JOURNAL OF GEOPHYSICAL RESEARCH, VOL. 95, NO. C10, PAGES 18,177–18,190, OCTOBER 15, 1990

The Weddell-Scotia Confluence in Midwinter

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The southern central Scotia Sea, site of the Weddell-Scotia Confluence where outflowing Weddell Sea waters converge with the eastward flowing waters of the Scotia Sea, was sampled during June–August (austral winter) 1988 with respect to temperature and salinity. Both drogued and ice-mounted drifters, tracked by Argos, were deployed in the region and yielded Lagrangian drift tracks of ice and water motion. The data substantiate past accounts of the region, based upon summer field research, as dominated by eastward flow upon which a complex array of mesoscale features is superimposed. Weddell-Scotia Confluence Water, documented by past summer work in the region and characterized by decreased static stability, was not detected, and the Scotia Front was not well defined. The region was one of intense mixing activity and primarily anticyclonic mesoscale features. Two such features, one an eddy and the other either an eddy or a meander in the Scotia Front, dominated the mesoscale field. With warm cores and containing Polar Front Water, they may have been advected eastward from Drake Passage or may have formed as detached eddies from a sharp northward bend in the Polar Front which typically lies just west of the study region. Several smaller eddies, primarily anticyclonic and some having warm cores, were also detected. There was no evidence of the deep convective mixing which has been hypothesized, on the basis of past summer data, to occur in winter, and vigorous vertical mixing was limited to a 100-m-thick upper mixed layer. Vertical stability in the upper layers was enhanced by low-salinity water derived from melting ice. Temperature-salinity analyses show that winter water in the study region can be derived through isopycnal mixing between waters from the Scotia Sea and waters from the northwestern Weddell Sea. This is in apparent contrast with summer conditions, wherein conditioning of water either through vertical mixing or via lateral mixing on continental margins has been invoked to arrive at the water mass characteristics which typify the Weddell-Scotia Confluence.

1. INTRODUCTION

In this paper we present physical oceanographic results from an austral winter 1988 field study in the southern central Scotia Sea. This study was part of the Antarctic Marine Ecosystem Research in the Ice Edge Zone (AMERIEZ) program, which has addressed the ecosystem associated with the Antarctic marginal ice zones [cf. *Nelson et al.*, 1987, 1989].

Scotia Sea circulation is dominated by eastward to northeastward flow associated with the Antarctic Circumpolar Current (ACC). Hydrographic conditions reflect the presence of water masses from the Bellingshausen and Weddell seas, Bransfield Strait, and the ACC. Water from Bransfield Strait and from the westernmost Weddell Sea may reflect modification by cooling, freshening, and mixing during its passage over the continental shelf surrounding the Antarctic Peninsula. Bellingshausen Sea and ACC waters, which enter the region via Drake Passage, comprise Scotia Sea Water. Weddell Sea Water enters the region from the northwestern Weddell Sea over the South Scotia Ridge, through the gaps west and east of the South Orkney Islands.

The transitional region in which Scotia Sea and Weddell Sea waters converge and mix is called the Weddell-Scotia Confluence (WSC). It is the site of an additional water type, which cannot be derived through mixing among the source waters but must rather originate through conditioning of the inflowing Weddell Sea and Bransfield Strait waters, called Weddell-Scotia Confluence (WSC) Water. Descriptions of this region have been provided by *Deacon and Moorey* [1975], *Deacon and Foster* [1977], *Gordon et al.* [1977], *Patterson and Sievers* [1980], and *Gordon* [1988].

The WSC dominates the oceanography of the southern Scotia Sea in summer. This zonally oriented, elongate feature, which has been observed to vary in width from a few tens of kilometers [*Deacon and Foster*, 1977] to several hundred kilometers [*Gordon et al.*, 1977; *Patterson and Sievers*, 1980], coincides along its axis roughly with the South Scotia Ridge. Bounded over at least part of its extent on the north and south by the Scotia and Weddell fronts, respectively, waters in the WSC tend in summer to be colder and less statically stable than those to the north and south. The locally decreased stability is due to increased upper layer and decreased lower layer salinity.

The weak summer stratification in the WSC has led previous investigators to postulate that the southern Scotia Sea is a site during winter for deep vertical mixing driven by cooling and brine rejection consequent to sea ice formation. The winter marginal ice zone traverses the Scotia Sea zonally, roughly coincident with the WSC (as defined using summer data), from about June to October [*Zwally et al.*, 1983; *Comiso and Zwally*, 1989]. Hence it has generally been assumed that convection might be abetted by the strong sea-air temperature differences which typify winter marginal ice zones in general. *Deacon and Moorey* [1975] speculated

Paper number 90JC01424.
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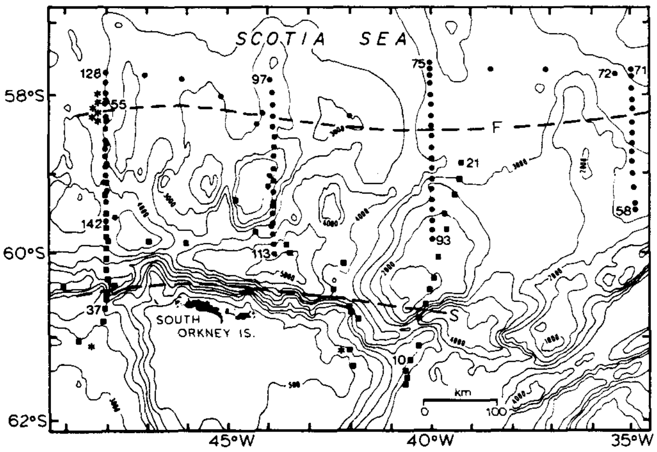

Fig. 1. Geographical location of the study region in the southern Scotia Sea. Locations for CTD casts are indicated by squares (leg 1) and circles (leg 2). Approximate locations for the ice edge at the start and at the finish of the program are shown by the two dashed curves labeled S and F, respectively. Asterisks indicate approximate drifter release sites. Station numbers show the ends of plotted vertical profiles, transects, and θ-S curves which are referred to specifically in other figures. Bathymetry is courtesy of B. A. Huber of Lamont-Doherty Geological Observatory, and depths are in meters.

that the combined high surface salinity and weak stratification might be remnants of vigorous vertical mixing through winter convection. *Patterson and Sievers* [1980] suggested, alternately, that the reduced stratification is due to oceanic lateral boundary layer mixing processes acting on Bransfield Strait and Weddell Sea waters in the continental boundary regions farther west, coupled with admixture of meltwater. There have been no published winter data obtained from the region prior to the study reported in this paper; therefore it has not been possible to estimate the extent to which the summer data represent year-round conditions or the extent to which deep winter convection may in fact occur.

The 1988 field program acquired temperature, salinity, and Lagrangian current data from the southern Scotia Sea region during midwinter. This paper describes and discusses these results. Field methods are described in section 2. The observations are described in section 3 and discussed in section 4.

2. The Field Program

The winter 1988 Scotia Sea field work was carried out in two successive legs from the Universities National Oceanographic Laboratories System (UNOLS) vessel Polar Duke. Leg 1 took place from June 9 to July 5 while the seasonal ice edge was still migrating northward toward its maximum extent. Activities during this leg focused on the immediate vicinity of the ice edge and southward into the multiyear pack ice. Temperature and salinity data were obtained on two relatively rapid transects (less than 72 hours each required for completion), using a shipboard conductivity-temperature-depth profiling system (CTD), nominally along 48°W and 40°W and in a loose grid which fell mostly between these transects (Figure 1). Nominal station spacing along the transects was 15 km. Leg 2 took place from July 18 to August 13 and encompassed the period over which the ice edge attained its maximum northward extent. Leg 2 emphasized water column measurements in and north of the ice edge, and transects having stations spaced nominally 15 km apart were occupied along 48°W, 44°W, 40°W, and 35°W (Figure 1). Individual meridional transects were occupied as rapidly as possible, each in less than 48 hours for leg 2. However, significantly longer periods elapsed between occupations of transects, as indicated by the dates given in the figure legends for each transect.

The CTD observations were carried out using a Sea Bird model SBE 9/11 profiling system. The sampling rate was 12 Hz on all casts. The lowering rate for the underwater unit was approximately 0.5 m/s for the upper 200 m and 1.0 m/s at greater depths. A hardware problem with the CTD limited data acquisition during the first part of leg 1 (including the first transect along 40°W) to depths shallower than 200 m and compromised the quality of salinity data on the later part of this leg. This problem was corrected prior to leg 2. The CTD temperature and salinity sensors were calibrated prior to and following the field work at the Northwest Regional Calibration Center in Bellevue, Washington. There were no significant differences between the two resulting sets of calibration values. The performance of the conductivity sensor was monitored during the field work by carrying out salinity analyses on rosette-derived water samples every third or fourth cast using a Plessey "suitcase" laboratory salinometer. The final data, excluding those salinity data obtained during the later part of leg 1, are accurate to within 0.01°C in temperature and 0.01 psu (practical salinity unit; 1 psu = 1‰ salt by weight) in salinity.

Two Polar Research Laboratories (PRL) (Carpinteria, California) Argos-tracked automatic data acquisition and position (ADAP) units were deployed early in leg 1 on ice floes in the southern portion of the study area (Figure 1). The first of these did not survive long enough to provide signifi-

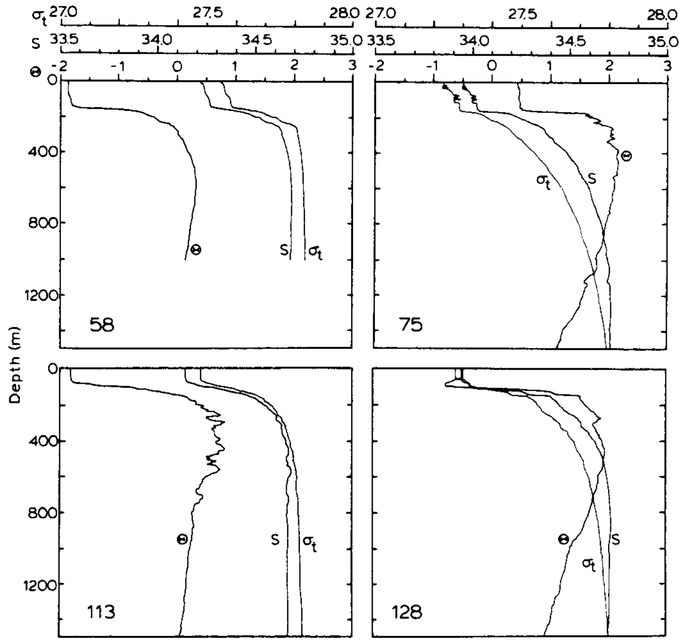

Fig. 2. Representative vertical profiles of potential temperature θ (degrees Celsius), salinity S (psu), and density σ_t. Cast numbers are at the bottom left of each set of profiles, and cast locations are shown on Figure 1.

cant ice drift information. The second survived throughout the field program. Six Argos-tracked, surface-drogued Tristar drift buoys (Technocean, San Diego, California) were deployed at the start of leg 2 along 48°W in the northwest corner of the study area (Figure 1). Three of these provided useful drift tracks, while three ceased to transmit too soon after deployment to provide records. The drift buoys with short survival times were probably destroyed by sea ice, which advanced rapidly because of off-ice winds directly following the buoy deployments.

3. Observed Conditions

3.1. *Hydrographic Conditions*

The regional vertical temperature, salinity, and density characteristics are summarized (Figures 2–11) as representative vertical profiles, as horizontal distributions, and as vertical distributions along transects.

3.1.1. *Horizontal and isopycnal distributions.* The southern Scotia Sea was typified by a relatively cold (−1.8°–1.0°C) upper layer of order 100 m in depth (Figure 2). It was usually, but not always, vertically homogeneous; at some locations more than one layer was present. The upper layer was underlain by a region of strong vertical gradients in temperature, salinity, and density. Below this gradient zone lay a warmer layer comprised of the Circumpolar Deep Water (CDW), having maximum temperatures which exceeded 2.0°C near 400 m depth in the northern part of the study area.

The distributions of temperature and salinity at 50 m depth are assumed to represent the upper mixed layer (Figure 3). South of the −1.75°C isotherm (58°30′–59°00′S) the upper layer was effectively at the freezing point (below −1.80°C) and had salinities of 34.0–34.2 psu. Warmer patches were present at about 59°S on 48°W and 40°W. North of the −1.75°C isotherm, temperature increased irregularly northward along 48°W and 40°W to the relatively warm patches of water (>1.0 and >0.0°C, respectively) at the northern ends of the transects. Temperatures increased less markedly (from the freezing point to more than −1.5°C) along 44°W and 35°W. Salinity decreased only weakly northward, from 34.0–34.2 to 34.0–34.1 psu, and exhibited considerable scatter. Slightly lower salinities (<34 psu) were associated with the warm patches at the northern ends of the 48°W and 40°W transects.

The 400-m level was selected because this was the approximate depth of the CDW temperature maximum (Figure 4). Temperature at this depth increased from $T < 0.5$°C in the south northward to more than 2°C. A cold (less than −0.5°C) patch was present at about 58°30′S on 48°W, and warm patches were observed near the southern end of the 48°W transect and the northern end of the 35°W transect. Warm (>2.0°C) patches at the northern ends of the 48°W and 40°W transects coincided with elevated temperatures and depressed salinities in the upper layer (Figure 3).

These horizontal temperature and salinity distributions can be compared with results from previous work. *Foster and Middleton* [1984, Figure 5] plotted potential temperature at the 50-dbar level. They depict a strong upper layer thermal front about 100 km north of the South Orkney Islands. Our

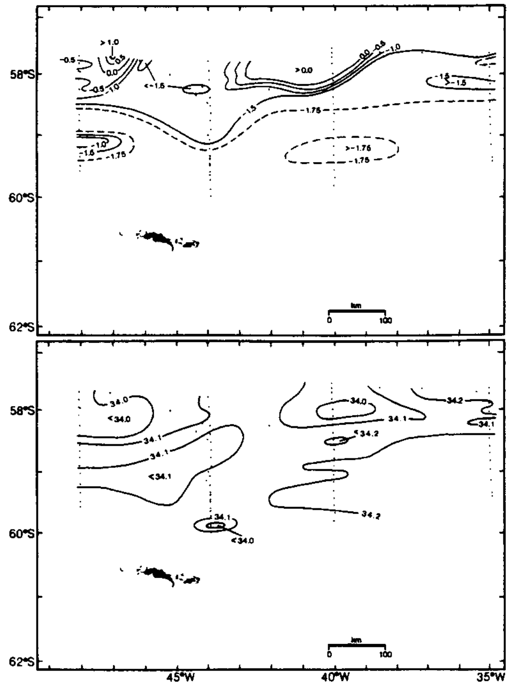

Fig. 3. Distributions of (top) temperature (degrees Celsius) and (bottom) salinity (psu) at 50 m during leg 2, July 18 to August 13, 1988.

winter data do not show such a feature, but they may not have extended far enough south to detect its presence. Foster and Middleton also noted the similarity of horizontal temperature patterns with increasing depth in summer data and suggested that the related features had developed about vertical axes. Both *Gordon et al.* [1977] and *Patterson and Sievers* [1980] have stated, on the basis of summer data, that the WSC is a site of significantly elevated upper layer salinity

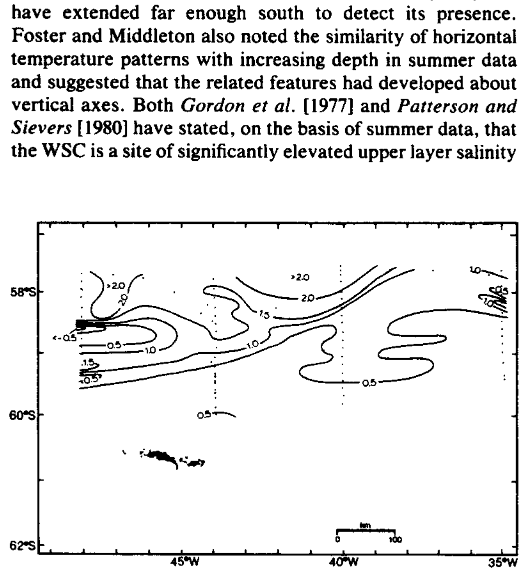

Fig. 4. Distribution of temperature (degrees Celsius) at 400 m during leg 2, July 18 to August 13, 1988.

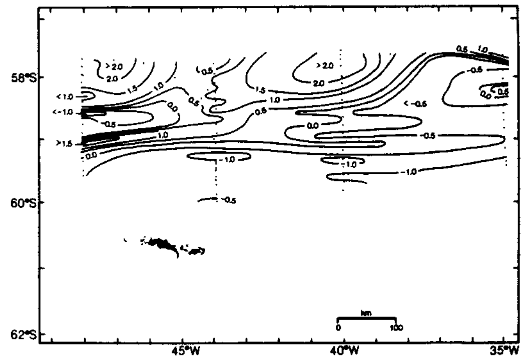

Fig. 5. Distribution of temperature (degrees Celsius) on the σ_t = 27.62 isopycnal surface during leg 2, July 18 to August 13, 1988.

compared with waters to the north and south. They showed surface salinities of order 34.3 psu in the WSC. Our winter data show surface salinities to be less, typically 34.1 psu. Hence the winter data do not show a well-defined salinity increase in the WSC over surrounding areas, as opposed to summer. *Deacon and Foster* [1977] suggested that winter cooling, brine rejection from freezing, and advection of Weddell Sea water into the region raise upper layer salinities in the WSC to more than 34.5 psu while lowering temperatures to the freezing point. While winter water temperatures were indeed near or at the freezing point, observed upper layer salinities were about 0.5 psu lower than predicted by Deacon and Foster. These differences will be discussed below (section 4).

The distribution of temperature is presented on the σ_t = 27.62 isopycnal surface (Figure 5). This surface was selected because it extended throughout the study area, deepening to more than 500 m in the warm (>2°C) patches along 48°W and 40°W to the north and shoaling to the bottom of the upper mixed layer in the south (see Figures 6–11). Temperature on this surface decreased from >2.0°C in the two northern warm patches to less than −1.0°C in the south. Greater complexity was present than for the temperature distribution at 400 m. Meridional gradients were qualitatively greater at 48°W than farther east and appeared to be associated with the two warm-core features along that transect. This pattern is consistent with formation of the frontal and other mesoscale features to the west and decay via lateral mixing processes as they progress eastward with the mean flow. *Patterson and Sievers* [1980] noted, in qualitative agreement with our observation, that the confluence and its associated fronts are quite well defined just off the Antarctic Peninsula where they are first formed, but that they become increasingly confused by mesoscale structures eastward from that point as they are advected downstream. *Foster and Middleton* [1984] noted that identifiable frontal structures were not present in the eastern Scotia Sea, but rather that the fronts had degenerated into a field of mesoscale eddies.

3.1.2. *Vertical distributions.* Vertical distributions of temperature, salinity, and density are presented for the meridional transects (Figures 6–11). These are presented sequentially starting with the westernmost transect, so that the distributions closest to the initial convergence of water

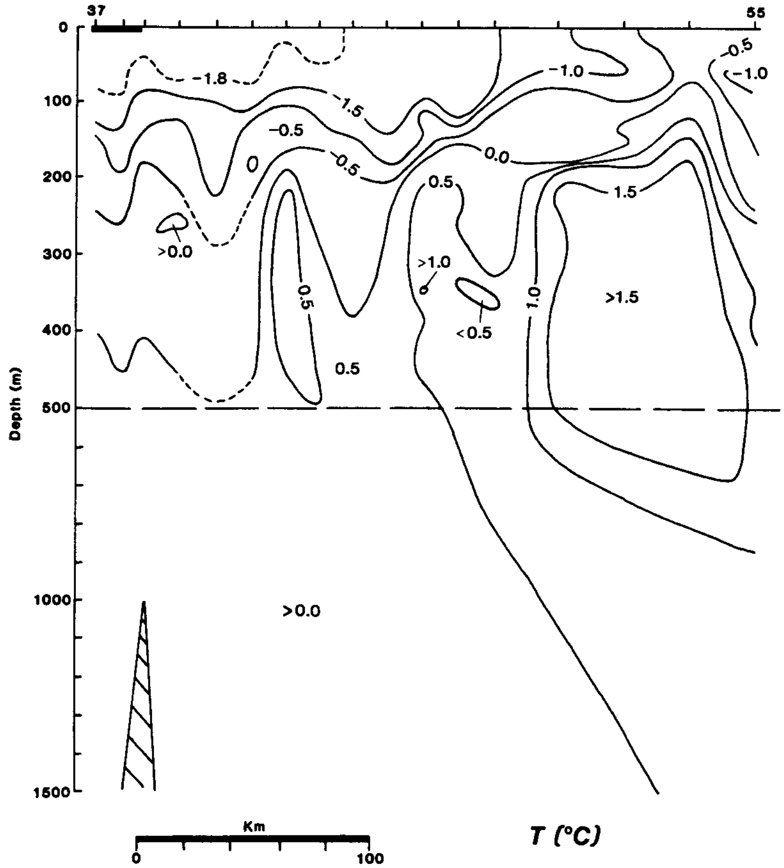

Fig. 6. Distribution of temperature (degrees Celsius) along 48°W during leg 1, July 2–5, 1988. Salinity (and derived density) data were unreliable along this transect because of hardware problems and are not presented (see text). For this and all subsequent transects the heavy line at the surface indicates sea ice cover, and the transect location is indicated in Figure 1.

types are presented first. Occupation of the 48°W transect during both leg 1 and leg 2 allows the comparison of temperatures at two different times. The transects occupied near and along 40°W on legs 1 and 2 did not coincide in space sufficiently to allow meaningful comparison.

Property distributions along the transects showed a cold upper layer, sometimes with low-salinity lenses, and a warm deep layer having maximum temperatures of 1.5°–2.0°C at 400- to 500-m depths in the northern portions. A weak salinity maximum was often present at the warm core, as shown, for example, by the 34.7-psu isohaline on three of the transects. Isopycnals sloped downward to the north in all the transects. This downsloping occurred primarily in the northern halves of the transects, consistent with a stronger baroclinic flow field in the northern than in the southern study region (see the discussion below on circulation, section 3.2). The strongest downward sloping occurred along 48°W and 44°W coincident with the southern edges of the warm-core features there.

There was no indication on the transects of deep-reaching vertical uniformity which may have reflected vertical mixing. The entire region appeared, conversely, to be well stratified in both temperature and salinity except for the relatively thin (compared to the overall water depth) upper mixed layer.

There was no indication that surface density (as σ_t) was elevated within the confluence by as much as 0.6 over Antarctic Zone surface values to the north, as reported by *Gordon et al.* [1977] on the basis of summer data. Rather, the winter surface densities varied from about 27.3 to 27.6 with the lower values typically being associated with the ice edge region. The lower densities tended to occur north of the ice edge, hence in the northern part of the study area. Along 44°W on July 31 to August 2, however, the surface density was 27.4 both north and south of the ice edge, showing little meridional change (Figure 8).

Property distributions along 48°W exhibited the greatest complexity, particularly with respect to temperature, seen

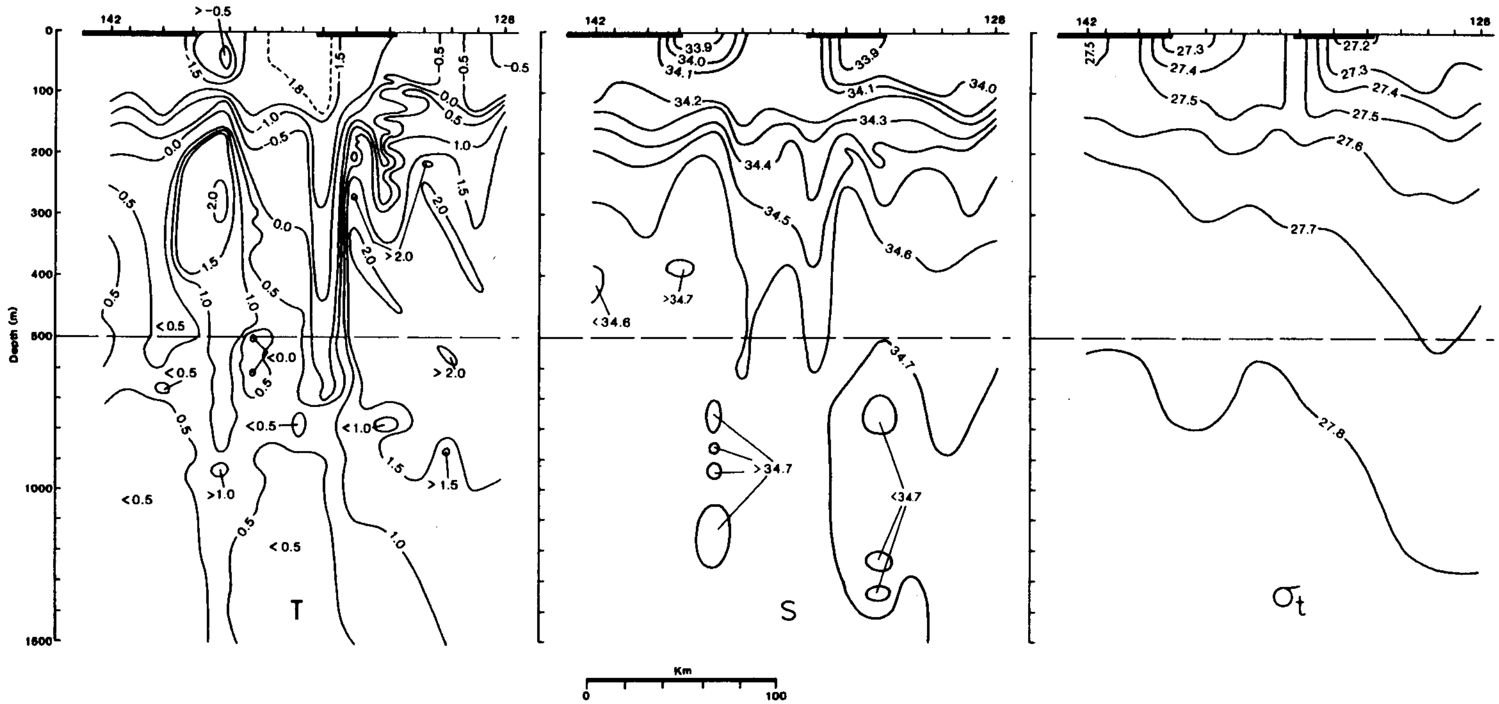

Fig. 7. Distributions of temperature (degrees Celsius), salinity (psu), and density (as σ_t) along 48°W during leg 2, August 10–11, 1988.

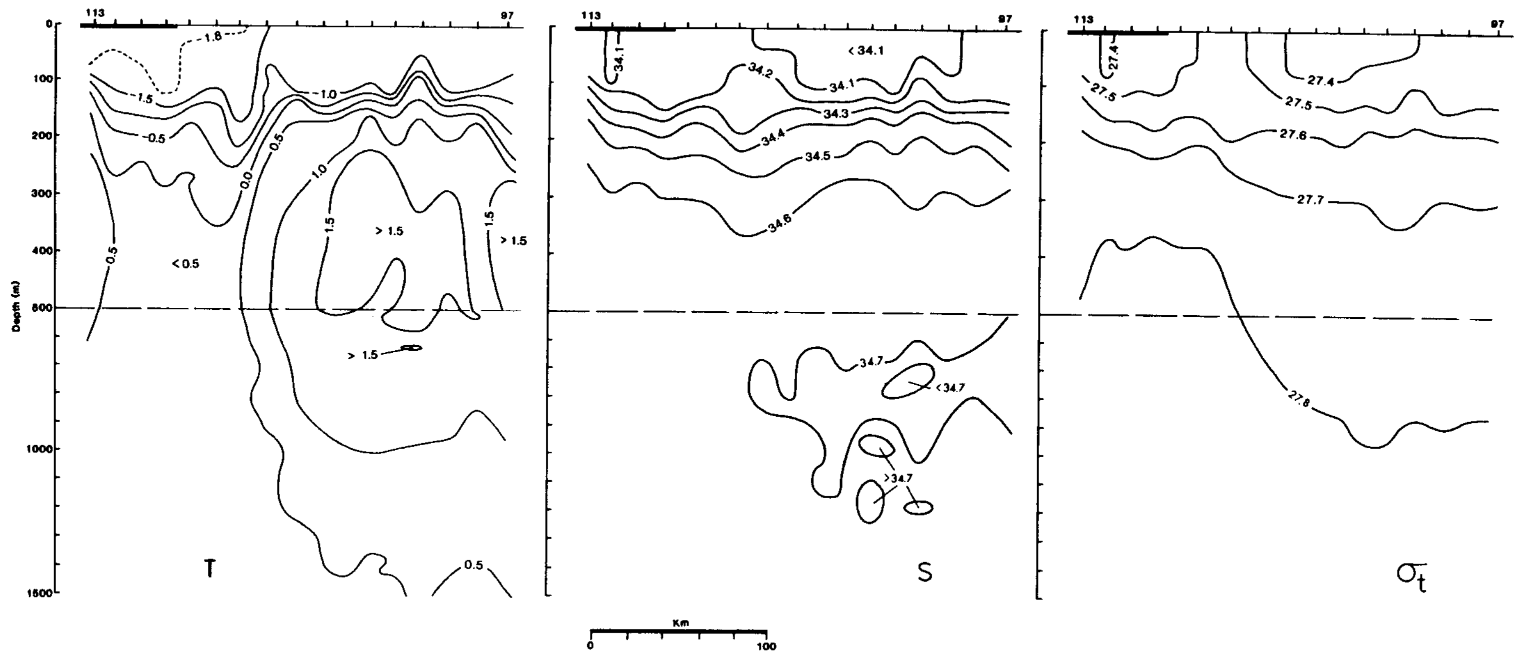

Fig. 8. Distributions of temperature (degrees Celsius), salinity (psu), and density (as σ_t) along 44°W during leg 2, July 31 to August 2, 1988.

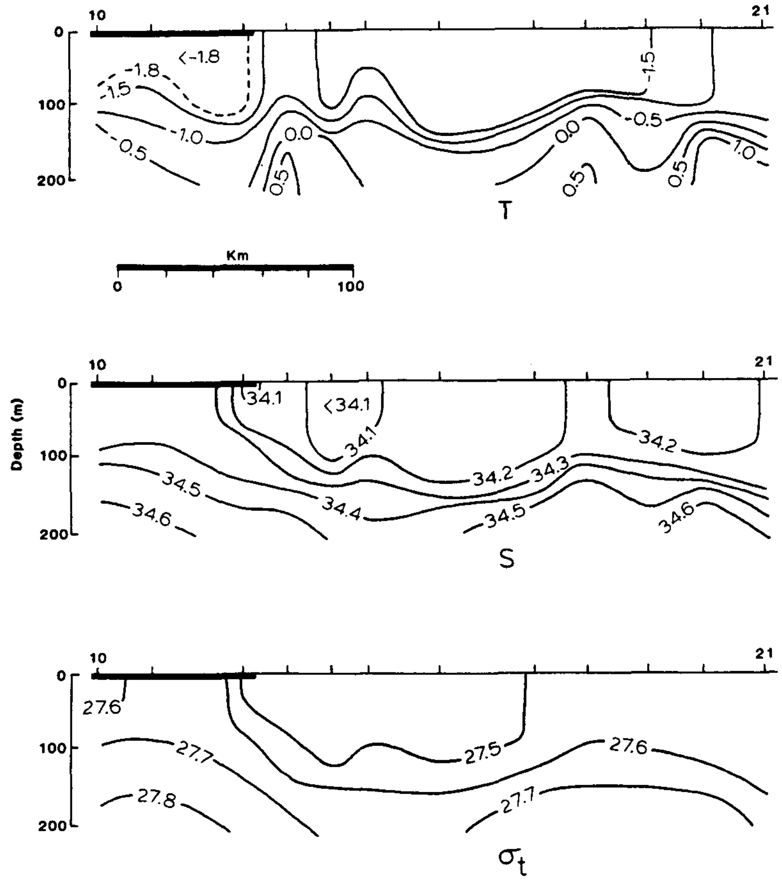

Fig. 9. Distributions of temperature (degrees Celsius), salinity (psu), and density (as σ_t) along 40°W during leg 1, June 16–18, 1988. Data below 200 m were unreliable and are not presented (see text).

on any of the transects (in qualitative agreement with Figure 5). This transect showed clearly, on both crossings, the cold upper layer and the warm CDW core at about 400 m. The warm core was more widespread and had higher temperatures (>2.0°C) in the later than in the earlier transect, and it extended to deeper than the maximum sample depth of 1500 m. The later transect exhibited considerably more complexity than the earlier one.

The literature alludes to a frontal structure, the Scotia Front, which forms the northern boundary of the WSC. *Patterson and Sievers* [1980] arbitrarily chose the 1.5°C isotherm in the CDW to define the Scotia Front. Each of the vertical transects shows a complex region wherein CDW temperatures increased from about 0.5° to 1.5°C. Given the large zonal gaps between our transects, and the regional mesoscale complexity, however, it is not possible to demonstrate continuity of an east-west frontal structure between the transects. The lateral deep temperatures (Figures 4 and 5) suggest that such a structure was present, but the interpretation shown invokes considerable interpolation between the transects. *Foster and Middleton* [1984] were unable to locate the Scotia Front, other than an apparent short segment north of the South Orkney Islands, during summer. They concluded, on the basis of their data from east of about 50°W, that the frontal system had disintegrated into a field of mesoscale eddies which broadened in the zonal sense toward the east.

The interrelations between the upper layer water temperature and the sea ice cover were typical of those observed elsewhere in the marginal ice zones [e.g., *Muench and Schumacher*, 1985]. The ice edge location as observed from the vessel fluctuated widely during the field program in response to local winds. The indicated locations (Figure 1) represent the location that was farthest north as observed from the vessel for each leg. On two occasions (at the northernmost ice edge crossings on August 10–11 and on July 26–28; Figures 7 and 10) the ice edge coincided roughly with the −1.0°C surface isotherm. This is the approximate ambient water temperature above which sea ice melts, and ice which is swept into water warmer than this normally melts within a few days. On other occasions the ice edge was in colder water well south of the −1.0°C isotherm, suggest-

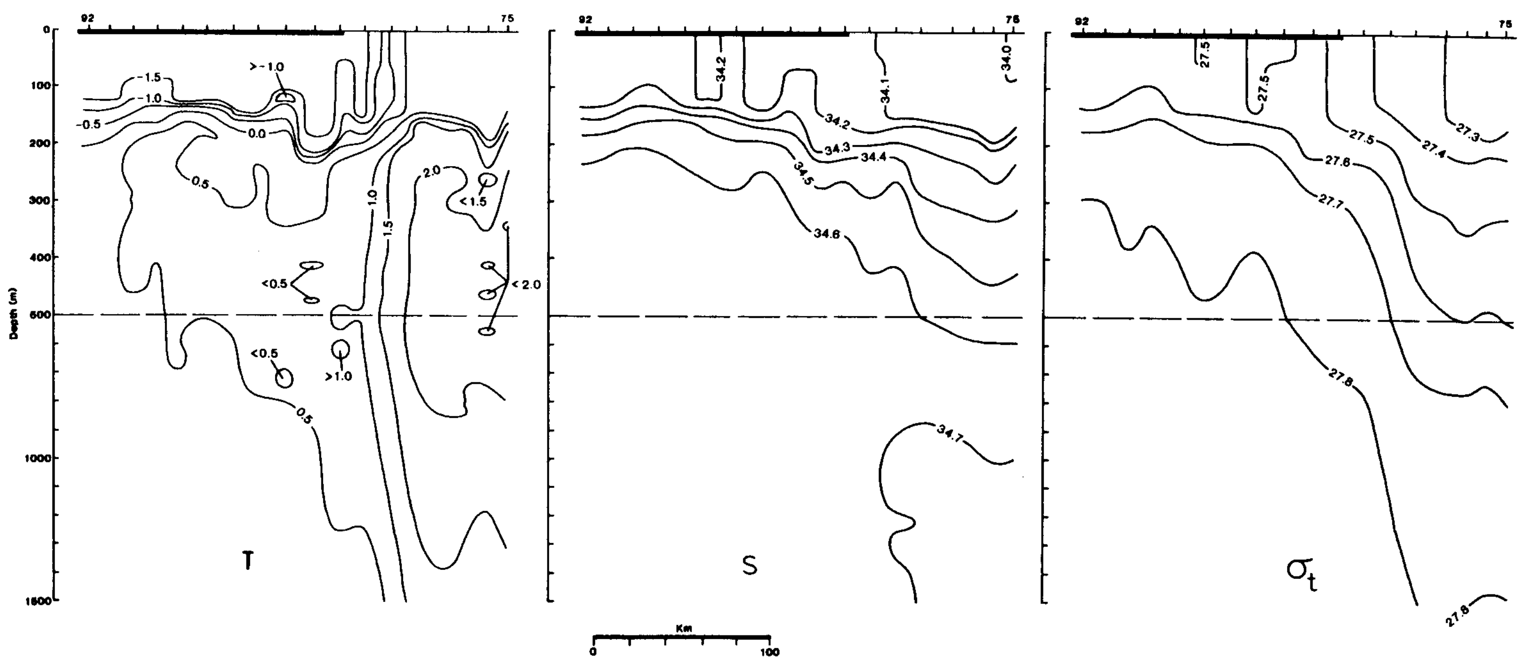

Fig. 10. Distributions of temperature (degrees Celsius), salinity (psu), and density (as σ_t) along 40°W during leg 2, July 26–28, 1988.

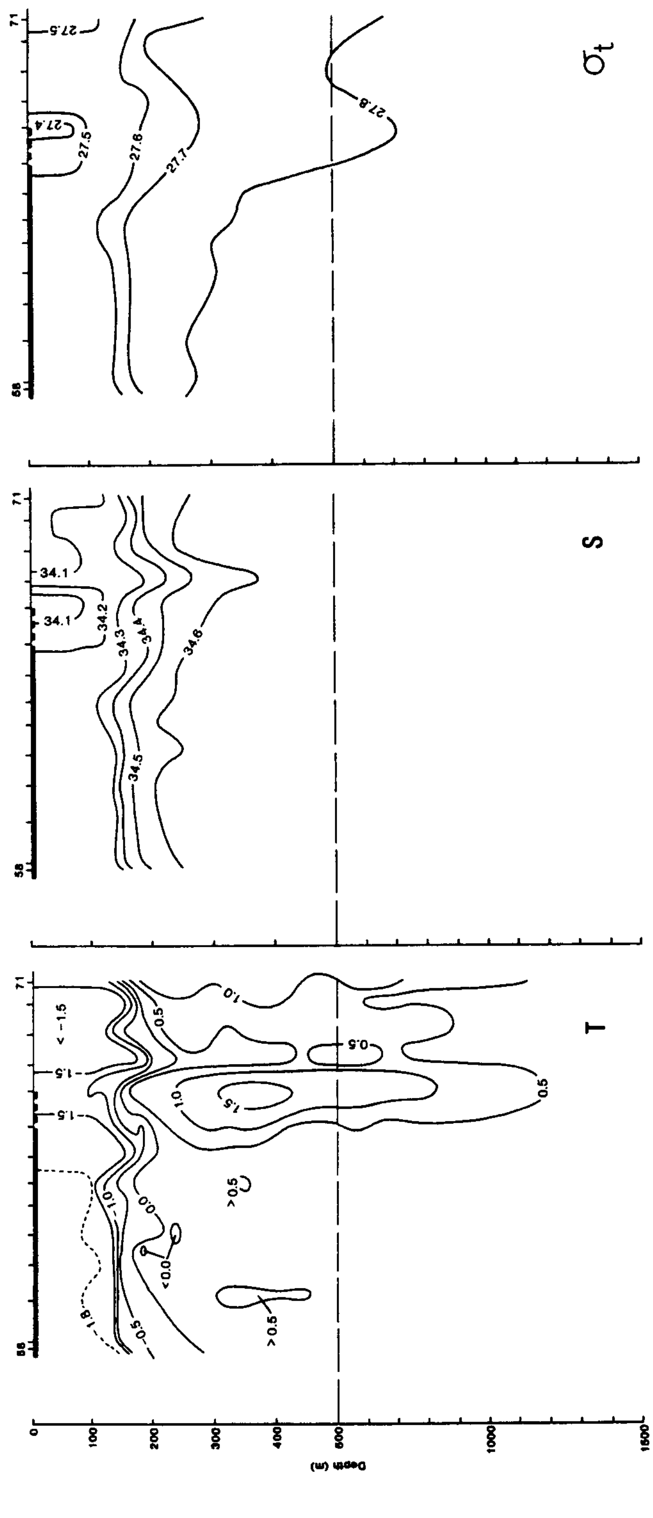

Fig 11. Distributions of temperature (degrees Celsius), salinity (psu), and density (as σ_t) along 35°W during leg 2, July 22–24, 1988.

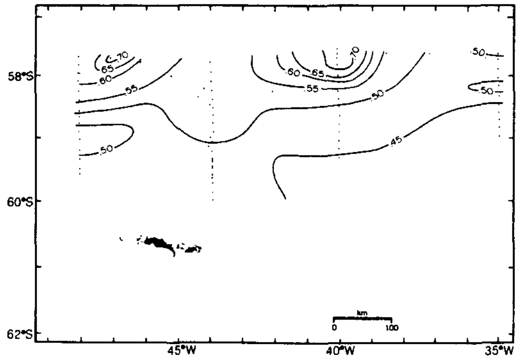

Fig. 12. Dynamic topography of the sea surface relative to the 1500-dbar level during leg 2, July 18 to August 13, 1988. The contour interval is 0.05 dyn m.

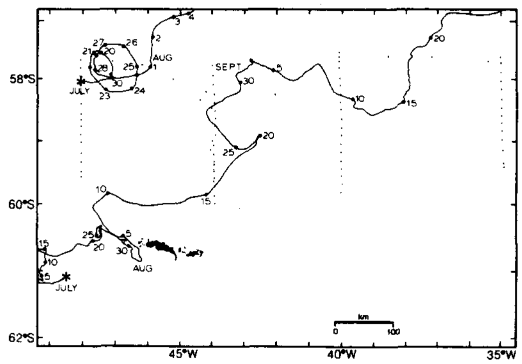

Fig. 13. Drift tracks for the longest record of three obtained from surface-drogued, open water drifters (entrained in the eddy, northwest part of study area) and for an ice-mounted drifter (see text). Circles indicate 1-day intervals for the drogued drifter and 5-day intervals for the ice-mounted drifter. Month and day are given as appropriate at interval points except where space did not permit this. Launch sites for drifters are indicated by asterisks.

ing that the edge had been blown southward from the equilibrium position where input of ice from the south is balanced by melting. Low-salinity lenses due to ice melting at the edge were present on the August 10–11 transect, the June 16–18 transect, and the July 22–24 transect. A low-salinity lens underlaid the ice south of the edge on the July 31 to August 2 transect and may have been the remnant of an earlier ice edge melting event.

3.2. *Circulation*

The baroclinic circulation was derived by constructing dynamic topographies of the sea surface relative to the 500-dbar, 1000-dbar, and 1500-dbar pressure surfaces. Each of these showed the same pattern, with baroclinic current speeds decreasing monotonically with increasing depth so that the surface topography relative to 1500 dbar showed features most clearly (Figure 12). The study region south of the 0.50 dyn m isoline showed little baroclinic flow. The northern part was dominated by a flow which trended slightly north of eastward, as shown by the 0.50 dyn m isoline, and had speeds of order 5–10 cm/s. This flow direction roughly parallels the trends suggested by the horizontal and isopycnal temperature and salinity distributions (Figures 3–5) and is consistent with the regional eastward transport.

The two deep warm patches along 57.5°S (Figures 3–5) were reflected in the baroclinic field as anticyclonic features imbedded within the eastward flow. These extended north beyond the study area and were not completely sampled; however, they appeared to be centered approximately on 40°W and 47°W. They were of order 100 km in horizontal dimension and had associated baroclinic surface current speeds exceeding 40 cm/s for the eastern feature and of nearly 30 cm/s for the western feature. The near-surface-drogued, Argos-tracked drifters which were deployed at the outset of leg 2 provided direct observations of the westernmost of these two features and defined it clearly as an eddy (Figure 13). One of these drifters became entrained within the eddy and executed more than two complete revolutions before exiting it toward the north. This drifted with speeds of 42–54 cm/s from the time it was actually entrained up to its final ejection from the eddy. Movement of the center of the roughly circular trajectory suggests that the eddy was migrating eastward with a speed of about 5 cm/s. Two additional drifters became briefly entrained in the eddy and then exited toward the north after being carried about 90° counterclockwise. Once beyond the influence of the eddy, they drifted toward the north-northeast out of the study region.

Smaller features which resembled eddies were identifiable in the dynamic topography just south of the center of the 48°W transect (an anticyclonic feature suggested by the 0.50 dyn m isoline) and in the northern part of the 35°W transect (an anticyclonic feature suggested by the 0.50 dyn m isolines). The two anticyclonic features had elevated core temperatures (Figures 3 and 4) and may have been warm-core eddies. The widespread occurrence of such eddies is consistent with previously published observations that the eastern Scotia Sea is rich in mesoscale eddies [*Foster and Middleton*, 1984].

Ice floe drift, observed with the single surviving Argos location unit, was as follows (Figure 13). Directly following its launch and prior to August 10 the floe appears to have been trapped in a region of sluggish net flow west of the South Orkney Islands. In late August the floe exited the region and became entrained into the stronger eastward regional currents. The mean drift speed from August 10 to September 15, after the floe had exited the region west of the South Orkneys and while it remained in the study region, was approximately 17 cm/s. Day-to-day drift speeds during this interval varied from about 40 cm/s eastward (August 10–15) to less than 5 cm/s and in a westward direction (August 20–25). A southward deflection of the floe trajectory on September 10–15 at 39°W coincided roughly with the eddy or meanderlike feature which had earlier appeared to be centered on 40°W. Interpretation of the ice floe movement requires caution, since ice drift is influenced by surface winds as well as by currents. Because of icing and hardware problems associated with the shipboard anemometer, no reliable wind observations were available during leg 2. Hence it is not possible to apply a wind correction to the floe drift. We cannot determine in the absence of wind data whether fluctuations in drift speed and direction were responses to circulation features or to local winds.

4. DISCUSSION

One of the potentially significant aspects of the study area lies in the possibility that it is a site for vertical convection and deep ocean ventilation. It was pointed out in the introduction that previous authors working with summer oceanographic data had suggested the WSC to be in part a vertically convective region characterized by reduced stability. Our winter data clearly encompass a region of mixing between Weddell and Scotia Sea waters; however, these data do not reveal deep convection. Only in one instance did the *TS* structure suggest a possible past convective event. In the central part of the final 48°W transect a narrow (less than 20 km, as defined by a single CTD cast), cold (<0°C), saline (0.1 psu more than ambient at the same depth) feature reminiscent of a convective feature was present (Figure 7). This feature was, however, decoupled from the surface by the upper mixed layer and extended down to only about 700-m depths. If it had been in fact a convective feature, it did not appear to be active at the time that the transect was occupied.

Occurrence of deep convection presupposes that stratification is decreased, either as a precurser and/or as a result of the convection. However, stratification, due primarily to salinity, was only slightly less than that measured in summer. Significant weakening of the stratification during winter would require brine enrichment of the upper layers through ice formation. This process would presumably be augmented by the cessation of input of low-salinity ice meltwater from upstream (i.e., the Antarctic Peninsula coastal region and Bransfield Strait). In fact, the regional ice advance appears to occur primarily through wind forcing of the existing ice floes toward the northeast, and the primary site of ice formation and brine rejection lies in the divergent ice field in the Weddell Sea farther south [*Ackley*, 1979, 1984; *Limbert et al.*, 1989]. It is probable that the winter presence of ice in the southern Scotia Sea actually inhibits deep convection by impeding local ice formation and consequent brine rejection. This was also borne out by shipboard ice observations during the 1988 program. Despite a net ice edge advance of as much as 200 km during the field program, the advancing ice was made up primarily of young ice and first-year floes with a few scattered multiyear floes, and little frazil or pancake ice was observed. The "upstream" area in the Weddell Sea is overlain year-round by multiyear pack ice and is not a site for significant new winter ice growth and brine rejection.

The two warm-core cyclonic features along 57.5°S are inherent to the typical regional pattern. There are several possible sources for such mesoscale features in the southern Scotia Sea. *Foster and Middleton* [1984] speculated that similar features observed by them were due either to baroclinic instability or to the development of a plane mixing layer due to turbulent shear flow between the converging water masses farther west. *Klinck* [1985] has documented the eastward passage of rings through Drake Passage at irregular 1.5- to 2-month intervals. Given the observed eastward eddy drift of 5 cm/s, the 400-km spacing between our two observed features would be approximately consistent with the intervals observed by Klinck. *Gordon et al.* [1977] observed a sharp northward bend in the Polar Frontal Zone slightly to the northwest of the observed eddy, and it is possible that segments of the Polar Frontal Zone detach periodically to form eddies which then drift eastward through the Scotia Sea. Vertical temperature inversions from 200 to 400 m in the northern part of the transect, combined with temperatures above 2°C, suggest that these waters were derived from the Polar Frontal Zone [see *Gordon et al.*, 1977]. This origin would be consistent with the apparent presence within the eddy of upper layer temperature inversions which typify Polar Front waters.

Once adrift in the Scotia Sea, the eddies or meanders might behave as planetary waves (see, for example, *LeBlond and Mysak* [1978]). Such waves tend in a beta plane, a reasonable approximation at 60°S, to migrate westward. Westward migration of order 5 cm/s superimposed on an eastward flow of order 10 cm/s would yield the observed eastward migration rate of 5 cm/s. Both a westward migration rate of 5 cm/s and the wavelength (spacing between) of the two observed features are consistent with planetary wave behavior, though a more quantitative analysis is beyond the scope of this paper.

Cyclonic eddies have been hypothesized to precondition water to deep convection by supplying deep water to the sea surface where it can be subjected to cooling and possibly brine enrichment [*Killworth*, 1979]. *Gordon* [1978] detected such a feature in the eastern Weddell Sea. The *MIZEX (Marginal Ice Zone Experiment) '87 Group* [1989] reported seeing deep convective features along the winter Greenland Sea ice edge. These latter features were, however, observed in regions where ice was forming and brine was being rejected into the water column, reducing the static stability interior to the preexisting eddy. The Scotia Sea feature which was identified above by a drogued drifter trajectory as a mesoscale eddy was anticyclonic and did not have reduced stability interior to the eddy. Rather, the upper layer overlying the eddy had increased static stability coincident with lowered salinity. It seems probable that the salinity lowering came about when sea ice, exposed to the higher temperatures overlying the eddy, melted and the meltwater became trapped in the closed eddy circulation. Given that the southern Scotia Sea is a site during winter of net ice melting, rather than of ice formation, this situation would be expected to occur quite commonly. The anticyclonic mesoscale feature farther west, while it did not show a lenslike, low-salinity upper layer, had increased static stability across the main pycnocline at 150–200 m because of increased temperatures and lowered salinities in the upper layer.

Water mass analyses suggest that the water characteristics observed in the southern central Scotia Sea during winter 1988 can be derived from mixing between Weddell Sea and Scotia Sea water types. A family of curves representing the potential temperature-salinity characteristics observed in winter 1988 is presented in Figure 14. Also shown is a potential temperature-salinity envelope derived from data obtained in the northwesternmost Weddell Sea during autumn (March) 1986 as reported by *Husby and Muench* [1988]. These latter data were taken between 48°–53°W and 62°–65°S, an area which we expect is the primary Weddell Sea source region for the WSC. The deepest waters observed in the southern Scotia Sea (1500 m) have θ-S properties identical to Weddell Sea water derived from just above the warm core of the Weddell Warm Water layer. This water type occurs at 300–400 m depth in the Weddell Sea and presumably flows northward into the study region through the gap west of the South Orkney Islands. Warm core water

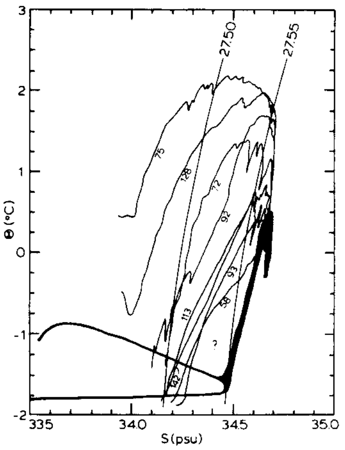

Fig. 14. Potential temperature-salinity curves representing the study region. Numbers adjacent to curves coincide with cast numbers given in Figure 1. The heavy curve is the envelope derived from autumn 1986 data taken in the northwestern Weddell Sea, a source region for the Weddell-Scotia Confluence, by *Husby and Muench* [1988].

from the Weddell Sea is denser than any observed in the southern Scotia Sea, and after entering the Scotia Sea it had probably sunk to its own density level below our 1500-m maximum sampling depth. Water observed in the upper 500–1000 m of the southern Scotia Sea during winter can be derived through isopycnal mixing between the CDW and the Weddell Sea's cold, uppermost layer. The numerous low-temperature layers which were present in the temperature maximum regions are aligned on the θ-S plots along equal potential density curves, consistent with isopycnal interleaving and mixing.

A significant gap exists on the θ-S plot between the southern Scotia Sea and the northwestern Weddell Sea waters, as indicated by a question mark in Figure 14. Presumably, the water types corresponding to this area were situated between our study area and the northwestern Weddell Sea (south of the South Scotia Ridge). The profile from station 58, which we might expect on the basis of the horizontal temperature distributions (Figure 3) to fall close to the Weddell Sea source, trends strongly at middepths toward the northwestern Weddell Sea profiles.

The concept of mixing between upper layer Weddell Sea waters and deeper waters in the Scotia Sea is not a new one. *Deacon and Foster* [1977, Figure 5] prepared a diagram showing lateral mixing paths between the Weddell Sea and the Scotia Sea along about 50°W. Their diagram shows clearly that upper layer Weddell Sea water can be expected to mix laterally with deeper Scotia Sea water. The above water mass analyses demonstrate that such mixing was occurring during winter 1988 and that it was able to account for the observed southern Scotia Sea water characteristics.

Our results do not support previous hypotheses, based upon summer data, which suggest that southern Scotia Sea water is modified through air-sea exchange or continental margin mixing processes, rather than being derived through mixing between Weddell Sea and Scotia Sea waters. In fact, the water type which has been characterized by previous authors as WSC water was not in evidence during winter 1988. We offer three possible explanations. One is that our study region was situated far enough east that WSC water which was presumably formed to the west had been diluted sufficiently through mixing that it was no longer identifiable. A second possibility is that WSC water was present along and south of the axis of the South Scotia Ridge, south of the study region. This seems unlikely, however, since the continual presence of sea ice in that region would have greatly inhibited surface freezing and the resulting convection presumably needed to form WSC water. A final explanation might be simply that there is a large year-to-year variability in the region. This last argument is consistent with the large variation in extent of WSC water which has been reported by the various authors cited above. If in fact the Weddell Sea upper layer water forms a significant proportion of WSC water, as implied by the above discussion, then unusually low upper layer salinities in the Weddell Sea might result in low salinities in the Weddell-Scotia Confluence. Resolution of these questions awaits the acquisition of midwinter data from a broader area, including regions to the west and south.

Acknowledgments. The authors would like to thank A. L. Gordon and B. A. Huber for useful discussions concerning the Weddell-Scotia Confluence region and three anonymous reviewers for suggestions which have improved the manuscript. This work has been carried out under National Science Foundation grants DPP-8420646 and DPP-8715979 to Science Applications International Corporation (R.D.M. and J.T.G.) and DPP-8513098 (D.M.H.) and by the National Marine Fisheries Service of NOAA (D.M.H.).

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(Received February 28, 1990;
revised May 29, 1990;
accepted June 8, 1990.)