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Changes in Antarctic Bottom Water properties in the western South Atlantic in the late 1980s

Victoria J. Coles,[1,2] Michael S. McCartney,[3] Donald B. Olson,[1] and William M. Smethie Jr.[4]

Abstract. Data collected in 1988–1989, as part of the South Atlantic Ventilation Experiment, have been combined with the historical database to study the circulation and water mass variability of the abyssal water in the Argentine Basin. A map of potential temperature at 4000 m used as an indication of geostrophic shear defines a south and western intensified crescent-shaped abyssal recirculation. Within this recirculation, and its northward extension to the Brazil Basin, Antarctic Bottom Water (AABW) properties have undergone two modifications during the 1980s: (1) The water mass cooled (0.05°C) and freshened (0.008 in salinity ratio) on surfaces of constant density. (2) The densest layer of AABW was altered to less dense water through mixing or advection out of the study area. This water mass change does not appear to have affected the flow pattern. Data collected in 1983 and 1988 to the north in the Brazil Basin show penetration of the freshwater mass in the deep western boundary current to between 18°S and 10°S, indicating very rapid propagation of the anomaly from the Argentine Basin into the Brazil Basin as a deep western boundary current. It is suggested that open ocean convective events within the Weddell Sea contributed to the change in AABW documented here.

1. Introduction

The North Atlantic's dominance in renewing the global deep water pool has been recognized since the 1920s [*Merz and Wüst*, 1922; *Reid and Lynn*, 1971]. Thus most studies of decadal timescale variations in the ocean's deep waters have concentrated on the far North Atlantic and changes in the major water mass formed there, the North Atlantic Deep Water (NADW) [*Lazier*, 1988; *Talley and McCartney*, 1982; *Brewer et al.*, 1983; *Schlosser et al.*, 1991; *Dickson et al.*, 1988; *Swift*, 1981; *Read*, 1992]. In the deepest levels of the Labrador Basin immediately downstream of the high-latitude sources for this deep water, salinity ratio changes of the order of 0.02 and temperature changes of order 0.2°C have been observed. Another source of variability to deep water of North Atlantic origin occurs in the South Atlantic through mixing with the underlying abyssal waters formed in the Antarctic. Variability in the southern component, Antarctic Bottom Water (AABW), thus has an effect not only locally in the formation region but also on a global basis through its modification of NADW properties. In this paper we discuss modification of AABW in the Argentine Basin, immediately downstream from its source in the Weddell Sea.

In 1988–1989, the South Atlantic Ventilation Experiment (SAVE) carried out a field program devoted to mapping not only traditional hydrographic properties but also the penetration and distribution of tracer substances in the ocean. The hydrographic measurements, coupled with earlier data sets collected in the southwestern Atlantic, provide the basis for a rough chronology of AABW water mass properties in this region over a 17-year time period. The data show decadal timescale variations in the northward flowing AABW comparable to documented changes in the northern formation regions. Our goal is to examine the magnitude, timescale, and dynamical effects of these fluctuations in the southwestern South Atlantic.

The outline of this paper is as follows. Some background materials on AABW are discussed in section 2, along with a review of the data set utilized in the study. We then present maps of the abyssal geostrophic circulation within the Argentine Basin in section 3.1. The magnitude and spatial scale of the variability are examined in sections 3.3–3.4. Section 4 discusses the spread of the anomaly out of the Argentine Basin, and finally, some possible causes for the variability in the AABW formation region are explored in section 5.

[1]Rosenstiel School of Marine and Atmospheric Science, University of Miami, Miami, Florida.
[2]Also at Horn Point Environmental Laboratory, Cambridge, Maryland.
[3]Woods Hole Oceanographic Institution, Woods Hole, Massachusetts.
[4]Lamont-Doherty Earth Observatory of Columbia University, Palisades, New York.

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2. Background

2.1. Circulation and Water Masses

The Argentine Basin is almost fully enclosed below the 3000-m isobath by the Rio Grande Rise and Falkland Escarpment to the north and south and by South America and the Mid-Atlantic Ridge to the west and east (Figure 1). These topographic features place a strong constraint on the inflow and outflow pathways and circulation of the deep and abyssal layers within the basin. The circulation in the abyssal layer can be summarized as an inflow to the south of the basin, gyral circulation/recirculations in the interior, and an outflow to the north.

At the inflow to the Argentine Basin, AABW generally has its most extreme (i.e., coldest and freshest) identifying characteristics; these are modified through diapycnal and advective mixing as the water mass travels equatorward. Within the ba-

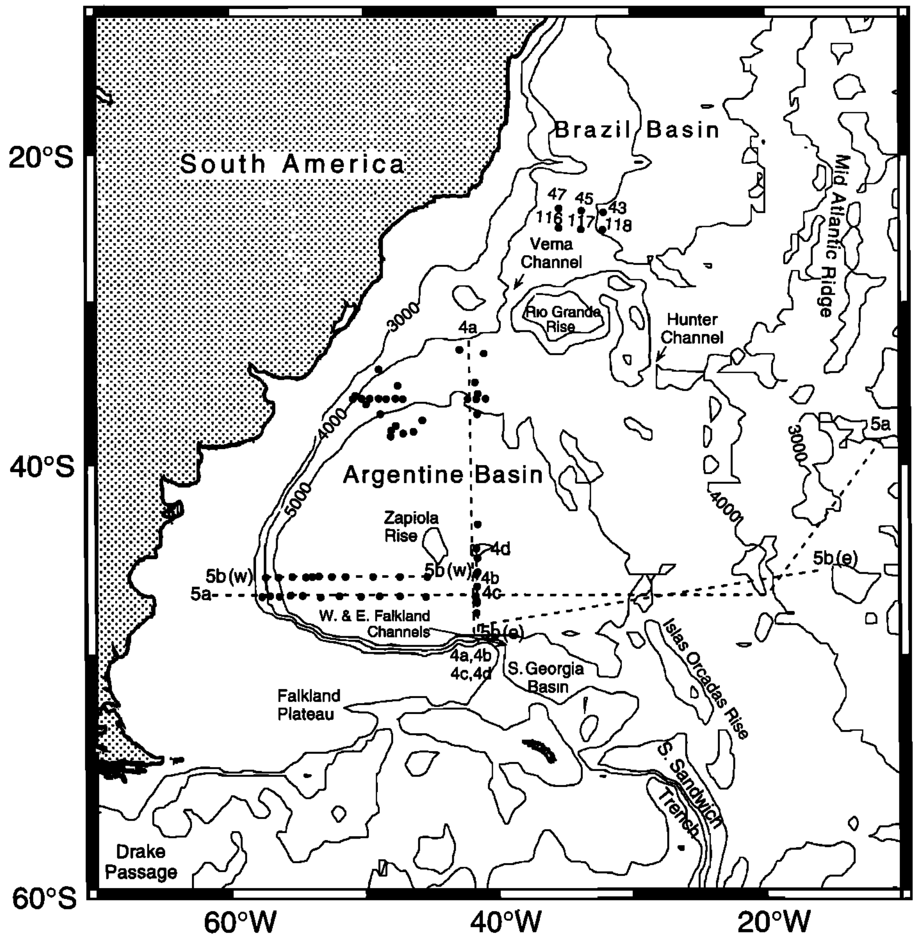

Figure 1. Bathymetric contours and geography of the western South Atlantic. The vertical sections discussed in section 3.2 are indicated with dashed lines and the figure number. The stations used in both the northern and southern property plots discussed in section 3.3 are shown as dots to the north and south of 40°S.

sin, AABW potential temperature ranges from −0.4°C to 0.2°C (all temperatures referred to in the following are potential temperatures), and salinity ratio is fairly uniform, 34.66–34.68. Following *Whitworth et al.* [1991], we use the 0.2°C isotherm to denote the upper horizon of water masses formed within the Weddell Sea, because water colder than 0.2°C does not pass through the Drake Passage [*Whitworth et al.*, 1991]. Water warmer than 0.2°C is designated Lower Circumpolar Deep Water (LCDW); this is the primary water mass in contact with AABW within the Argentine Basin.

Southern inflow to the Argentine Basin is limited by the topography of the Falkland Plateau and Ridge south of 48°S. The East and West Falkland Channels are thought to be the major passages for AABW entering the Argentine Basin from the Georgia Basin [*Georgi*, 1981]. *Whitworth et al.* [1991] speculate that this inflow to the Argentine Basin is regulated by long-period (≈70-day) meandering of the Antarctic Circumpolar Current. The transport into the Argentine Basin is difficult to estimate from the current meter measurements of *Whitworth et al.* [1991], because the net flux is the difference of highly variable eastward and westward flows. Estimates range from 1 to 3 × 10^6 m^3 s^{-1} [*Georgi*, 1981; *Whitworth et al.*, 1991]. An alternative route for inflow to the basin from around the Islas Orcadas Rise has been postulated but never confirmed [*Whitworth et al.*, 1991].

Abyssal water exits the Argentine Basin to the north through channels associated with the Rio Grande Rise. Outflow is partitioned between the Hunter Channel (0.7 × 10^6 m^3 s^{-1} [*Speer and Zenk*, 1993]), the Vema Channel (3.9 × 10^6 m^3 s^{-1} [*Speer and Zenk*, 1993] to 4.1 × 10^6 m^3 s^{-1} [*Hogg et al.*, 1982]), and the Santos Plateau shelf (2.0 × 10^6 m^3 s^{-1} [*Speer and Zenk*, 1993]). *McCartney and Curry* [1993] combine these estimates and show that this net northward flow into the Brazil Basin of 6.8 × 10^6 m^3 s^{-1} warms and upwells as it continues northward.

Less dense than AABW, LCDW from the Drake passage enters the Argentine Basin beneath the Malvinas Current [*Peterson and Whitworth*, 1989] and has a major eastward transport through the southern Argentine Basin with the Antarctic Circumpolar Current. LCDW is characterized by higher salinity and temperature than AABW, reflecting its mixture of waters from northern and southern sources. Within the Argentine Basin, we delineate LCDW as water with temperature between 0.2°C and 1.4°C and with salinity ranging from 34.68 to 34.75.

2.2. Data

In this work we compare the hydrographic data from the 1988–1989 (SAVE) cruises to earlier expeditions in the western South Atlantic (Table 1). Artifacts of the measurement

Table 1. Hydrographic Data Included in This Study

Cruise	Dates	Data Source
CATO VI	Nov.–Dec. 1972	*SIO* [1979]
GEOSECS Leg 6	Dec. 1972	*Bainbridge* [1981]
INDOMED Leg 13	Nov.–Dec. 1978	NDOC
Atlantis II cruise 107, Leg 3	Dec. 1979 to Jan. 1980	*Guerrero et al.* [1982]
Oceanus	Feb.–March 1983	*McCartney and Curry* [1993]
Marathon Leg 9	Nov.–Dec. 1984	*Zemba* [1991]
ABCS	Jan.–Feb. 1986	*Worley et al.* [1986]
ABCS	March–April 1987	*Whitworth et al.* [1988]
SAVE Leg 2	Dec. 1987 to Jan. 1988	*SIO* [1992a]
SAVE Leg 3	Jan. 1988 to March 1988	*SIO* [1992a]
SAVE Leg 4	Dec. 1988 to Jan. 1989	*SIO* [1992b]
SAVE Leg 5	Jan.–March 1989	*SIO* [1992b]
HYDROS Leg 4	March–April 1989	*SIO* [1992b]

CATO, INDOMED, and HYDROS are code names; GEOSECS, Geochemical Ocean Sections Study; ABCS, Abyssal Boundary Current Studies; SAVE, South Atlantic Ventilation Experiment; SIO, Scripps Institution of Oceanography.

processes can appear to indicate significant variability in the property fields over the timescales we are interested in. Salinity analysis methodology in particular has changed considerably in recent decades. SAVE salinity measurements were calibrated according to the new Practical Salinity Scale Standard Seawater (batch P106), relative to KCl standards, whereas data from the 1972 CATO VI expedition, for example (Table 1), were calibrated relative to standard seawater batch P56 on the previous salinity scale. The CATO salinities may be systematically high relative to the SAVE data by 0.001 based on known standard seawater variations alone [*Mantyla*, 1986]. This potential discrepancy is smaller than the observed salinity variability discussed below.

3. Analysis

3.1. Circulation

Some understanding of the deep flow field is central to isolating the effect of water mass variability on the circulation and to determining the pathways through which signals of variability propagate. In this section we consider the abyssal flow, combining new data from the SAVE expedition with historical data (see Table 1) to look in detail at the circulation in just this region. *Olbers et al.* [1992] have published a map of the entire Southern Ocean which shows a temperature pattern within the Argentine Basin similar to the one inferred here but with much coarser resolution due to the map scales involved.

In Figure 2 the temperature at 4000 m is compiled to suggest the geostrophic flow pattern in the abyssal water. Through most of the basin, the water mass sampled at 4000 m is LCDW. The distribution of temperature at 4000 m is a proxy through the thermal wind relation, for the vertical shear between AABW and the overlying LCDW. We have used the information from this map and from abyssal property distributions to draw the flow schematic illustrated in Figure 3.

The tight packing of isotherms to the west in Figure 2 shows an intense deep western boundary current (DWBC) flowing northward along the bathymetry. The origins of this flow can be traced back to the outflow through the Falkland Ridge in the extension of these packed isotherms eastward past 41°W. Between 35°S and 40°S, the DWBC bifurcates, with one branch (0.0°C to 0.3°C) flowing through the Vema Channel into the Brazil Basin and the other branch (0.3°C to 0.7°C) flowing east then south. This eastward flowing branch feeds a C-shaped recirculation cell conformed about the sedimentary Zapiola Ridge, with what appears to be an intensity almost comparable to the DWBC. Previous studies have suggested an interior three-gyre recirculation [*Georgi*, 1981] or a somewhat crescent-shaped recirculation [*Reid et al.*, 1977; *Peterson*, 1992; *Flood and Shor*, 1988] such as we see here. *Reid* [1989] shows no sign of this recirculation in his abyssal maps of adjusted steric height; however, his maps showing the depth of deep isopycnals do include a C-shaped feature in the western basin. The recirculation, with the 0.7°C isotherm at its core and defined by the 0.5°C and 0.3°C isotherms, seems doubly topographically steered. The flow is constrained by continental slopes to the west and south and by the Zapiola Ridge within the arms of the C shape. The northern limit of this recirculation is not topographically steered; it corresponds to the region where the surface Brazil Current turns offshore to extend eastward through the basin. The surface flow in the confluence area is known to be strongly barotropic [*Peterson*, 1992; *Weatherly et al.*, 1993]. It may constrain the northward extent of the abyssal recirculation gyre which one might otherwise expect to extend northward to the Rio Grande Rise following the topography (see Figure 1).

The northern limb of the recirculation flows eastward to about 55°W, although there is some indication in the 0.3°C isotherm that the recirculation continues out to 40°W (this is a relatively data poor area). The southern limb is more sharply defined than in the north, probably because it is tied to the stationary topography, rather than a meandering and highly variable front. Extending eastward to 40°W the southern limb retroflects sharply to the south to join the DWBC just shy of the inflow of AABW from the Falkland Channels. This narrow intense recirculation, here defined by the retroflected 0.3°C and 0.5°C contours and partly by the 0.1°C contour, is well documented in the Abyssal Boundary Current Studies (ABCS) 1986–1987 current meter mooring observations [*Whitworth et al.*, 1991].

Figure 2 also shows the isotherms exiting the Argentine Basin. The 0.5°C and 0.7°C isotherms in the Vema Channel have their origin in the southeast, not from a northward continuation of the DWBC isotherms. Because there is little data coverage between 40°W and 30°W in the northern part of the basin, the detailed structure of the isotherms is not well established. However, the pattern suggests that a significant fraction of the flow out of the Argentine Basin to the north originates

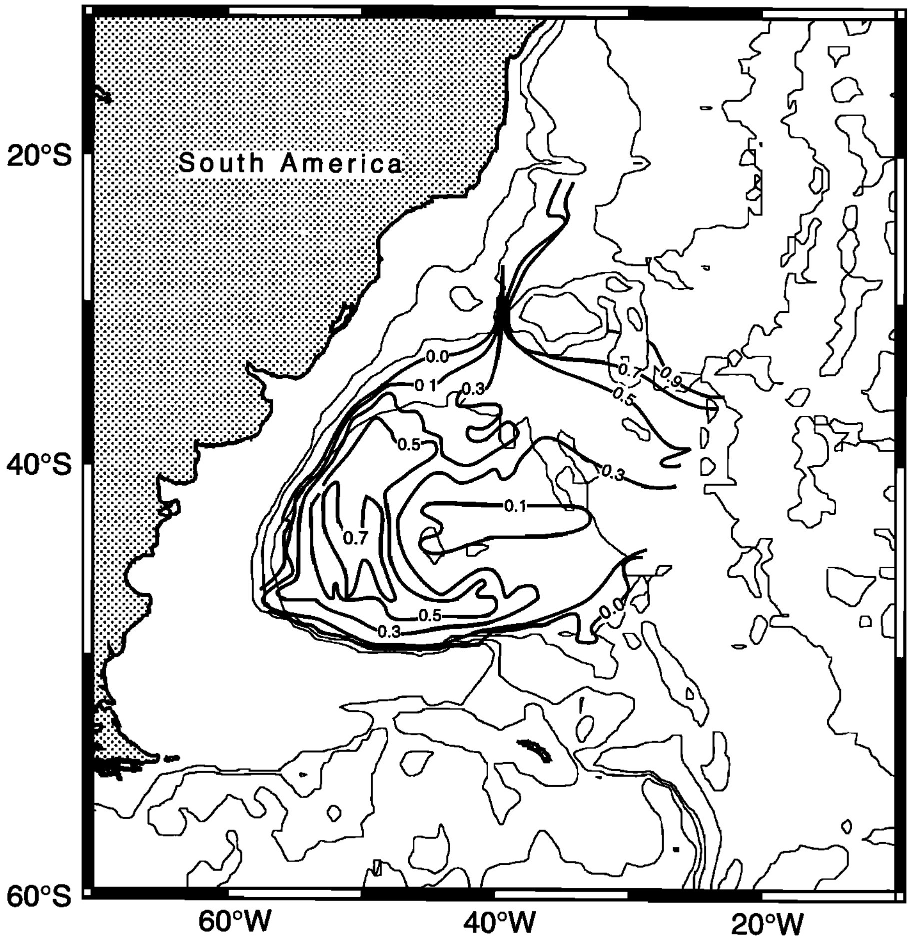

Figure 2. Potential temperature is contoured at 4000 m depth within the Argentine Basin. The contours are a proxy for the deep flow field invoking the thermal wind relation and considering a reference level of no motion above 4000 m.

in the eastern part of the Argentine Basin. This is contrary to prevailing ideas that most of the transport through Vema Channel is supplied from the DWBC. Some support for throughflow originating in the eastern basin comes from a higher-resolution model of the Vema Channel region nested within the Fine Resolution Antarctic Model. In this study, *Wadley and Bigg* [1994] show that flow into the Vema Channel originates in the eastern Argentine Basin rather than from the western boundary current. We have indicated the flow from the eastern basin in Figure 3 with a question mark to suggest the uncertainty in the flowpath given the absence of direct measurements. The flow direction is dependent on the choice of a reference level. Here we are implicitly assuming a reference level of no motion above the 4000-m isopleth. This is based on the large-scale view of southward flowing NADW overlying northward flowing LCDW and AABW; however, the actual reference level for geostrophic computations could be highly spatially variable in this area.

Not evident in Figure 2 is the direct inflow of abyssal water into the central Argentine Basin from the Falkland Channels that has sometimes been suggested by the following observations. Abyssal chlorofluorocarbon measurements [*Warner et al.*, 1990] and hydrographic properties [*Smethie et al.*, 1990] measured during SAVE, as well as flow inferred from mud waves [*Flood and Shor*, 1988] suggest that inflow to the basin bifurcates with some portion of the flow entering the DWBC, and some flow passing fairly directly into the interior around the eastern flank of the Zapiola Ridge (see Figure 3). This inflow forms an anticyclonic recirculation in the bottom property maps similar to the interior closed circulation seen in the 0.1°C isotherm in Figure 2. Anticyclonic flow around the Zapiola Ridge was observed directly using a lowered ADCP during World Ocean Circulation Experiment (WOCE) cruise A11 [*Saunders and King*, 1995]. It seems likely that the inflow feeding this recirculation is not reflected in the map at 4000 m due either to the sporadic nature of the inflow or to the different inflow pathways taken by the topographically steered AABW and the less constrained LCDW.

Advective-diffusive modification of the water mass properties characteristic of the new influx into the southern Argentine Basin occurs along the paths indicated. Abyssal temperature and salinity (not shown, but see maps of deep salinities of *Reid et al.* [1977]) reflect cold fresh tongues of AABW which extend along the western boundary and into the center of the basin, eroding along their pathways through a combination of diapycnal and isopycnal mixing. Downward turbulent diffusion of heat and salt from the overlying layer reduces vertical gradients in water mass properties. Mixing along isopycnal surfaces also occurs, as the new influx of very cold and freshwater circulates and recirculates with interior water which has been

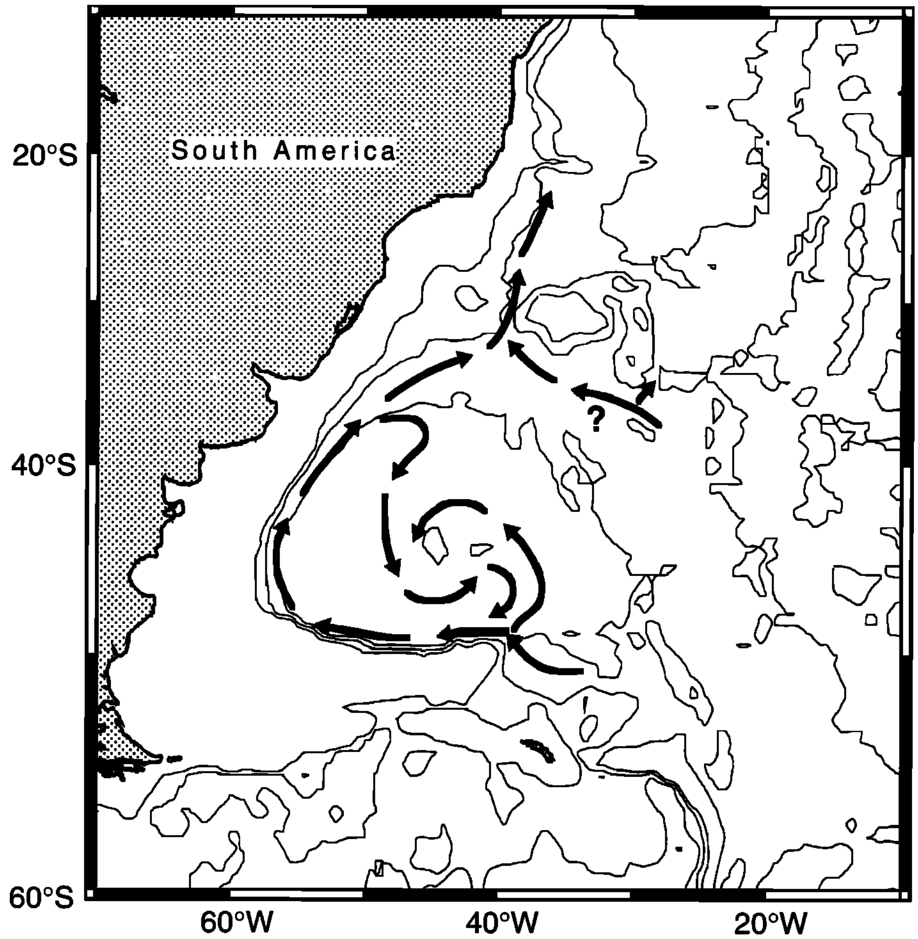

Figure 3. Flow schematic which combines the information from Figure 2 with abyssal tracer and hydrographic measurements to yield a qualitative abyssal flow field. The transports are undetermined, although estimates of the boundary current transport are given by *Whitworth et al.* [1991] and *Peterson* [1992].

resident much longer, and has less extreme properties. This modification of water mass properties has been well documented within the Brazil Basin by *McCartney and Curry* [1993] and farther north by *Whitehead and Worthington* [1982] and can be used in cases where the basin inflow is well constrained, to determine vertical mass fluxes. In the Argentine Basin, inflow is insufficiently well known to attempt this type of calculation, and our data suggest a time dependent water mass distribution, so the relative magnitudes of these types of vertical and horizontal processes can not be determined in this way.

Below, we present evidence of variability in the water mass properties of the abyssal layer unrelated to this gradual equatorward modification. These changes do not directly affect the circulation in Figure 2, as only the AABW layer is affected, and the variability is small relative to the vertical gradients in the property fields. The circulation appears to have remained stable over the 17-year span of the data included in the map, as Figure 2 uses both SAVE and earlier data without contouring inconsistencies. This fact allows us to focus on the pathways such as the DWBC and the interior recirculation discussed above, along which we anticipate that anomalous water masses will travel.

3.2. Vertical Sections

The tilt of the isotherms in the vertical sections shown in Figures 4 and 5 reflects the basic circulation discussed above. The meridional sections at 40°W (Figure 4), pass through the lower recirculation limb of the basinwide gyre. The zonal sections at 47°S (Figure 5) cut across the DWBC which transports much of the AABW inflow, the C-shaped internal recirculation, and the AABW pathway east of the Zapiola Ridge, as discussed in section 3.1.

The ABCS sections in 1986 and 1987 (Figures 4c and 4d) show 400 and 200 m respectively, thick layers of water colder than −0.2°C. These layers are strongly tilted, suggesting the rapid, westward flowing DWBC, with an offshore recirculation shown in Figure 2. The Indomed section in 1978 (Figure 4b) and the SAVE Leg 5 section in 1989 (Figure 4a) have a similar flow structure, despite the time gap between expeditions. This implies little if any shift in the circulation pattern in this area over the time period shown. However, the layer colder than −0.2°C in 1989 has shrunk to a thickness of order 100 m south of 45°S where it can be compared to the earlier data. The layer does thicken slightly to the north, which could suggest that the older thick layer has been pushed into the central basin or that the anomaly has not propagated into the central basin. Given the circulation pattern discussed above, the bullet of cold water at 44°S in the SAVE section could reflect a steady direct inflow of AABW from the Falkland Channels into the central basin or could be some part of the C-shaped recirculation not yet "flushed" out with the anomalous water.

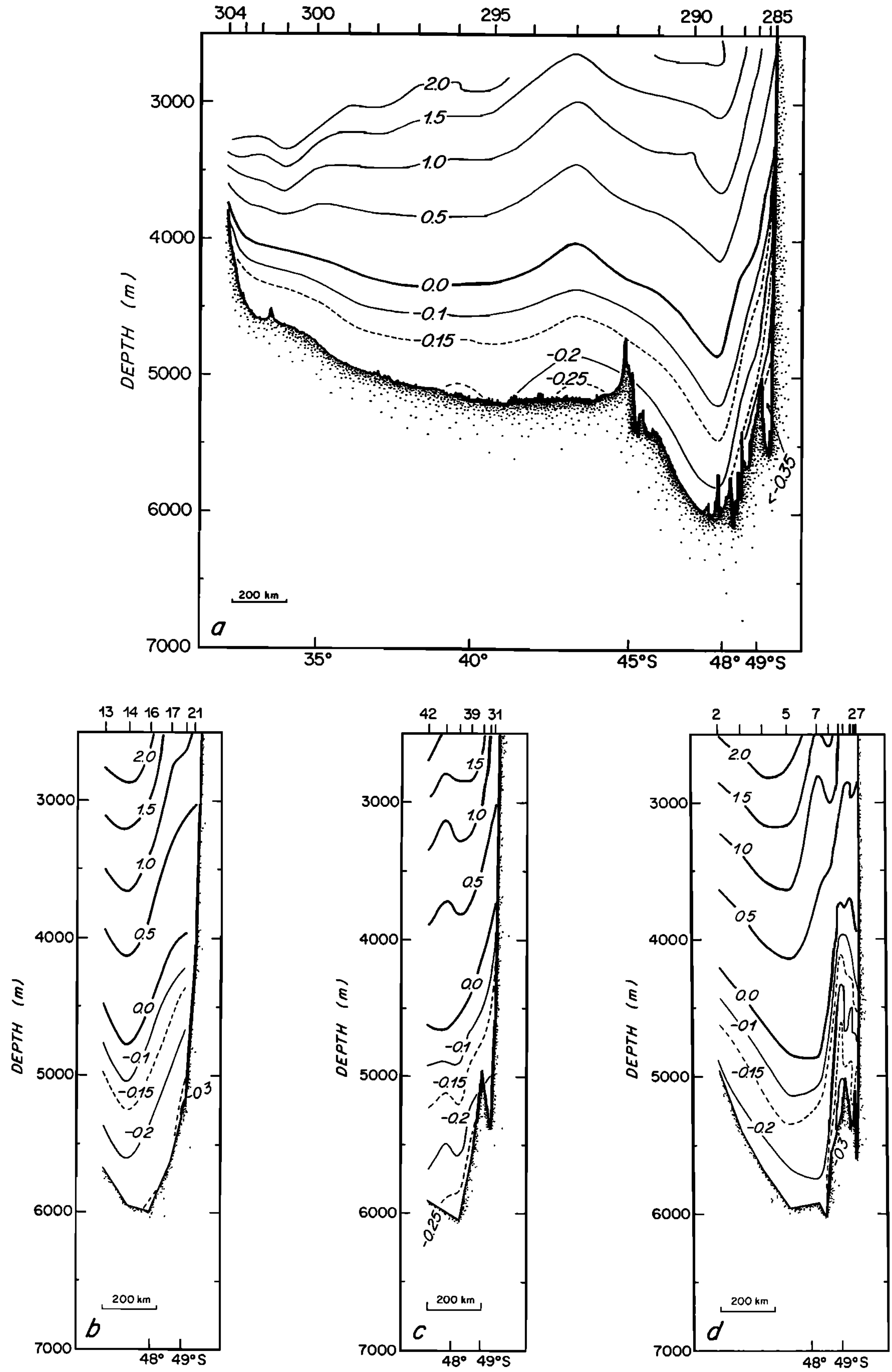

Figure 4. Meridional vertical sections of potential temperature along 40°W originating at 50°S (see Figure 1). These sections show both the similarities in the flow field over the time period spanned and the change in abyssal potential temperature in 1989. (a) SAVE 1989, (b) INDOMED 1978, (c) ABCS 1986, (d) ABCS 1987.

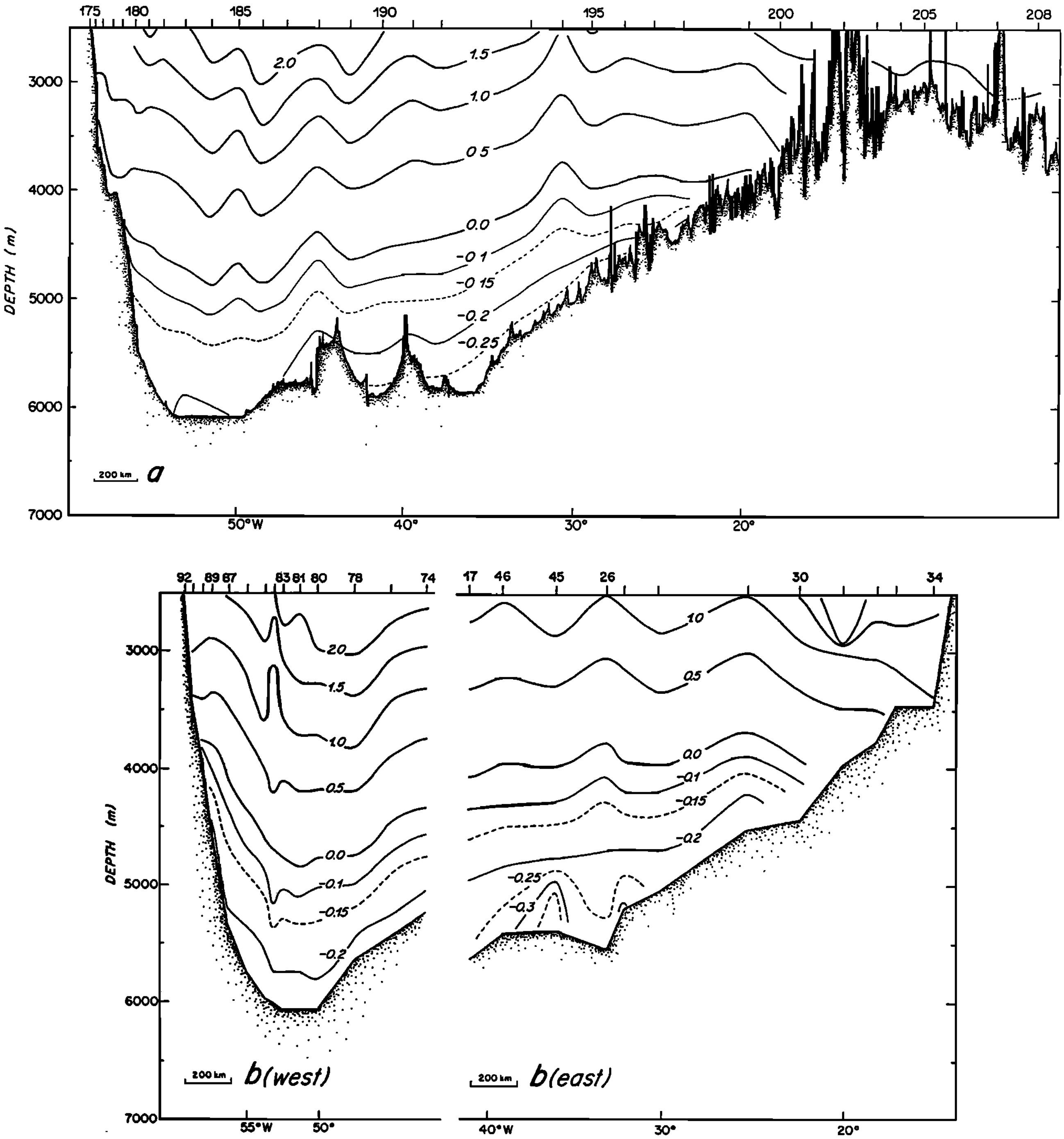

Figure 5. Zonal vertical sections of potential temperature along 47°S, originating at −60°W, (see Figure 1). The flow pattern indicated by the tilt in the isotherms is similar over the three time periods. However, the coverage of the −0.2°C isotherm in particular is much reduced from 1978 to 1988. (a) SAVE 1988, and (b) *Atlantis* cruise 107 1980 (west) and INDOMED 1978 (east).

Figure 5b shows a section at 47°S of potential temperature from the *Atlantis II* cruise 107 data set collected in 1979–1980 (west), and the INDOMED 13 expedition in 1978 (east). A layer of water colder than −0.2°C blankets the topography averaging about 500 m in thickness. The coldest water found in the section occurs at 36°W, evidence for direct abyssal inflow of AABW into the central Argentine Basin. This inflow can also be seen in the SAVE 4 section from 1988 (Figure 5a) between 40°W and 30°W, supporting our view that the essential abyssal circulation within the basin is unaffected by processes controlling the observed water mass variability. Both sections reflect the circulation pattern discussed above; that is, strong tilting of the isotherms up toward the western boundary indicates a rapidly flowing DWBC. Here we assume a reference level of no motion between the NADW and LCDW as discussed in section 3.1. A reverse tilt observed between 50°W and 45°W indicates southward flow in the interior recirculation.

These sections illustrate the spatial extent of the anomalous

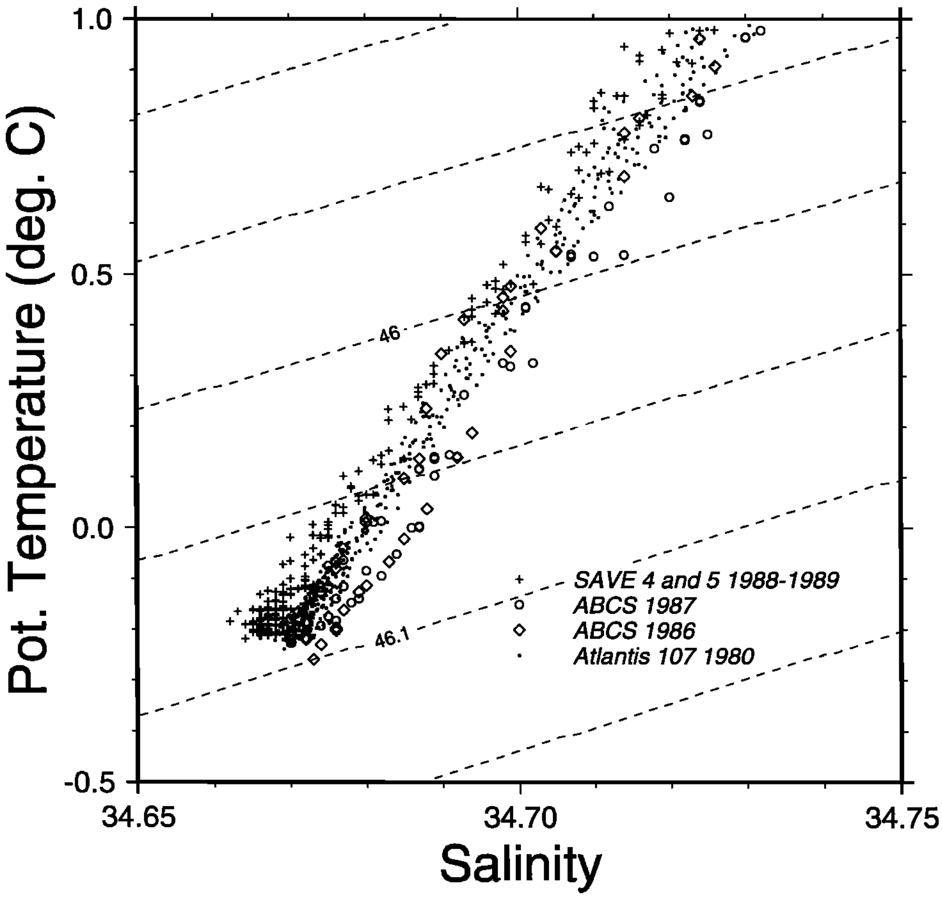

Figure 6. Temperature (in degrees Celsius) versus salinity plots from data in the southern Argentine Basin. The dashed lines represent isopycnal surfaces referenced to 4000 m depth. Two types of change can be seen: (1) cooling and freshening on isopycnal surfaces and (2) loss of the most dense abyssal water (slight warming and freshening of densest water).

cold fresh, yet less dense AABW found in 1988–1989. Whereas a layer of water colder than −0.2°C blankets the topography, with an average thickness of 500 m in Figure 5b, this layer is patchier in Figure 5a. Figure 5a, in particular, suggests a scenario where the C-shaped recirculation is flushed out with less dense waters fairly completely but with little change east of the recirculation. An isolated lens of water is found in the deepest part of the basin, perhaps reflecting a residual amount of the old water. In the east, Figure 5a shows a layer of water colder than −0.2°C, although it is somewhat thinner than in the sections from 1979 to 1980. This could indicate that the eastern part of the basin has not been ventilated with anomalous AABW or, alternatively, that new inflow to the east has returned to the pre-SAVE state.

3.3. Property-Property Plots

Seventeen years of temperature-salinity data are compiled to suggest a chronology of AABW variability. The data are grouped into northern and southern regions for comparison (see Figure 1). AABW properties are altered through mixing and heating as the water mass flows from the southern to the northern parts of the basin, and the bifurcation of the DWBC into a recirculating, and throughflow component suggests two domains. The northern region lies outside the recirculation in the northward extension of the DWBC toward the Brazil Basin, and the southern region lies within the southern limb of the crescent-shaped recirculation.

ABCS and SAVE data south of the southern limb of the recirculation were not used in Figure 6 because the meandering of the Antarctic Circumpolar Current causes large fluctuations in the temperature-salinity distributions over short temporal and spatial scales. This variability swamps any signal of the longer-term trend. Because of this exclusion, the data from the ABCS study in 1986–1987 do not necessarily reflect the inflow characteristics of the AABW but rather some mixture of the inflow and resident water mass.

In the southern region the characteristics of AABW are generally most extreme, i.e., coldest and freshest. Figure 6 shows the temperature-salinity relation for data collected in 1988–1989, 1987, 1986, and 1980. The scatter within the individual data sets shows that the natural variability in the water mass averaged over such a large spatial range is fairly large. However, the data previous to 1988 are consistently warmer and saltier along isopycnal surfaces, than the SAVE data. Between 1987 and 1989, the water mass properties of AABW near the southern inflow region of the Argentine Basin shifted to become somewhat colder and considerably fresher. It is clear from Figure 6 that this type of shift cannot be accomplished by simple diapycnal mixing. It seems unlikely that simply halting inflow into the basin, given the advective-diffusive nature of the system, could generate a local extreme of temperature or salinity. In addition to the change in properties on a common isopycnal surface, the densest water seen in 1988–

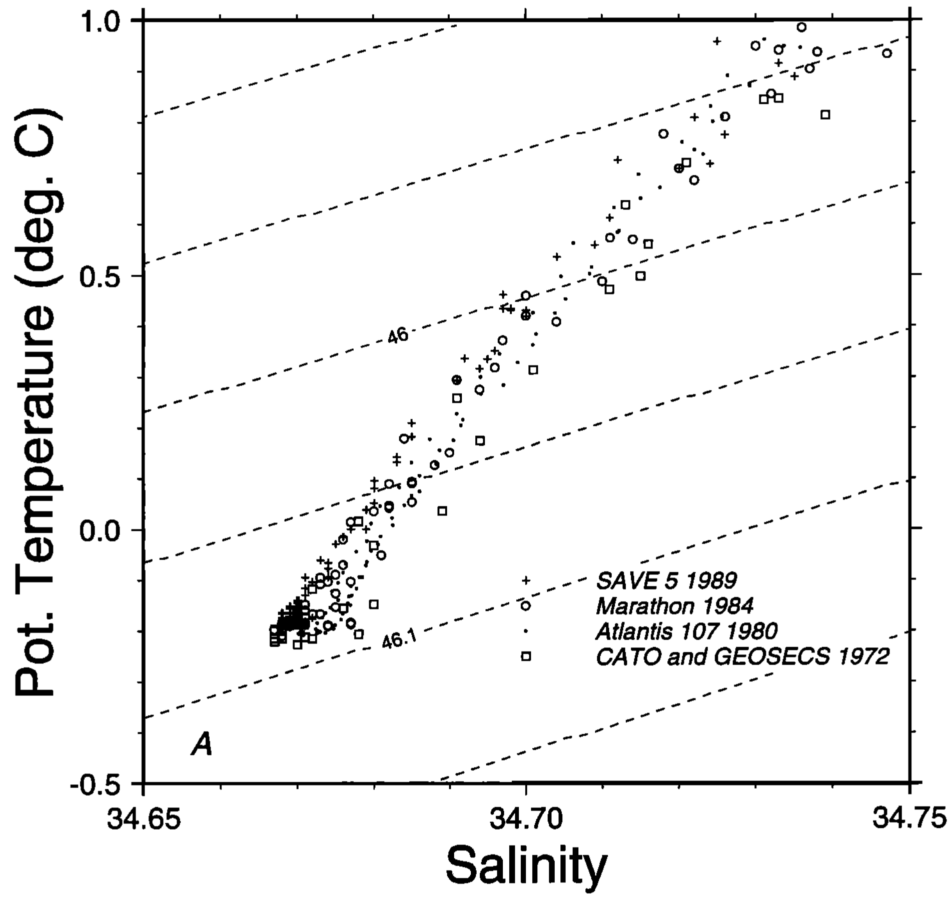

Figure 7. Temperature (in degrees Celsius) versus salinity plots from data in the northwestern Argentine Basin. The two types of water mass modification are still seen; however, the signal outside the recirculation is reduced from the magnitude seen in Figure 6.

1989 is lighter than the bottom water seen previously. While this seems to be a small difference in the temperature-salinity plot, the vertical scale of the anomaly is quite large (see Figure 5). In the abyssal layer the SAVE data are considerably less dense than the earlier data; at the bottom this difference amounts to about 0.05 $m^3\ kg^{-1}$ in potential density anomaly (~0.008 difference in salinity ratios and ~0.05°C difference on the densest SAVE isopycnal). We find no evidence of a thickening of this layer in the SAVE data from the eastern Argentine Basin (see Figure 5); however, the coverage in this area is poor, and it is possible that the layer is banked up against the Mid-Atlantic Ridge.

The data from the northern basin shown in Figure 1 show a similar change over time. Figure 7 compares abyssal properties in 1972, 1980, and 1984 to those found in 1989. Again, the SAVE data are colder and fresher on isopycnal surfaces than the AABW observed previously. Here the anomaly is less pronounced, reflecting the greater distance from the source waters, or dilution, and also possibly that the anomalous AABW has not yet completely filled the northwestern portion of the abyssal basin. There seems to be no consistent trend in the data from 1972 and 1980, suggesting that the AABW had fairly constant water mass properties over that time period. The stations in 1984 do seem to show a trend to a warmer and fresher water mass; however, they are still cooler and saltier on isopycnal surfaces than the SAVE data of 1989. The bottommost AABW found in 1989 is also less dense than that found previously, as in the southern regime.

The modifications observed in the AABW, cooling and freshening on density surfaces and decrease in maximum density, must be the result of two processes, mixing and inflow of "new" water with anomalous properties. Inflow of cool, fresh bottom water can not displace denser water, and this denser water has not been observed leaving the basin through the Vema Channel; thus another process such as diapycnal mixing must be invoked to absorb the dense bottom water. However, mixing alone without a new end-member is not sufficient to shift temperature-salinity properties off the mixing line shown in Figures 6 and 7.

One possible scenario is that anomalously cool and fresh, yet less dense, inflow from the south fills the western abyssal basin as denser water is advected to the northeast with the basinwide circulation. This water mass forms a new end-member on the temperature-salinity diagram, leading to the offset observed. As there are few data in the eastern portion of the basin, we can not determine whether the densest layers have been mixed vertically into the overlying water or whether they have been advected into the poorly sampled areas.

3.4. Volumetrics

Volumetric calculations over the Argentine Basin below 3000 m for both the historical data set and the SAVE data

Table 2. Volume and Average Salinity Associated With Each Temperature Layer for Augmented SAVE Data, Historical Data, and Their Difference

	SAVE Data		Historical Data		
	Volume, 10^{14} m^3	Salt	Volume, 10^{14} m^3	Salt	Volume Difference, 10^{14} m^3
Level, °C					
0.3 to 0.2	3.6847	34.6872	3.3214	34.6911	0.3633
0.2 to 0.1	4.5822	34.6826	5.0075	34.6865	−0.4253
0.1 to 0.0	6.2528	34.6780	7.2147	34.6808	−0.9619
0.0 to −0.1	9.6484	34.6741	9.7532	34.6769	−0.1048
−0.1 to −0.2	20.9370	34.6700	20.5220	34.6727	0.415
−0.2 to −0.3	7.0207	34.6639	11.8830	34.6641	−4.8623
−0.3 to −0.4	0.4532	···	0.4517	···	0.0015
−0.4 to −0.5	0.0586	···	0.5500	···	−0.4914
Total below 0.2°C	52.6376	58.7035	−6.0659		

Colder than −0.3°C, there is insufficient data to interpolate salinity correctly.

provide a means to quantify the magnitude of the AABW anomaly. Because the historical data have more even spatial coverage, though poorer vertical and spatial resolution, the data from SAVE were augmented by historical data (three stations) in the western Argentine Basin. This addition fills in the structure of the boundary current for the interpolation. The use of these earlier data means that differences between the data sets will be reduced.

The calculations were made by vertically interpolating station data onto set isotherms. The unequally spaced data points were then gridded spatially, and volumes were computed down to the bathymetry at 5-min resolution.

The results are summarized in Table 2. Volume within abyssal temperature layers is plotted in Figure 8. The data sets both show the same large-scale water mass distributions, with differences within roughly 10% of the volume within the layer except for the volume within the −0.2 to −0.3°C class. As expected, we see a large volume of abyssal water and a relatively abrupt transition between AABW and LCDW, with a temperature greater than 0.2°C.

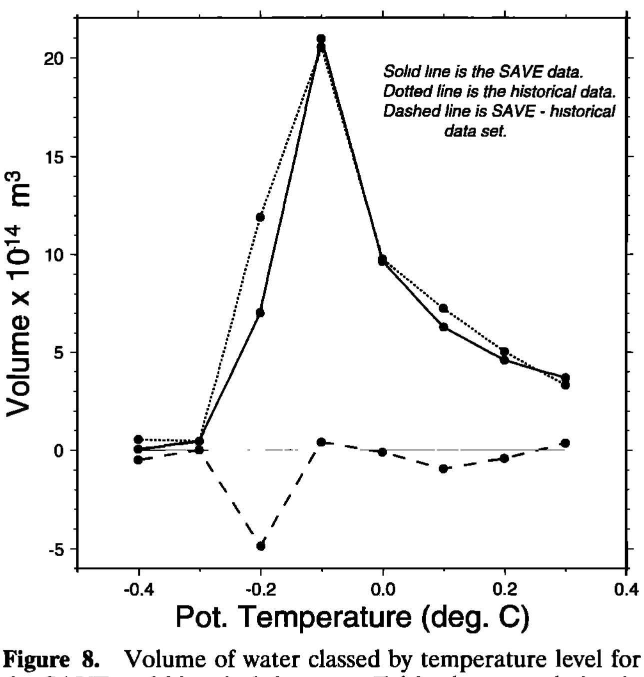

Figure 8. Volume of water classed by temperature level for the SAVE and historical data sets. Fairly close correlation in local maxima suggests that temperature data are accurate in both data sets and that gridding error is not too large. Volume difference shows conversion from colder to warmer modes within the −0.2°C level of AABW.

Figure 8 also shows the difference of the two data sets. The volume change in the abyssal layer at −0.2°C clearly dominates. Waters with a temperature between −0.2 and −0.3°C have been vertically mixed and/or advected out of the domain sampled causing a net change in the amount of water colder than 0.2°C (AABW). Essentially no volume change is seen in waters colder than −0.3°C; this is partly due to the limited areal extent of these layers within the basin.

Because volume below the 0.2°C horizon has not been conserved, a change in the inflow rate of AABW is required to account for the change. However, as discussed above, a simple cessation of inflow cannot account for cooling and freshening on isopycnal surfaces. A reduction in inflow must be coupled with changes to the source waters for the inflow to account for the observed variability. It is also possible that the inflow warmed sufficiently that waters warmer than 0.2°C are partly of Antarctic origin, and that the inflow rate has not varied significantly.

4. Spreading

The previous sections have shown variability in the temperature and salinity of AABW within the Argentine Basin, over very short timescales (~5 years). The Argentine Basin is located close to the Weddell Sea AABW source region (roughly 15° latitude separation). Thus changes in AABW formation should be relatively rapidly transmitted into the Argentine Basin. However, the timescales involved suggest a more rapid and vigorous abyssal circulation than has been predicted. While we are unable to determine how fully the AABW anomaly had spread throughout the basin due to lack of data in the east, we can examine the spread of the anomaly in the DWBC and compare this "apparent" velocity to estimates of NADW spreading rate in the northern hemisphere. In the following, SAVE data collected just north of the Vema Channel outflow from the Argentine Basin are compared with historical data to show the spread of the anomalous AABW into the Brazil Basin.

The northern limit of the anomalous AABW is fairly well resolved by the SAVE data in the Brazil Basin. Zonal transects crossing the boundary current were carried out in 1983 (R/V *Oceanus* at 23°S), and in 1989 (SAVE Leg 3 at 25°S). In Figure 9, three stations which span the boundary current are shown for each time period. In 1983 (stations 43, 45, and 47) they show a zonally homogeneous abyssal layer extending from the boundary into the interior of the basin, suggesting that prior to 1983 the AABW had a very stable water mass composition over the basin circulation timescale.

In contrast, the SAVE section (stations 116, 117, and 118) shows a clear distinction between the water mass properties of the DWBC and the interior recirculation. SAVE station 116 is the westernmost of the three stations shown; it is fresher than the *Oceanus* data, reflecting inflow of anomalous water in the boundary current. SAVE station 118, to the east, shows no sign of change in abyssal temperature and salinity, suggesting that the anomaly has not spread to the basin interior. The intermediate station 117, located at 33°W, shows the fresher abyssal layer but reverts at middepth to the more saline, colder water which fills the basin interior.

This suggests that the anomalous water mass is entering the Brazil Basin in the DWBC, banked up in a wedge-shaped feature against the western continental rise. The anomaly appears to have penetrated in the DWBC to at least 25°S but not yet spread into the interior. This is consistent with the concept of a western boundary current as transporting fluid and tracers rapidly in a narrow band. Recirculation then transports the properties of the fluid in the boundary current into the basin interior on longer timescales and acts to decrease tracer concentrations in the DWBC through mixing.

The northward penetration of the anomaly in the boundary current is of sufficient magnitude to impact the northward AABW transport mode. Transport calculations for the two sections are computed as by *McCartney and Curry* [1993]. The shear between station pairs is calculated above the greatest common depth and is extrapolated downward using a weighting factor dependent on the vertical density gradient. A zero motion reference level of 1.9°C is used, as this isotherm marks the conventional horizon between the southward flowing NADW and the northward flowing AABW in the Brazil Basin [*Wright*, 1970].

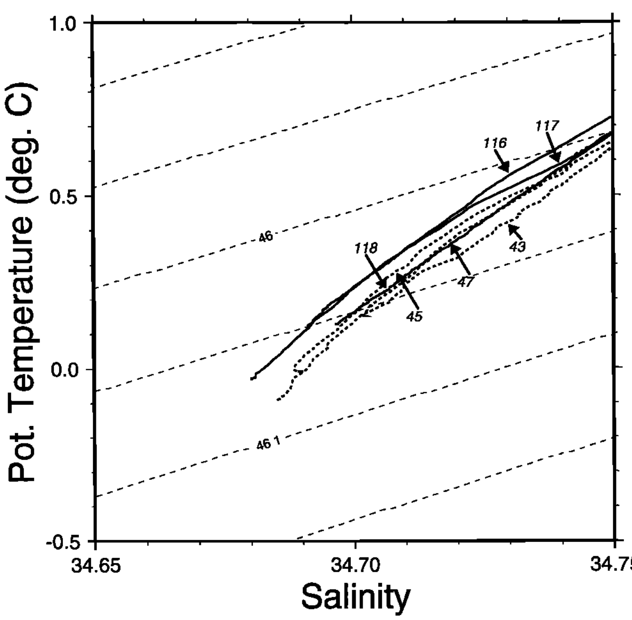

Figure 9. Temperature (in degrees Celsius) versus salinity for SAVE and *Oceanus* data in the Brazil Basin. Stations are as follows: Solid line, SAVE stations from west to east, 116, 117, 118 (1989), dashed line, *Oceanus* stations from west to east 47, 45, 43 (1983).

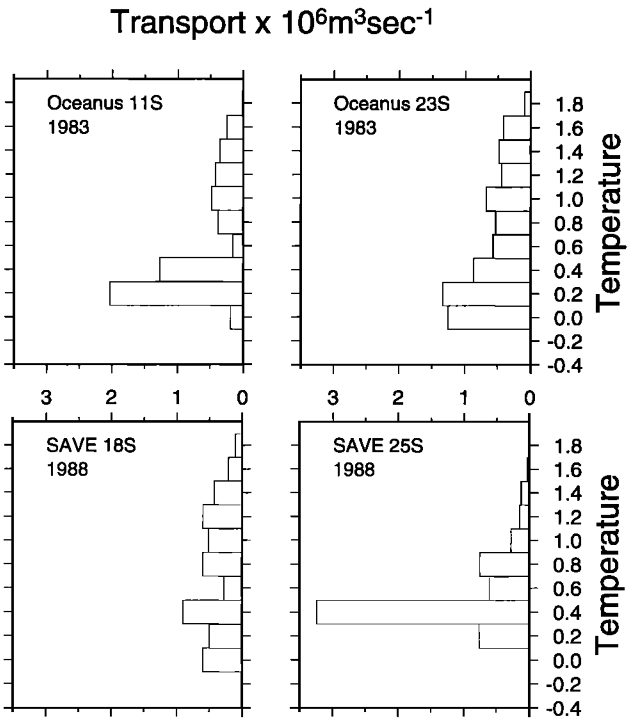

Figure 10. Transport histograms over two time periods, SAVE in 1988 and *Oceanus* in 1983. While the *Oceanus* section at 23°S and the SAVE leg at 25°S are directly comparable, the SAVE leg at 18°S lies intermediate between the two *Oceanus* sections and must be considered accordingly.

Mixing with the overlying water mass modifies the temperature of the northward flowing AABW. This warming is discussed by *McCartney and Curry* [1993] and *Whitehead* [1989]. In the *Oceanus* data collected in 1983 at 23°S, the maximum northward transport occurred in the temperature range between −0.1°C and 0.3°C (Figure 10). Five years later at 25°S, the dominant mode of northward transport had "warmed" to between 0.3°C and 0.5°C. This warming appears to be unrelated to the normal advective diffusive mixing which reduces the property maxima along the path of the AABW. Instead, it is a result of the decrease in maximum density of AABW due to the water mass anomaly seen in the Argentine Basin.

The distinction between strong western boundary and less dynamic midbasin flow is well defined in the transport calculations. To the west of SAVE station 118, 5.1 × 10^6 m^3 s^{-1} of AABW flowed northward in the boundary current. Between stations 118 and 117, 2.0 × 10^6 m^3 s^{-1} of AABW flowed northward in the transition regime, and there was negligible northward flow to the east of station 117.

AABW warming is less evident farther to the north, possibly because the altered AABW may not yet be the dominant water mass in the boundary current. The SAVE section at 18°S shows no dominant transport mode. This is in part due to an interior northward flow path for AABW [*Durrieu De Madron and Weatherly*, 1994]. Within the boundary current the northward transport mode is warmer, whereas the flow in the interior is

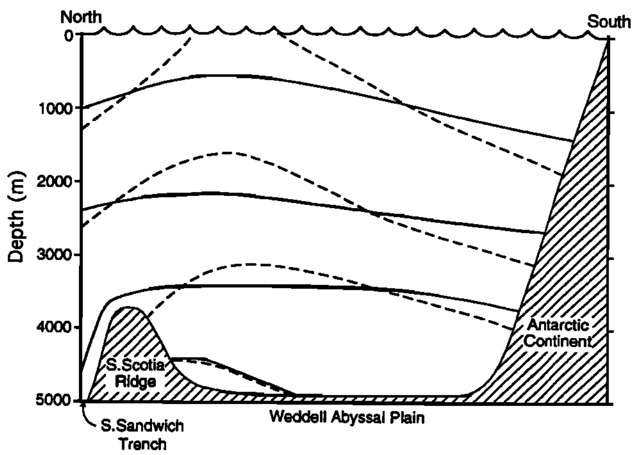

Figure 11. Conceptual picture of the isopycnals within the Weddell Sea during "normal" deep water formation (solid lines) and with preconditioning or gyral spinup leading to midgyre convection (dashed lines). As a result of the preconditioning, less dense fluid impacts the sill at the outflow of the basin, as shown.

composed of older water with preanomaly colder and fresher temperature and salinity characteristics.

The SAVE leg 1 transport calculations cannot be made across the whole basin (and are thus not shown), as the section crossed the equator to the east. However, the western portion of the section suggests that the warming of the dominant transport mode had not propagated this far north. Thus the anomaly had clearly penetrated north of the 25°S section by 1988, and its influence seems to have affected the 18°S section, but there is no evidence that it reached SAVE Leg 1 at 10°S.

Assuming that the Atlantis data collected in 1980 reflect the inflow characteristics of AABW better than the ABCS data shown in Figure 6, which is from the basin interior, then a propagation velocity along the western boundary can be estimated at at least 2.3 cm s^{-1}. This is comparable to than the 2.2–2.5 cm s^{-1} southward velocity of the lower NADW core in the North Atlantic estimated by *Smethie* [1993]. Such rapid response to changes in water mass properties in the source region suggests a close coupling between the ocean-atmosphere system in high latitudes where deep water masses are ventilated and the deep ocean reservoir.

5. Processes

In the AABW flow system, a cold, fresh inflow to the Argentine Basin is mixed advectively and diffusively with the surrounding warmer, more saline, and less dense water mass. Thus the AABW transport equatorward becomes gradually warmer and saltier. This process of mixing is a simple two end-member process, resulting in a linear temperature-salinity profile through the deep water up to the NADW. Two types of change to this system have been isolated; a loss of the densest end-member in the mixing profile, and a shift to AABW which is colder and fresher on density surfaces. A reduction or cessation of inflow of "new," very dense AABW would lead to a gradual loss of density, as mixing acted to reduce the property gradients in the deep water. A shutoff of inflow, however, can only result in mixing along the temperature-salinity profile established; it cannot cause a shift in the deep water to a cooler and fresher state. For this, a new mixing end-member is required.

The water mass variability appears to be limited to the AABW and does not extend vertically into the LCDW. This suggests that the source of the changes is located in the Weddell Sea or in the control of AABW flow between the Weddell Sea and the Argentine Basin. Two instances of midgyre convection in the Weddell Sea, in 1973 [*Gordon*, 1982; *Martinson et al.*, 1982] and 1983 [*Bersch*, 1988], have been shown to modify middepth water within the central Weddell Gyre which forms a major constituent of AABW.

The first, in the mid 1970s, was due to a large, persistent polynya which drifted over the course of several years in a westerly direction with the mean flow in the eastern Weddell Gyre. The opening in the sea ice permitted large air-sea heat flux, which cooled the surface waters sufficiently to permit deep convection. Convection generated −0.2 to −0.75°C cold anomalies in the water from 500 to 2500 m and −0.2 to −0.4 fresh anomalies from 500 to 3500 m [*Gordon*, 1982]. Thus water from 2500 to 3500 m was fresh and had a temperature of about −0.5°C. If this water replaced or mixed with Weddell Sea Deep Water to form AABW, it would cause a cool fresh anomaly, consistent with our observations.

The second observations in 1983 found convectively cooled, fresh WSDW in the central gyre. Temperature anomalies of between −0.5 and −1.0°C and salinity anomalies of −0.03 to −0.06 occurred in the warm deep water between 500 and 3000 m. Convection did not appear to penetrate as deeply as the previous event, but the water column was more depleted in salt than during the polynya convective plume [*Bersch*, 1988]. Because this event was more recent and had the stronger salinity signal, it seems a plausible source for the AABW seen in the Argentine Basin in 1988–1989.

While we cannot isolate a specific event as leading to the water mass variability seen, these two instances of central gyre deep convection give a possible mechanism for producing anomalous AABW. This convective middepth water mass ventilation must be directly related to atmospheric forcing within the Weddell Gyre. It seems reasonable that preconditioning of the water column through updoming of isopycnals may have occurred as a precursor to convection. *Martinson et al.* [1982] speculate that such preconditioning could occur in the central Weddell sea, leading to locally elevated isopycnals that are advected around the gyre with the mean flow. A cartoon showing the effects of such a preconditioning scenario, albeit larger scale, is shown in Figure 11. Such an updoming might result from spinup of the gyre caused by anomalous wind forcing. Conservation of mass in isopycnal layers would require a corresponding downwelling at the northern boundary of the gyre as shown in the dashed portion of Figure 11. Thus lighter fluid would impact the sills along the Scotian Arc, leading to less dense flow out of the Weddell Sea.

In such a scenario, the maximum density of the overflow water is reduced, corresponding to the loss of the densest isopycnal layers in the Argentine Basin, as observed. The outflow water is also colder and fresher due to the midgyre convection. While this outflow water mass mixes with overlying LCDW and underlying bottom water before entering the Argentine Basin as AABW, its temperature and salinity are sufficiently lower than the observed anomalous AABW within the Argentine Basin to compensate for some mixing.

If midbasin deep convection within the Weddell Sea is the origin of the anomalous AABW seen in the SAVE data within the Argentine Basin, then the effects of the atmospheric conditions within the Weddell Sea which caused middepth convection are propagated to the deep ocean on very short timescales. The anomalous water mass would have traveled to the Argentine Basin and continued on to roughly 18°S in the Brazil Basin in 5–11 years depending on which instance of convection generated the anomaly. Information about the transient tracer concentrations within the abyssal waters may serve to confirm or deny this hypothesis.

6. Conclusions

AABW water mass properties within the Argentine Basin have changed significantly in less than 10 years. Two types of variability have been isolated: (1) The AABW in 1989 was fresher (~0.008 salinity ratio) and somewhat cooler (~−0.05°C) than the AABW of 1980–1987 on density surfaces, although these differences vary with position. (2) The densest water noted between 1980–1987, with a vertical extent of 500 m in some areas of the basin, disappeared almost completely by 1989, resulting in a warming of the abyssal layer. The greatest change is associated with the C-shaped recirculation cell in the southwestern Argentine Basin. Weaker change signals are found in the northwestern part of the basin, but no clear indication of this anomaly exists in the eastern part of the basin along the flank of the Mid-Atlantic Ridge.

The change in dominant AABW transport mode within the Brazil Basin suggests that the signal has propagated into the western boundary of the Brazil Basin, to roughly 18°S, requiring an equatorward velocity of at least 2.3 m s^{-1}.

Variability in AABW properties in the Argentine Basin may be linked to the Weddell Sea source. Convection to middepth in the central Weddell Gyre has been documented in 1983 and the mid 1970s. Both of these events caused freshening and cooling of the middepth water consistent with what we see in the Argentine Basin AABW. Preconditioning of the water column through updoming of isopycnal surfaces associated with deeply penetrating convective events might also lead to a less dense overflow out of the Scotian Arc.

WSDW formation may be at least as variable as deep water formation in the Labrador and Greenland Seas. *Brewer et al.* [1983] showed a salinity change in the deep waters in the northern North Atlantic along 58°N of 0.02, about twice the signal in salinity that we see. In an average of their data, however, no appreciable change in the density of the deepest waters referenced to 2000 m is found. These measurements at 58°N are considerably closer to source waters than the measurements discussed here (50°S–30°S). Thus we feel that the magnitude of the signal, given the advective diffusive modification of the deep waters as they flow from their sources, is comparable in magnitude to the variability found in the North Atlantic. While *Brewer et al.* [1983] suggested that the variability seen in the North Atlantic reflected a decadal scale fluctuation, observations within the Labrador Sea [*Lazier*, 1988] show interannual variability in the Denmark Strait Overflow Water of a similar magnitude. Observations within the Argentine Basin suggest a long period of stable water mass properties within the AABW, followed by a rapid shift to a new state. However, interannual variability cannot be ruled out.

We do not propose that variability in the deep Argentine Basin is necessarily a result of long-term or global climate change. However, high-frequency variability, such as that seen in the Argentine Basin, may indicate that the deep oceans are not as static as many studies have suggested. Thus their role as a buffer of short-term atmospheric variability may be more effective than previously believed. Furthermore, such rapid shifts in conditions may indicate that oceanic climate fluctuations can occur very rapidly, on decadal timescales, as has been suggested by *Broecker* [1987] and others.

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V. S. Coles, HPEL, P.O. Box 775, Cambridge, MD 21613. (e-mail: vcoles@seagrape.hpel.umd.edu)

M. S. McCartney, Woods Hole Oceanographic Institution, Woods Hole, MA.

D. B. Olson, Rosenstiel School of Marine and Atmospheric Science, MPO, 4600 Rickenbacker Causeway, Miami, FL 33149.

W. M. Smethie Jr., Lamont-Doherty Earth Observatory, Palisades, NY.

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