

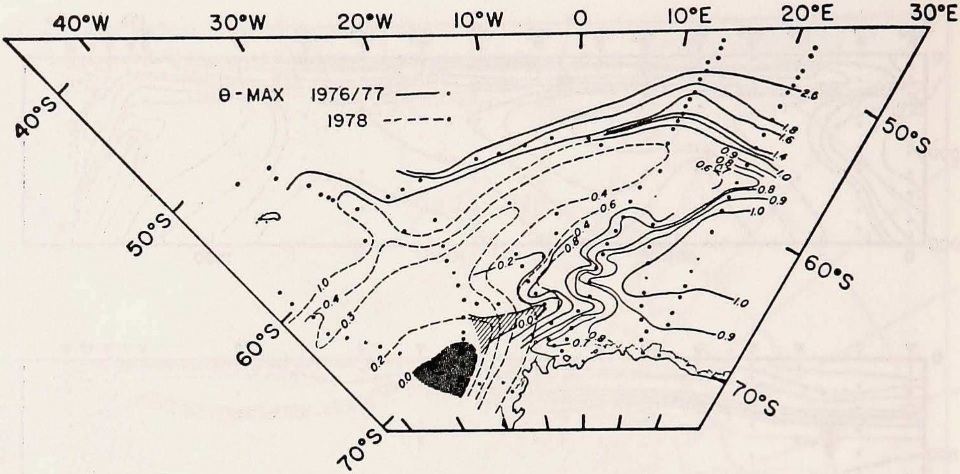

Figure 5. Potential temperature at the temperature maximum core layer within the Weddell Basin for the austral summer of 1976/77 (*Islas Orcadas* cruises 11 and 12, solid isotherms) and for the austral summer 1978 (*Islas Orcadas* cruise 15, dashed isotherms). The 0.4°C isotherm of the 1978 data set is extended eastward to the 10E 1976/77 data. Temperature maximum of less than 0°C is patterned (striped for 1976/77, dots for 1978).

The *Islas Orcadas* stations 123 and 125 are within the "cold spot," station 124 is slightly to the north. The *Glacier* stations 11, 13 and 16 are within the same latitude range as the *Islas Orcadas* stations.

The *Islas Orcadas* stations are significantly colder than the *Glacier* stations from 200 m to 2700 m. The extent of cooling varies with depth and station. Using temperature profiles for a regional selection of stations from cruise 12 and 15 (stations 108, 115, 121, 123-125, 155, 156, 166 and 167) relative to the *Glacier* stations 11, 13 and 16, the mean temperature depression is 0.2°C, with depression of over 0.4°C within the "cold spot." The total areal extent of the WDW cooling is not fully resolved by the data set. Assuming the cooling is primarily in the area covered by the less than 0.4°C isotherm (Fig. 5) south of 65S and that cruise 12 measured only the eastern half of the area, a total area for the cooled WDW amounts to about

Table 1. Position of *Glacier* 1973 and *Islas Orcadas* 1977 hydrographic stations used in water column comparisons (Figs. 5, 6, 7, 8).

Glacier				*Islas Orcadas*			
Station	Date	Latitude °S	Longitude °W	Station	Date	Latitude °S	Longitude °W
11	Jan. 28, 1973	65°34.7′	23°55.2′	123	Feb. 12, 1977	68°20.7′	23°58.7′
13	Jan. 28, 1973	67°14.5′	21°26.3′	124	Feb. 13, 1977	65°30.8′	18°30.6′
16	Jan. 29, 1973	68°50.0′	18°19.5′	125	Feb. 15, 1977	67°26.0′	22°41.2′

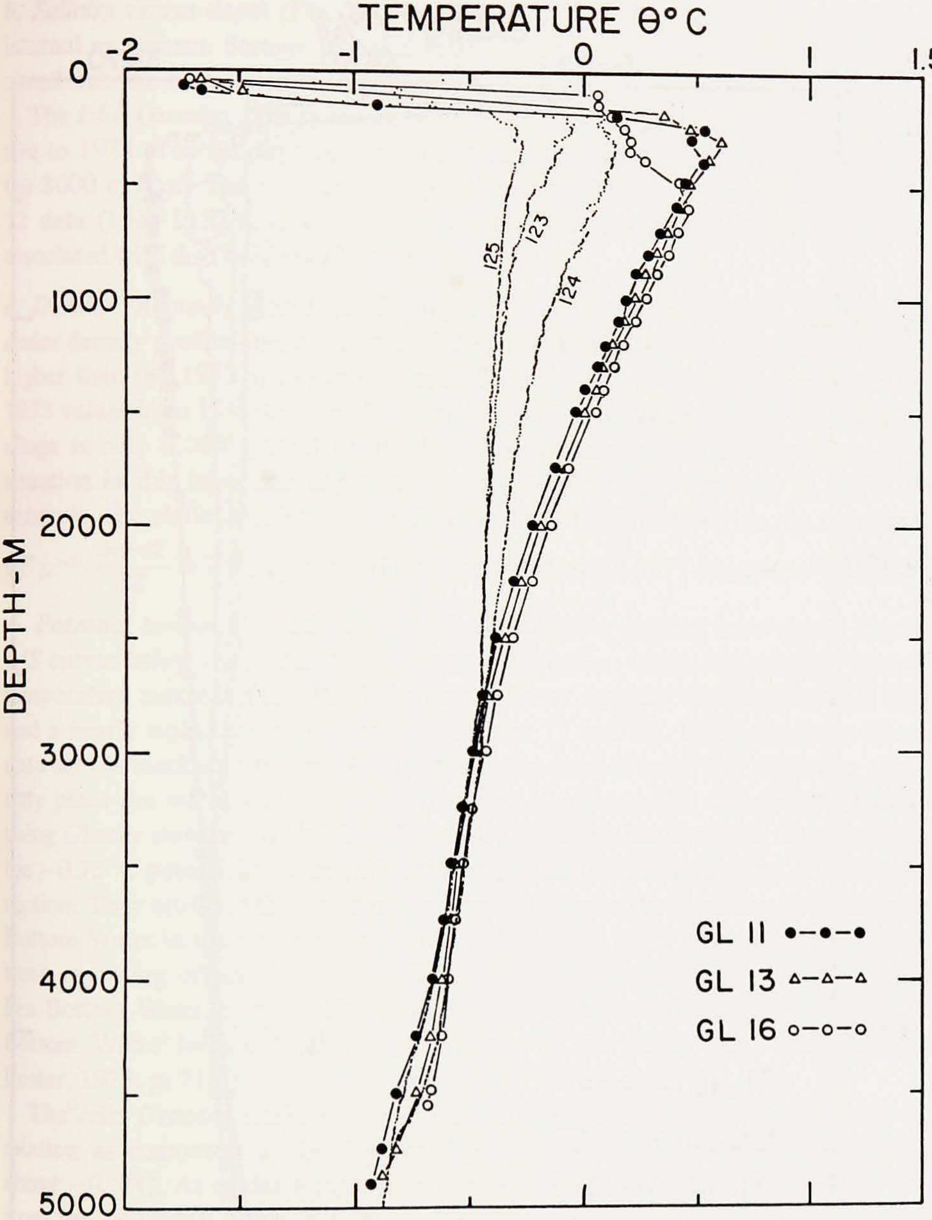

Figure 6. Potential temperature versus depth at standard levels for *Glacier* stations 11, 13 and 16 obtained in 1973, and from the CTD listings for *Islas Orcadas* cruise 12 stations 123, 124 and 125 obtained in 1977.

0.6×10^6 km². Using a mean cooling of 0.2°C for the depth interval 200 to 2700 m yields a total heat deficit of 3×10^{20} calories.

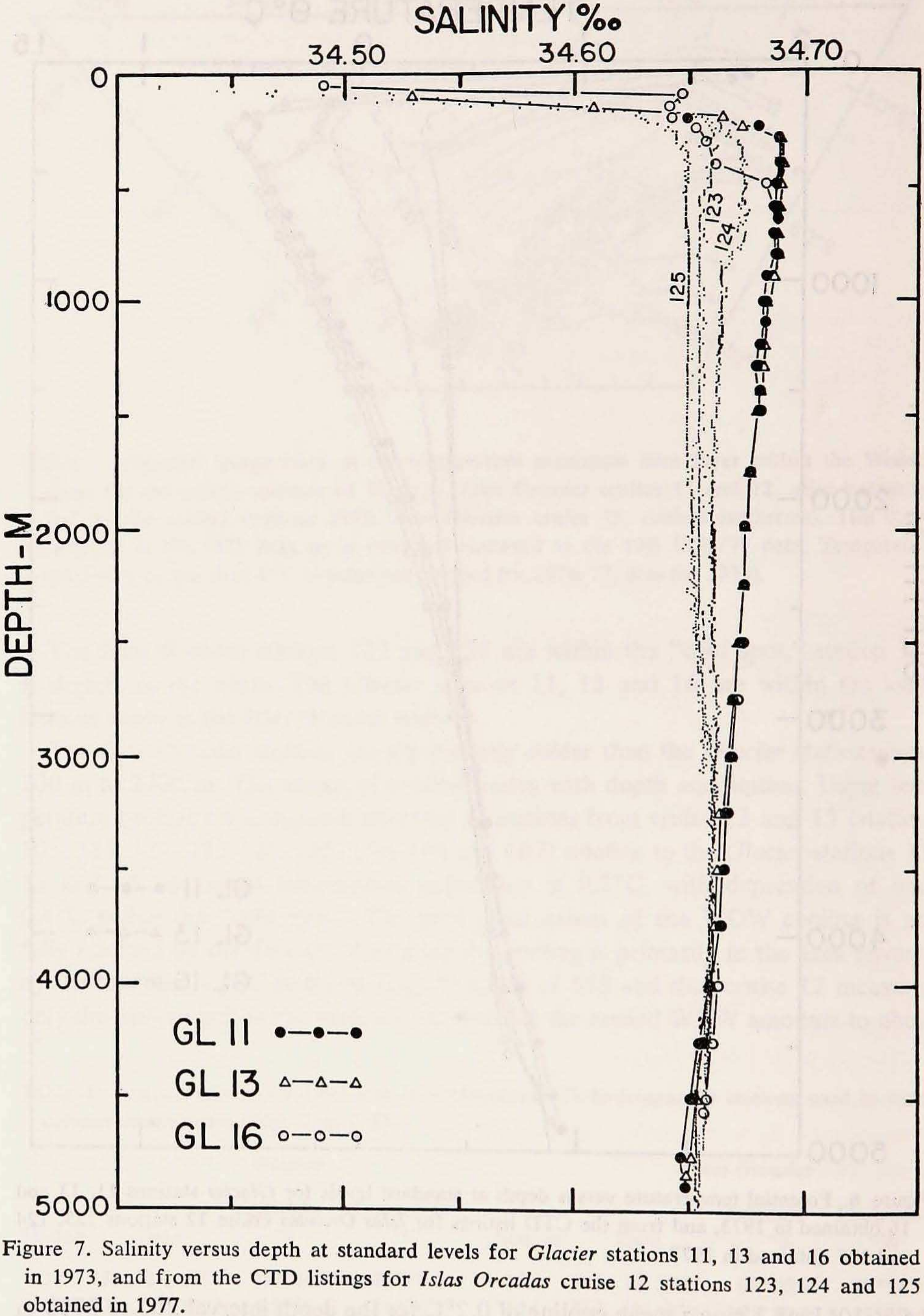

Figure 7. Salinity versus depth at standard levels for *Glacier* stations 11, 13 and 16 obtained in 1973, and from the CTD listings for *Islas Orcadas* cruise 12 stations 123, 124 and 125 obtained in 1977.

b. Salinity versus depth (Fig. 7). The *Glacier* station salinity structure shows good internal agreement. Station 16 has a relative salinity deficit of about 0.015‰ associated with the heat deficit in the 200 to 500 m layer.

The *Islas Orcadas* salinity versus depth plots indicate lower salinity in 1977 relative to 1973. The salinity decrease is significant (above 0.005‰) to approximately the 3000 m level. The average salinity decrease to 3000 m for a selection of cruise 12 data (108, 115, 121, 123-125) is 0.02‰ with extreme differences of 0.03‰ associated with the "cold spot" stations.

c. Density (sigma-θ) versus depth (Fig. 8). Comparison of *Glacier* and *Islas Orcadas* density profiles indicates small differences. Cruise 12, 1977 density is slightly higher than the 1973 situation between 250 m to 600 m, and somewhat less than 1973 values from 1700 to about 3500 m. While the mean density in the 200-3000 m range is only 0.005 greater in 1973, which is essentially insignificant, the 1973 situation in this layer has higher static stability than that observed in 1977. For example, the static stability of the interval from 1500 to 2500 m, as represented by $\rho^{-1} \dfrac{\partial\sigma{-2}}{\partial Z}$ is 2.6 larger in 1973 as compared to the 1977 condition.

d. Potential temperature-salinity (Fig. 9). The *Glacier* stations have nearly identical θ/S curves below the potential temperature maximum: linear θ/S relation from the temperature maximum to −0.4°C; a deeper linear segment with a lower θ/S ratio; and a nearly isohaline benthic layer at stations 11 and 13. Using the same *Glacier* data set Carmack and Foster (1975) discuss the deep abyssal θ/S structure, though they place the warmer discontinuity (between the linear θ/S segments) at −0.5°C using *Glacier* stations 10 and 21. The nearly isohaline benthic layer is found below the −0.75°C potential temperature level only in the northern segment of the *Glacier* section. They attribute the origin of the benthic layer to the injection of Weddell Sea Bottom Water in the western part of the Weddell Sea gyre. They offer two speculations regarding origin of the −0.5°C discontinuity: recirculated "older" Weddell Sea Bottom Water or ". . . it may represent an additional component of Antarctic Bottom Water formed further east, perhaps in the Davis Sea." (Carmack and Foster, 1975, p. 718.)

The *Islas Orcadas* stations 123-125 show a significant displacement of the θ/S relation as compared to the *Glacier* data, particularly at potential temperatures above −0.5°C. At colder temperatures small salinity differences displace the 1977 from the 1973 θ/S curve, but by an amount close to the accuracy of the salinity. However, the lower θ/S ratio revealed in the *Glacier* data below the −0.4°C level is not present in 1977.

3. Hypothesis: Open ocean convective cooling of WDW

The significant alteration of WDW down to the 2700 m depth is most likely a

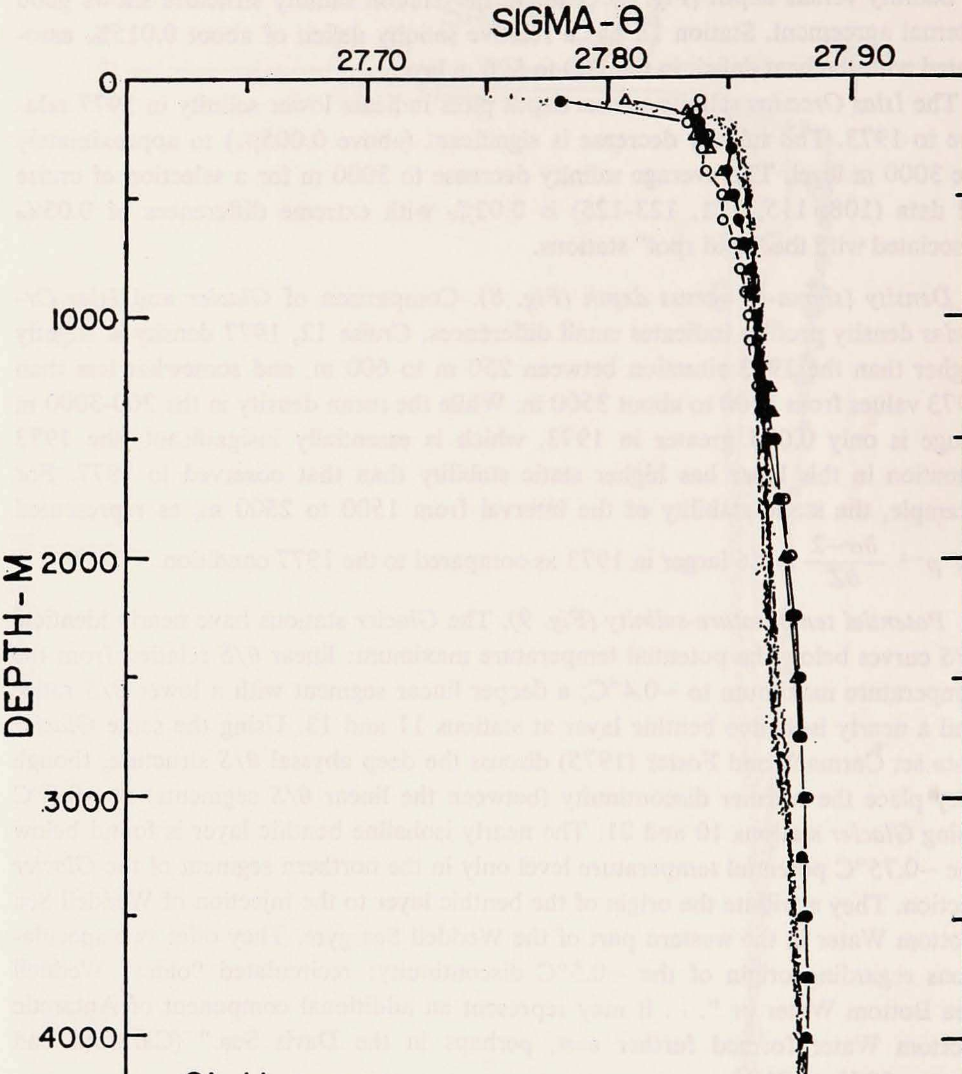

Figure 8. Sigma-θ versus depth at standard levels for *Glacier* stations 11, 13 and 16 obtained in 1973, and from the CTD listings for *Islas Orcadas* cruise 12 stations 123, 124 and 125 obtained in 1977.

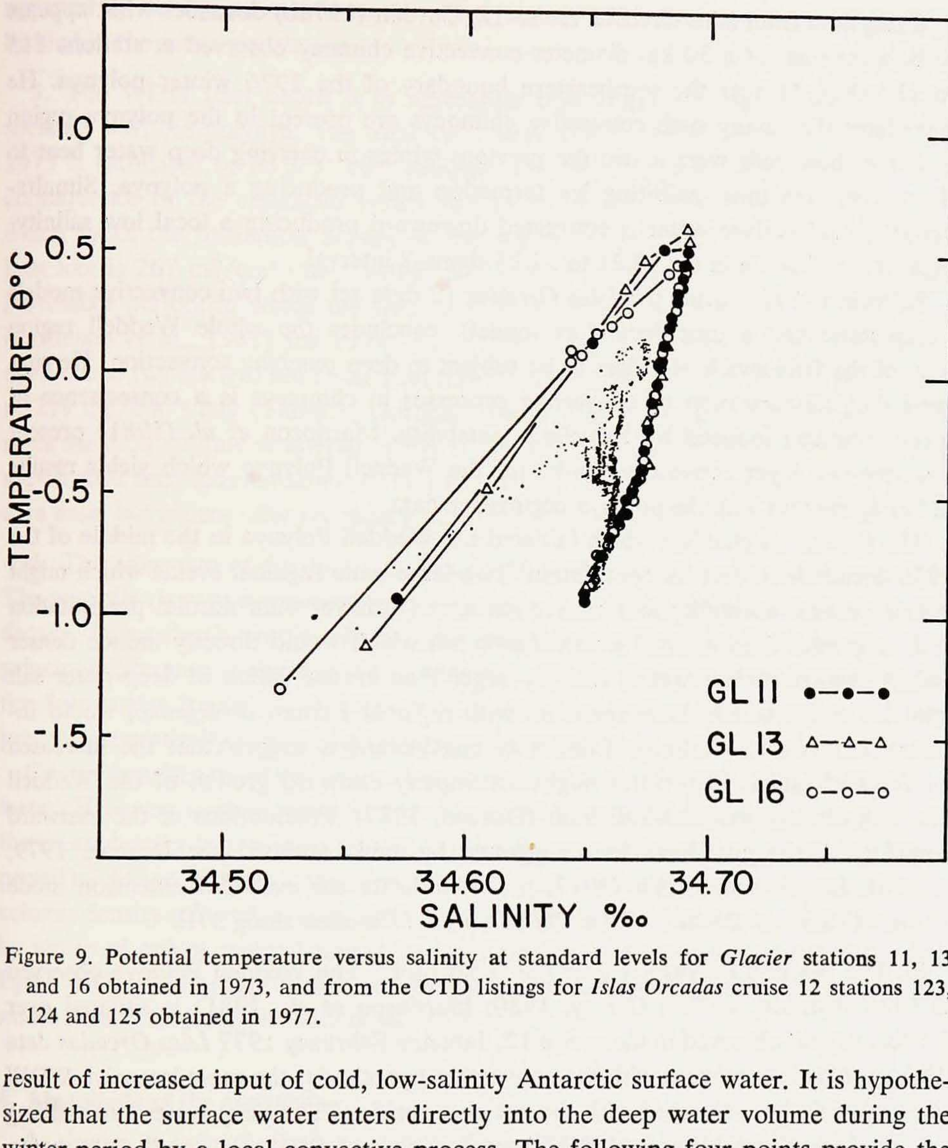

Figure 9. Potential temperature versus salinity at standard levels for *Glacier* stations 11, 13 and 16 obtained in 1973, and from the CTD listings for *Islas Orcadas* cruise 12 stations 123, 124 and 125 obtained in 1977.

result of increased input of cold, low-salinity Antarctic surface water. It is hypothesized that the surface water enters directly into the deep water volume during the winter period by a local convective process. The following four points provide the basis for this hypothesis:

1. The Weddell Polynya is induced by deep convection. During the austral winters of 1974-1976 an 0.2 to 0.3 $\times$ 10^6 km^2 area of ocean, in the same region in which WDW cooling is observed, was revealed by satellite imagery to be nearly or entirely free of sea ice, while the surrounding region had a nearly complete ice cover (Carsey, 1980). This anomalous region is now called the Weddell Polynya (Gordon, 1978b).

Using data from *Islas Orcadas* cruise 12, Gordon (1978b) describes what appears to be a remnant of a 30 km diameter convective chimney observed at stations 115 to 117 (Fig. 1) near the southeastern boundary of the 1976 winter polynya. He speculates that many such convective chimneys are present in the polynya region and that these cells were active the previous winter in carrying deep water heat to the sea surface thus inhibiting ice formation and producing a polynya. Simultaneously, cold surface water is convected downward producing a local low salinity-high oxygen feature in the 37.21 to 37.23 sigma-2 interval.

Killworth (1979) using the *Islas Orcadas* 12 data set with two convective models (quasi-static and a time-dependent model), concludes the whole Weddell region west of the Greenwich Meridian to be subject to deep reaching convection. He suggests that concentration of convective processes in chimneys is a consequence of preconditioning induced by baroclinic instability. Martinson *et al.* (1981) present a simple two layer convective model for the Weddell Polynya which yields results in fair agreement with the polynya occurrence data.

The specific mechanism which initiated the Weddell Polynya in the middle of the 1970 decade is subject for speculation. Two large-scale regional events which might induce vertical instability and convection are: (1) lower than normal precipitation and/or greater than normal sea ice formation which would directly induce denser (saltier) winter surface water; and (2) larger than normal influx of deep-water salt into the region which, in conjunction with regional Ekman divergence, would increase surface-water salinity. Continuity considerations suggest that the increased injection of saltier deep water might accompany eastward growth of the Weddell gyre from its reported 20-30E limit (Deacon, 1937). Fluctuations of the eastward boundary of the gyre have been suggested by many authors (see Deacon, 1979, p. 888). Jacobs and Georgi (1977, p. 70) indicate the eastward extension mode was in existence at the time of the 1974 *Conrad-17* section along 37E.

2. The Weddell Polynya is over the "cold spot." The Weddell Polynya observed in 1976 (Gordon, 1978b; Carsey, 1980; Martinson *et al.*, 1981) is situated over the "cold spot" observed in the cruise 12, January-February 1977 *Islas Orcadas* data (Fig. 5). This is consistent with the convection hypothesis: the most intensive WDW cooling would be associated with the polynya position. The deep water surrounding the "cold spot" would be cooled by lateral mixing of the "cold spot" deep water. As discussed above, the "cold spot" in 1978 is observed farther west than the 1977 position, requiring an annual average 1.3 cm/sec drift rate. A study of satellite microwave images (Carsey, 1980) determines that the polynya feature drifts westward at a rate 1 ± .6 cm/sec. In winter 1977 a weak polynya (short lived and/or only partial clearing of ice) was observed in satellite data near 20-25W and 70S (Carsey, 1980, Table 3) which is in the vicinity of the 1978 "cold spot." Thus, as

required by the hypothesis, both the polynya and "cold spot" move in unison, embedded in the mean circulation.

3. The WDW heat deficit is in agreement with enhanced oceanic heat loss associated with the polynya. The total 1977 heat deficit within the WDW relative to 1973 is estimated to be 3×10^{20} calories. The hypothesis requires that this be a consequence of the enhanced winter sea to air heat transfer within the polynya which lacks the insulation benefit of the sea ice cover. The corresponding daily heat loss is 267 cal/cm^2 • day, within the 0.25×10^6 km^2 polynya for the July to November period, in which the well developed polynya is observed (see Fig. 1 of Martinson *et al.*, 1981) for 1974-1976. This compares favorably to the estimate open ocean (no sea ice) sea to air heat flux for the 60-70S belt for July to November of 219 cal/cm^2 • day (Table 3, Gordon, 1981). This agreement may be fortuitous since the total volume of altered WDW is not well resolved by the data sets and the sea-air heat flux determinations of Gordon (1981) are subject to uncertainties in the data base, but it does offer secondary support for the convection hypothesis.

4. The alteration of the density versus depth profile is consistent with convection. The partially density compensating temperature and salinity alterations yield a 1977 density versus depth profile which differs only slightly from the 1973 profile. The principal difference is that the 1977 density profile is closer to neutral stability for the deep water from 200 to 3000 m. The altered density profile suggests an isopycnal, but vertically homogenizing, process as expected for convection.

Convection begins when the surface density slightly exceeds that of the deep water. Hence as surface water is mixed into the deep water by convective process, the mean density (referred to the sea-surface pressure) of the deep water is not expected to change. However, the deep water is cooled by this process, hence mean water column density referred to deep water pressure levels would be expected to increase, by virtue of colder water being more compressible than warmer water (Reid and Lynn, 1971). After the initial convective event continued sea-air buoyancy flux would lead to increased density of the water column.

4. Magnitude of the convection

As shown in point 3, the heat deficit of the WDW in 1977 relative to 1973 can reasonably be concluded to be a product of enhanced winter heat loss within the polynya. Determination of the volume of water involved with the overturning involves specific knowledge of the sequence of winter events, or of the four stages in a simple two layer, one-dimensional model outlined by Martinson *et al.* (1981). Presumably the low salinity, end of summer surface water is cooled to, or near, the freezing point to initiate convection, after which the saline deep water is exposed to the atmosphere to continue the convection at a somewhat higher temperature. Lateral input of surrounding low-salinity surface water or sea ice might be additional factors which would lower the temperature of the surface water input to deep levels.

In order to determine the minimum volume of surface water input the following calculation is carried out in reference to freezing point surface water. Since the surface water is probably above freezing the actual overturning would be significantly higher to account for the observed cooling. For reference, the maximum volume involved in the overturning would occur if all of it is due to direct cooling of the 1.5×10^{15} m^3 altered WDW. This determination yields an overturning rate of 15.9×10^6 m^3/sec in the three polynya years, or 38.2×10^6 m^3/sec if all the activity occurs during the five polynya months of July to November in each of the three years.

The minimum overturning rate is determined with the aid of the schematic potential temperature-salinity diagram (Fig. 10) showing linear mixing curves between freezing point winter surface water and deep water when the sea level referenced sigma σ-0 of the two water types match and when the 2000 dbar referenced sigma (σ-2) match. The altered WDW of 1977 is about 20% surface water if convection responds to the σ-0 stratification, and about 10% if it responds to the σ-2. The percentage of surface water varies by only 5 percentage points as the deep water type end number is moved along the *Glacier* 1973 curve. The actual mixing curve initiated by the convective process responds to the continuous σ-p density stratification and would fall within the σ-0 and σ-2 mixing lines. Hence an estimate for the amount of winter Antarctic surface water incorporated into the deep water mass varies from 10% to 20%. Using 2500 m for the thickness of effected WDW, the required total thickness of surface water to account for the observed thermohaline alterations ranges from 250 to 500 m. Since the summer surface water mass is only 100 m thick, re-establishment of the surface water mass would be required during the three year polynya period. Naturally, if convective overturning involves cooling of newly exposed deep water, the requirement for low salinity influx is relaxed. Using the estimated total area of 0.6×10^6 km^2 of altered WDW, the total volume of Antarctic surface water convected into the deep water is 1.5×10^{14} to 3.0×10^{14} m^3. This implies an average convection rate of 1.6 to 3.2×10^6 m^3/sec over the entire three years (1974-1976) of the fully developed polynya condition. However, since active convection of the surface would be confined to the July to November period (when the polynya is actually present) for each of the three polynya years, the convection rate during the active state is 3.8 to 7.7×10^6 m^3/sec.

Assuming all of this volume flux is accomplished in half of the polynya area (the other half having the corresponding upward transfer) the lower limit of average vertical velocity (lower since the actual convection may be more confined than the polynya; i.e., within a series of chimneys) is 3 to 6×10^{-3} cm/sec during the five

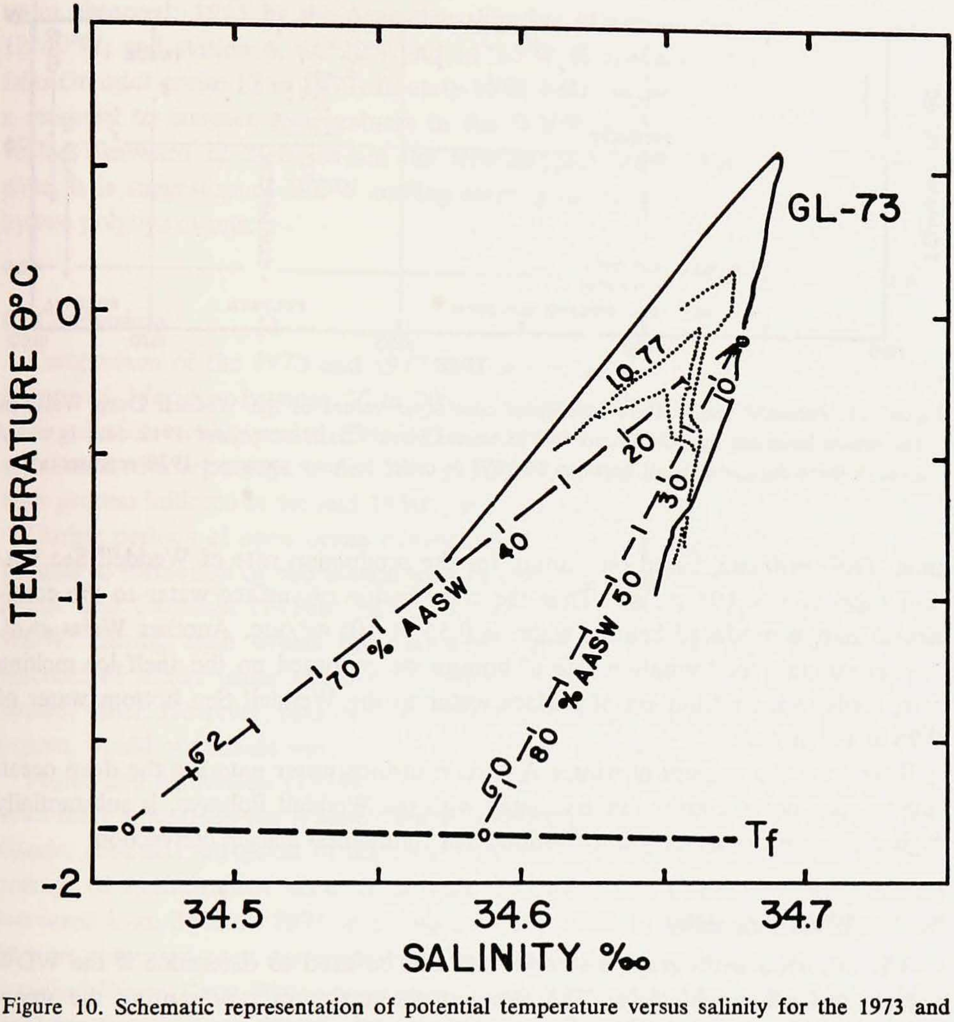

Figure 10. Schematic representation of potential temperature versus salinity for the 1973 and 1977 condition (based on Fig. 8). Two mixing lines are included: for a 1973 WDW end member mixture with winter (freezing point) surface water at the same sigma-0 value (density referenced to sea surface pressure) and for the same sigma-2 value (density referenced to pressure at a depth of 2 km) match. The mixing lines are divided into ten parts to denote percentage of the surface water end member in the mixture.

polynya months. The maximum vertical velocity (if all overturning were cooled WDW) is 30.2×10^{-3} cm/sec.

Weiss *et al.* (1979), using the Foster and Carmack (1976b) recipe for bottom waters derived from the continental margin of the western Weddell Basin, determine the surface water contribution to the Weddell Sea bottom water to be 19% by vol-

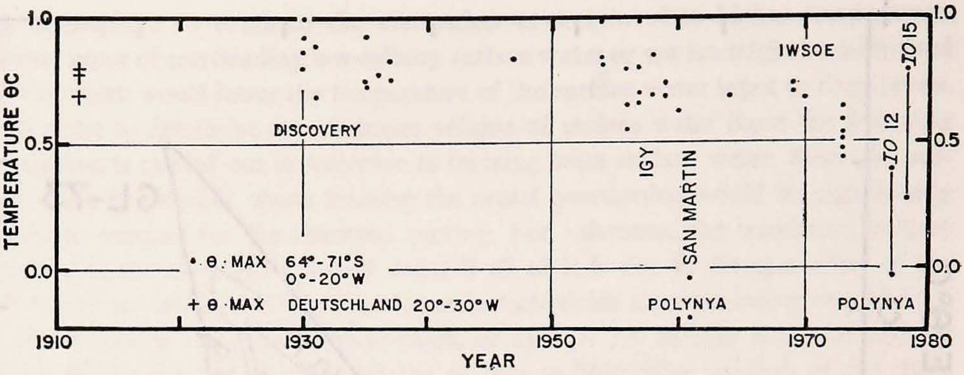

Figure 11. Potential temperature maximum core layer values of the Weddell Deep Water in the region bounded by 0-20W and 64-71S versus year. The *Deutschland* 1912 data is added though these data points fall between 20-30W in order to have some pre-1930 representation.

ume. Their estimate, based on tritium, for the production rate of Weddell Sea bottom water is 3×10^6 m^3/sec. Thus the contribution of surface water to the continental margin produced benthic water is 0.55×10^6 m^3/sec. Another Weiss *et al.* (1979) estimate for formation rate of bottom water, based on the shelf ice melting rate, leads to a contribution of surface water to the Weddell Sea bottom water of 0.95×10^6 m^3/sec.

Therefore, the amount of winter Antarctic surface water entering the deep ocean due to open ocean convection associated with the Weddell Polynya, is substantially higher than surface water lost to Weddell Sea continental margin convection.

5. The WDW time series

The historical hydrographic station data can be used to determine if the WDW experienced other variations. The temperature-maximum representing the upper WDW for the region 0 to 20W and 64 to 71S is presented as a time series (Fig. 11). Clusters of *Discovery* Expeditions data exist for the 1930's, during the 1950's as part of the post-war renewed interest in Antarctica and the IGY episode, and in many austral summers of the 1960's and 1970's as part of the *IWSOE* and the *Islas Orcadas* cruises. The *Deustchland* 1912 data are included, though they fall near 30W, to allow for a pre-1930 contribution to the time series. The time series is far from complete, hence, detection of a "historical cold spot" crossing the region is unlikely.

The potential temperature-maximum core layer values in 1912, the 1930's, 1947, and into the middle 1950's is generally above 0.7°C. Beginning in the mid-1950's the temperature maximum decreases, first to approximately 0.7°C, and then in 1973 (*Glacier* data) to well below 0.7°C. In addition, there are two cold (polynya) episodes observed: 1961 by the Argentine ship *San Martin* (station 5 at 68°39′S and 12°40′W; and station 6, 66°20′S and 11°15′W in January 1961) and during the *Islas Orcadas* cruise 12 in 1977. In early 1978 *Islas Orcadas* cruise 15 data indicate a recovery to warmer temperatures in the 0-20W region as the cold WDW advected westward of 20W. While the hydrographic station time series is incomplete, it is suggestive of WDW cooling starting in the mid-1950's and punctuated by two polynya events.

6. Conclusions

Comparison of the 1973 and 1977 data in the Weddell Deep Water west of the Greenwich Meridian between 65 to 70S indicates significant cooling and freshening during the intervening period. This WDW alteration is produced by deep convection associated with a polynya event. It is possible that the cooling is part of a longer term process initiated in the mid-1950's, as suggested by Figure 11.

During periods of open ocean convection the winter water which normally contributes to formation of continental margin deep reaching plumes, according to the Foster and Carmack (1976b) mixing model, would be diverted directly into the WDW. During such events the continental margin process may involve lesser amounts of winter water component, reducing the formation rate of Weddell Sea bottom water. However, open ocean convection, without the dense shelf water component, would not enable winter water to penetrate the sea floor.

Foster and Middleton (1979), discuss variability in the bottom water characteristics outflowing from the Weddell Sea as revealed by comparison of 1975 and 1976 *Glacier* data sets northwest of the area discussed in this study. They show the percentage of western shelf water (following the Foster and Carmack mixing model) increased from 38% in 1975 to between 54 and 64% in 1976. This change could also be presented as a corresponding decrease in the contribution of open ocean modified Warm Deep Water component. Such a decrease may be associated with the loss of such water directly into the Weddell Deep Water reservoir by convection, as suggested in this study. In this way the bottom water formation at the continental margin receives lesser amounts of modified Warm Deep Water and so is enriched in the shelf water component.

Cooling of the WDW by open ocean convection and the (speculated) accompanying decrease in continental margin produced bottom water may actually increase the Southern Ocean impact on the world ocean abyssal water, since it is the density surfaces associated with the WDW which, by isopycnal processes, spread across the confining ridge system around Antarctica, into the northern ocean (Reid *et al.*, 1977; Gordon, 1978b), whereas the bottom water is confined by these ridges.

Worthington, in 1977 and again in 1981, makes an important point: the present production rates of water masses may not be in equilibrium with the present day abyssal water characteristics. Time scale of the abyssal water residence is of the order of one thousand years, and the replacement rate is small relative to the total volume of abyssal water. Hence a change in formation rates would not become obvious on a world ocean scale for decades to centuries. Remnants of water formed in the past under different climatic conditions might exist. Perhaps small volumes of "anomalous" water, such as the cooled WDW, may abound in the abyssal ocean, and our convenient steady state assumptions might be misleading.

7. The data

1. The *Islas Orcadas* hydrographic data for cruises 11, 12 and 16 are reported by Huber *et al.* (1981a,b,c). The cruise 15 CTD salinity values are not yet processed. The delay is due to faulty conductivity cells which have complicated the editing tasks. Cruise 15 rosette bottle data salinity and oxygen are used in Figure 3. The potential temperature maximum core layer map (Fig. 5) is based on the temperature records of the CTD, which are corrected, and on XBT data obtained between the hydrographic stations. The CTD values were edited and then calibrated to the data obtained by a 24 bottle rosette sampler. Standard deviations of the differences between CTD and rosette thermometers and salinity values are better than 0.005°C and 0.004‰, respectively.

2. The *Glacier* 1973 STD data were collected by the Scripps Oceanographic Institution and standard level values supplied to the author by Prof. T. Foster, who gives the standard deviation between STD and bottle data for temperature and salinity data as 0.006°C and 0.005‰, respectively.

Acknowledgments. The analysis of the ARA *Islas Orcadas* data set is funded by grant DPP 78-24832 from the National Science Foundation Division of Polar Programs. The aid of Bruce Huber and Sarah Rennie in providing pre-data report graphics and listings for the *Islas Orcadas* data is appreciated. Comments by Stan Jacobs and the Journal referees of the study significantly improved the presentation. Lamont Doherty contribution no. 3327.

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Received: *7 August, 1981;* revised: *5 January, 1982.*