Tracing Southwest Pacific Bottom Water Using Potential Vorticity and Helium-3

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ABSTRACT

This study uses potential vorticity and other tracers to identify the pathways of the densest form of Circumpolar Deep Water in the South Pacific, termed “Southwest Pacific Bottom Water” (SPBW), along the 28.2 kg m$^{-3}$ surface. This study focuses on the potential vorticity signals associated with three major dynamical processes occurring in the vicinity of the Pacific–Antarctic Ridge: 1) the strong flow of the Antarctic Circumpolar Current (ACC), 2) lateral eddy stirring, and 3) heat and stratification changes in bottom waters induced by hydrothermal vents. These processes result in southward and downstream advection of low potential vorticity along rising isopycnal surfaces. Using $\delta^3$He released from the hydrothermal vents, the influence of volcanic activity on the SPBW may be traced across the South Pacific along the path of the ACC to Drake Passage. SPBW also flows within the southern limb of the Ross Gyre, reaching the Antarctic Slope in places and contributes via entrainment to the formation of Antarctic Bottom Water. Finally, it is shown that the magnitude and location of the potential vorticity signals associated with SPBW have endured over at least the last two decades, and that they are unique to the South Pacific sector.

1. Introduction

The divergence of tectonic plates along midocean ridges introduces a gap in the earth’s crust, forming an axial valley, from which geothermal heat is released at a relatively high rate, by both conductive and convective processes (Wilson 1965; Morgan 1971). Stommel (1982) concluded that heat in the buoyant plumes from hydrothermal vents is effectively diffused unless “competing mechanisms are not overpowering.” Stommel’s paper considered the circulation above the East Pacific Rise, located deep below the South Pacific subtropical gyre, where little “overpowering” circulation would be expected. Hydrothermal vents are apparently common along midocean ridges, but with a highly intermittent distribution; they have recently been identified on parts of the 2600-km-long Pacific–Antarctic Ridge, southwest of the East Pacific Rise (Winckler et al. 2010). In complete contrast to the East Pacific Rise, the
Pacific–Antarctic Ridge lies within the main flow of the Antarctic Circumpolar Current (ACC), presumably an “overpowering” flow. In addition, vigorous eddy activity and the upwelling of deep waters all coincide with the multiple hydrothermal sources along this ridge.

Hydrothermal plumes from the major ocean ridges have been traced both near and far afield from the vent sources. Veirs et al. (1999) used a stability function along the Juan de Fuca Ridge to identify which vent sources along the ridge are associated with temperature and light attenuation signals. They concluded that the stability anomaly was confined to within about 50 km from the vent source. Hydrothermal signals have also been detected at larger distances from the sources on the ridge crests. For example, $\delta^{3}$He and manganese have been used to trace hydrothermal fluid originating at the East Pacific Rise and flowing more than 2000 km west of the ridge (Lupton and Craig 1981; Klinkhammer 1980). In addition, Johnson and Talley (1997) compared stratification measures and temperature-salinity anomalies with $\delta^{3}$He to trace various pathways of the hydrothermal plumes that had originated along the East Pacific Rise. In the South Atlantic Ocean $\delta^{3}$He indicates hydrothermal plumes also extending thousands of kilometers from the hydrothermal sources along the southern half of the Mid-Atlantic Ridge (Rüth et al. 2000).

Winckler et al. (2010) identified the Pacific–Antarctic Ridge as a major hydrothermal plume source that can be traced via a $\delta^{3}$He plume along the 28.2 kg m$^{-3}$ neutral density surface ($\sigma_2 \approx 1037.12$ kg m$^{-3}$) south of the ridge along 150°W and across the 67°S cruise transect. They concluded that this $\delta^{3}$He plume, distinct from the much stronger $\delta^{3}$He source along the East Pacific Rise (Lupton 1998), could be used for tracing the South Pacific abyssal circulation. However, $\delta^{3}$He measurements in the South Pacific are sparse, and thus this tracer cannot solely be used to describe in detail the large-scale circulation. Here, we expand upon the analysis of Winckler et al. (2010) by showing that the $\delta^{3}$He signature they observed coincides with a distinct potential vorticity signal in the deep South Pacific sector of the Southern Ocean.

In the South Pacific sector of the Southern Ocean, strong interactions occur between the eastward-flowing ACC and topography (Gnanadesikan and Hallberg 2000; Rintoul et al. 2001). The ACC is comprised of fronts or jets (cf. Orsi et al. 1995; Sokolov and Rintoul 2000; Rintoul et al. 2001). The ACC and topography (Gnanadesikan and Hallberg 2000; Orsi and Wiederwohl 2009; Orsi 2010). Here we refer to water in the ACCbw density range in the South

$$1 \delta^{3}\text{He} = 100 \times \frac{[3\text{He}/4\text{He}]_{\text{water sample}} - [3\text{He}/4\text{He}]_{\text{atmosphere}}}{[3\text{He}/4\text{He}]_{\text{atmosphere}}}$$

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Table 1. Cruises of analyzed hydrographic (including $^3$He) data. Note: the CLIVAR S4P line also includes western (172°E), central (170°W), and eastern (150°W) Ross Sea sections that extend south of 67°S to about 72°S. Also, the P06 cruise had three legs and the chief scientists for each leg are listed. The cruise chief scientists and the Principal Investigator (PI) for $^3$He are also included.

<table>
<thead>
<tr>
<th>Line</th>
<th>Latitude</th>
<th>Longitude</th>
<th>Ship</th>
<th>Year</th>
<th>Months</th>
<th>Chief Scientist(s)</th>
<th>$^3$He PI</th>
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<td>62°–21°S</td>
<td>150.5°W</td>
<td>Knorr</td>
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<td>150°W</td>
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<td>2005</td>
<td>Jan–Feb</td>
<td>B.Sloyan/J.Swift</td>
<td>P.Schosser</td>
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<tr>
<td>WOCE S03</td>
<td>66°–44°S</td>
<td>140°E</td>
<td>Aurora Australis</td>
<td>1994</td>
<td>Dec–Feb</td>
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<tr>
<td>WOCE A09</td>
<td>19°S</td>
<td>37°W–9°E</td>
<td>Meteor</td>
<td>1991</td>
<td>Feb–Mar</td>
<td>G.Siedler</td>
<td>A. Putzka</td>
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<td>42°W–13°E</td>
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<td>B.King</td>
<td>—</td>
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<tr>
<td>CLIVAR A13.5</td>
<td>54°S–5°N</td>
<td>0°</td>
<td>Ronald H. Brown</td>
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<td>Mar–Apr</td>
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<td><strong>Indian</strong></td>
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Pacific Ocean as “Southwest Pacific Bottom Water” (from herein, SPBW), based on the potential vorticity and δ$^3$He signals it acquires at a middepth source located far from the Antarctic continental slope. The name “ACCbw” refers to a water type restricted to the ocean floor in the ACC region and then exported equatorward (see temperature and salinity distribution in plates 220–221 in Orsi and Whitworth 2005). Traces of SPBW signals are found downstream, both to the east at the bottom of southern Drake Passage, and to the west at deep levels of the southern Ross Gyre. The interior vertical mixing of new bottom waters from the Ross Sea needed for the regional AABW to continue equatorward over the ridges is therefore partly reflected in the SPBW characteristics.

The vertical layering of water masses along the path of the ACC and the formation of AABW in the Ross Sea have been extensively documented; however, our understanding of the interaction between these major elements of the deep stratification remains incomplete. We focus our study on the potential vorticity and other tracer signals associated with three major dynamical processes occurring in the vicinity of the Pacific–Antarctic Ridge: 1) the flow of the Antarctic Circumpolar Current (ACC), 2) lateral eddy stirring, and 3) heat and stratification changes in bottom waters induced by hydrothermal vents. We trace the circulation of a water mass we term Southwest Pacific Bottom Water over roughly two decades, using results from the World Ocean Circulation Experiment (WOCE; 1991–2000) and Climate Variability and Predictability (CLIVAR; 2000 onward).

The outline of the remainder of the manuscript is as follows. We begin with a description of the cruise data and define potential vorticity used here as a tracer. We then describe the pathways of distinct potential vorticity signals in the Southwest Pacific bottom water density class from the Pacific–Antarctic Ridge to the Ross Sea and Drake Passage. We assess the influence of other major ridges and finally summarize and discuss our results.

2. Data and methods

a. WOCE and CLIVAR cruise data

This study uses hydrographic data from multiple cruises (Table 1; Fig. 1). All of the data and final cruise reports are available from the CLIVAR and Carbon Hydrographic
Data Office (CCHDO; http://cchdo.ucsd.edu/). We have also analyzed 1973/74 hydrographic data from the Geochemical Ocean Sections Study (GEOSECS), also available from CCHDO. None of these data were collected during winter; however, our research focuses on deep and abyssal waters so we do not expect any seasonal bias. The main variables of interest are pressure, potential temperature, and salinity, measured using conductivity–temperature–depth (CTD) instruments with an accuracy of at least 2 m, 0.005°C, and 0.005, respectively.

CTD results are routinely reported at 1-dbar resolution, but we subsampled that data every 10 dbar. Horizontal station spacing for WOCE and CLIVAR is routinely 55 km. Potential vorticity, a measure of stratification and a focus for our study, is calculated using neutral density. Neutral density is based on temperature and salinity and spatial location of the cast, and is calculated using software detailed in Jackett and McDougall (1997). The tracers, $\delta^3$He, silicate and chlorofluorocarbons were sampled using Niskin-type bottles and a Rosette and typically have accuracies of 1.5%, 2%, and 1%, respectively. Most of these cruises collected 36 bottle samples at each station; however, for these tracers only silicate was measured in all bottles. All of the WOCE and CLIVAR cruise data have been subjected to multiple levels of quality control and are believed to be of very high quality by modern standards. The GEOSECS data were high quality for the time (1970s), but are not quite as accurate or precise as the newer data.

b. Potential vorticity

Potential vorticity is a measure of stratification in the water column, and we are able to detect subtle changes in the abyssal ocean stratification with the use of hydrographic data that is high resolution in the vertical direction. Potential vorticity ($Q$) quantifies the ratio of the combined relative vorticity ($\zeta$) and planetary vorticity (or Coriolis parameter; $f$; negative in the Southern Hemisphere) to the thickness of the water column, and is given by

$$Q = -\frac{(\zeta + f) \frac{\partial \rho}{\partial z}}{\rho},$$

where $\rho$ is the mean density, and the local vertical change in density over that in depth is given by $\frac{\partial \rho}{\partial z}$. We assume that the relative vorticity is negligible for large-scale flows, and we use neutral density $\rho'$. Hence,

$$Q = -\frac{f}{\gamma'} \frac{\partial \gamma'}{\partial z}. \quad (2)$$

We assume a positive sign for potential vorticity throughout our text, figures, and equations. A small potential vorticity corresponds to a small change in density with respect to depth, and hence weak stratification (and a well mixed fluid).

In a layered framework, the potential vorticity equation can be written as

$$\frac{\beta u}{f} + u \frac{Vh}{h} = \frac{w^r - w^s}{h} + \frac{h}{f} \nabla^2 \left( \frac{f}{h} \right). \quad (3)$$

The first term represents the meridional advection of planetary vorticity. The second term ($h$ is layer thickness)
is vortex stretching because of the isopycnal component of the velocity (conservative stretching). The third term represents the diapycnal (nonconservative) stretching because of diapycnal fluxes at the top (T) and bottom (B) of the layer, and the fourth term is the isopycnal diffusion of potential vorticity (where mixing $\kappa$ is assumed to be constant).

Potential vorticity is not conserved near the plume source. Heat injected from the earth’s mantle and flow over the bottom can generate diapycnal fluxes $w^*$ that modify potential vorticity. Frictional torques experienced by the abyssal ocean will alter the local density field and break the conservation constraint for potential vorticity as well. Large-scale lateral eddy stirring $\kappa$ does not directly modify density since the stirring is essentially along isopycnals, but it does smooth out potential vorticity gradients on these isopycnals, spreading tracers away from sources. These nonconservative influences may be a factor in setting the potential vorticity signals of SPBW once it moves away from the direct influence of the plume source.

3. Results

a. Identification of potential vorticity signals along the Pacific–Antarctic Ridge

Winckler et al. (2010) illustrated the existence of a unique $\delta^3$He plume south of the Pacific–Antarctic Ridge along the WOCE P16 section, as well as along the 67°S WOCE S4P section. We found that this distinct $\delta^3$He plume signal of over 10% at the ridge crest around 59°S coincides with a potential vorticity minima signal (Fig. 2a). The minimum in the potential vorticity of less than $5 \times 10^{-12}$ m$^{-1}$ s$^{-1}$ rises above the ridge along the 28.2 kg m$^{-3}$ neutral density surface (see yellow contour). Directly below the potential vorticity minima, we find a maximum in the potential vorticity of $1.1 \times 10^{-11}$ m$^{-1}$ s$^{-1}$ that is evident south of 60°S. We note that the potential vorticity of the plume is not conserved near the ridge crest, where it is influenced by geothermal heat, eddies, and bottom friction. South of 61°S and downstream of the ridge, the plume retains a potential vorticity close to $9 \times 10^{-12}$ m$^{-1}$ s$^{-1}$. The density range of SPBW encompasses
the low potential vorticity signal along the 28.2 kg m\(^{-3}\) surface.

The WOCE P16 section at 150°W lies within 30 km downstream of potential plume source regions identified by Winckler et al. (2010) (see their Fig. 3). We infer that the WOCE P16 section is close to a plume source, though we cannot determine the exact distance with the presently available measurements. We calculate the temperature anomaly in density space along the 28.2 kg m\(^{-3}\) surface, defined as the difference between the background temperature away from the ridge at 56°S (taken to be 0.6°C) and the temperature at the ridge crest at 59.5°S. We find that the Pacific-Antarctic Ridge crest has a 0.07°C temperature anomaly, which compares well with the 0.04°C anomaly found at the Juan de Fuca Ridge (Cannon and Pashinski 1997).

The localized minimum in the vertical temperature gradient (i.e., the local change in temperature with depth) along WOCE P16 clearly coincides with the potential vorticity minima (Fig. 2b); however, the relatively shallower minimum of the vertical salinity gradient is unrelated to the potential vorticity signals around the 28.2 kg m\(^{-3}\) surface (black contours in Fig. 2c). We also observed a strong correlation between the vertical temperature gradient and the potential vorticity signal for the CLIVAR P16 section (Fig. 2d). The temperature gradient is weaker in the modern data, possibly because of the small offset in location and large time difference between occupations. Along the WOCE P16 line, there is a clear \(\delta^3\)He maxima within the vicinity of the Pacific-Antarctic Ridge, between the mean temperature 0.4° and 0.6°C (Fig. 3).

Hydrothermal plumes generated close to, but upstream, of the WOCE P16 section undergo several modifications. Geothermal heat released into the water column at the Pacific–Antarctic Ridge would destabilize the water column, thus reducing the potential vorticity. However, potential vorticity can also be influenced by the enhanced flow over a rough bottom and mixing induced in the canyons on the flank of the ridge (Thurnherr et al. 2005).

To determine if additional mixing sources are at play, we can estimate the scale \((H)\) over which the plume effects might occur (without knowledge of the exact depths of the sources we are limited to scaling estimates). We use the equation from Speer and Helfrich (1995), \(H \approx 3.76(F_oN^2)^{1/4}\), where 3.76 is a coefficient based on laboratory and in situ observations. The local buoyancy frequency squared, \(N^2 = -gQ/f\) is based on the Coriolis parameter \((f)\), the acceleration due to gravity \((g = 9.81 \text{ m s}^{-2})\), and the potential vorticity \((Q)\) estimated in Eq. (2), and has units of s\(^{-2}\). The source buoyancy flux \((F_o)\) is typically between \(10^{-2}\) and \(10^{-1}\) m\(^4\) s\(^{-3}\) (Speer and Helfrich 1995). Taking the local background potential vorticity of the vent source along the WOCE P16 section to be \(5 \times 10^{-12}\) m\(^{-1}\) s\(^{-1}\) and the range of \(F_o\) values, we predict a vertical scale between 300 and 530 m. Given that the plume appears to have vertical scales substantially larger, rising to the 28.2 kg m\(^{-3}\) surface that flows ~1 km above the ridge crest (Fig. 2), it is likely

![FIG. 3. WOCE P16 potential temperature (°C) vs \(\delta^3\)He (%) for stations south of 55°S, between 800 and 3500 m. The potential temperatures associated with the potential vorticity \((Q)\) minima and maxima signals \((9 \times 10^{-12} \text{ m}^{-1} \text{s}^{-1}\) and \(1.1 \times 10^{-11} \text{ m}^{-1} \text{s}^{-1}\), respectively) are denoted by the red vertical lines.](Unauthenticated)
that other mixing mechanisms do occur at the ridge crest and on its flank. However, distinguishing these mixing mechanisms is beyond the scope of this study.

As it rises, the plume transports warmer waters into the surrounding fluid and entrains cooler ambient waters until it reaches a height where the background density is similar to its own (Joyce et al. 1986; Speer and Rona 1989). The plume (along with bottom mixed layer contributions) attains eventually a minimum value in potential vorticity of about $9 \times 10^{-12}$ m$^{-1}$ s$^{-1}$ and is advected along the 28.2 kg m$^{-3}$ surface both poleward into the Ross Gyre and downstream along the ACC southern fronts. We find that the layer affected by the plume flows deeper than the salty LCDW (Fig. 2c).
There is in addition a dynamical effect of the mixing and plume source near the bottom. Once the plume reaches its maximum vertical spreading height, it is transported westward via a diffusive phase speed or eastward via advection (Joyce and Speer 1987). The 28.2 neutral density surface lies within the coldest portion of LCDW, and hence within the deep meridional overturning cell (Speer et al. 2000), so when the d³He tracer is injected from the vent, it spreads laterally in the absence of ambient currents. We estimate the phase speed of the long baroclinic Rossby waves (Joyce and Speer 1987) using the formulation:

$$C_m = \frac{\beta g \alpha}{f^2 m^2 \bar{\theta}_z},$$

(4)

where the beta plane approximation $\beta = \frac{\partial f}{\partial y} \approx 10^{-11} \text{ m}^{-1} \text{ s}^{-1}$ is the latitudinal variation in the Coriolis parameter, $f \approx -1.25 \times 10^{-4}$ s$^{-1}$. The thermal expansion coefficient, $\alpha \approx 3 \times 10^{-4} \text{ deg}^{-1}$, and $g$ is acceleration due to gravity. The background vertical temperature gradient $\bar{\theta}_z$ is estimated as $5 \times 10^{-4} \text{ deg} \text{ m}^{-1}$ from Fig. 3. The vertical wavenumber $m = n \pi / H$, where $n = 1, 2, \ldots$ is the forcing mode, and $H$ is the previously calculated height the plume rises before equilibrating with the background buoyancy. The first three modes and an average plume rise height of 450 m give a westward phase speed of less than $2 \times 10^{-3}$ cm s$^{-1}$.

We estimate the eastward advection along the WOCE P16 section using the thermal wind equation:

$$u = \frac{g}{\rho_0 f} \int \frac{\partial \rho}{\partial y} dz,$$

(5)

where the $\rho$ is the ocean density, with a mean of $\rho_0 = 1028 \text{ kg m}^{-3}$. The meridional density gradient is integrated over depth ($z$), giving a mean eastward flow of at least 1 cm s$^{-1}$, 500 m above the ridge crest. The eastward advection, due to the ACC in this case, is one to three orders of magnitude greater than the diffusive phase speed, implying that the plume is entirely passive and swept downstream along the 28.2 kg m$^{-3}$ surface. Other studies have quantified a larger zonal velocity within the ACC (e.g., Renault et al. 2011; Zhang et al. 2012), further emphasizing our conclusion that advection via the ACC dominates the plume’s pathway.

Fig. 5. Schematic of the Pacific Ocean bathymetry (blue hues; m) overlaid with the fronts that encapsulate the ACC (from north to south): the SAF, PF, SACCF, and the SBDY. GEOSecs stations are indicated by magenta dots with stations shown adjacent in white boxes with magenta text. The WOCE P14, P15, P16 (P16A), diagonal P16 line (P16D), P17 (P17E), P18, P19, and S4P lines are denoted in red. The CLIVAR P16, P18 and S4P cruises are shown by blue dashed lines. The CLIVAR S4P cruise is divided into the main cruise section along 67°S (S4P), and the western, central, and eastern Ross Sea sections (RW, RC, and RE, respectively). Potential plume source locations are taken from Winckler et al. (2010). The bold part of the cruise lines denotes regions where we observe SPBW.
b. Tracing SPBW downstream of the Pacific–Antarctic Ridge

We trace SPBW potential vorticity signals along the path of the ACC in the WOCE cruises south and east of the Pacific–Antarctic Ridge (Fig. 4). We find that potential vorticity signals along the WOCE P16 diagonal section sampled between (135°W, 62°S) and (147°W, 58°S), as well as the southern parts of the meridional sections (P16, P17, P18, P19; Figs. 2 and 4), coincide with the $\delta^3$He tracer (~10%). The stronger $\delta^3$He signal (~14%) in the WOCE P18 and P19 sections are associated with the East Pacific Rise at a lower density (~28.0 kg m$^{-3}$; Fig. 4). SPBW is carried along the ACC between the Polar Front (PF) and Southern Boundary Front (SBDY), as shown by the bold parts of the cruise lines in Fig. 5. Along the WOCE S4P section, the SPBW potential vorticity and $\delta^3$He signals are present across the entire section, as noted in the Winckler et al. (2010) study. This indicates that SPBW flows poleward of the Pacific–Antarctic Ridge, in the Ross Gyre (discussed further in section 3d) and downstream along the ACC.

East of the S4P section, we have no evidence that the potential vorticity or $\delta^3$He signals associated with SPBW survive the intense mixing encountered in Drake Passage (Fig. 4; bottom right panel). The stronger $\delta^3$He signal (~17%) on the 28.0 kg m$^{-3}$ surface represents the diluted signal from the East Pacific Rise hydrothermal vents that is advected into the ACC from the

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**Fig. 6.** Potential vorticity (color and black contours, $\times 10^{-12}$ m$^{-1}$ s$^{-1}$) overlaid with 28.2 kg m$^{-3}$ neutral density surface (yellow) for Pacific lines P16, P18, and S4P. (top) Cruises during the WOCE period are shown; (bottom) CLIVAR cruise lines are shown.
north in the Southeast Pacific (Lupton 1998; Well et al. 2003; Naveira Garabato et al. 2007). The observations presented in our results thus far show strong evidence of SPBW potential vorticity and \( \delta^3\text{He} \) signals that are advected along the path of the ACC east of the Pacific-Antarctic Ridge. In the next section we will demonstrate that the SPBW we trace in this study originates along the Pacific-Antarctic Ridge, and not upstream of the ridge.

Remarkably, across the WOCE and CLIVAR periods the magnitude and location of the potential vorticity minima and maxima in the SPBW density range remain unchanged (Fig. 6), and the 28.2 kg m\(^{-3}\) surface remains almost fixed in depth for the P16, P18 and S4P sections. We can also trace the \( \delta^3\text{He} \) signal pre-WOCE using Pacific data collected during the GEOSECS (Fig. 7). There are two \( \delta^3\text{He} \) signals present south of 40\(^\circ\)S. The first is the strong \( \delta^3\text{He} \) signal of approximately 20\% near 28.0 kg m\(^{-3}\) that originates along East Pacific Rise (station 322; 43\(^\circ\)S, 130\(^\circ\)W), and dissipates by the time it reaches stations 282 and 290. The second \( \delta^3\text{He} \) maxima of \( \approx 10\% \) occurs near 28.2 kg m\(^{-3}\) for Stations 286 and 287 (67\(^\circ\)S to 56\(^\circ\)S, 170\(^\circ\)E to 170\(^\circ\)W), that lie around the Pacific–Antarctic Ridge. This second \( \delta^3\text{He} \) signal agrees in magnitude and density with the WOCE P16 (150\(^\circ\)W) observations and other Pacific sections. The GEOSECS, WOCE and CLIVAR \( \delta^3\text{He} \) data presented here provide evidence of hydrothermal vent activity along the Pacific–Antarctic Ridge near the 28.2 kg m\(^{-3}\) surface with \( \delta^3\text{He} \) of \( \approx 10\% \) since the early 1970s.

c. Signals upstream of the Pacific–Antarctic Ridge

To examine the uniqueness of the SPBW potential vorticity and \( \delta^3\text{He} \) signals found in the Pacific basin, we analyze cruise data directly upstream of the WOCE P16 section (WOCE P14 and P15; Fig. 8) and three Indian Ocean lines upstream of the Pacific–Antarctic Ridge, namely I08S, I09SS, and S03 (Fig. 9 and Table 1). In Fig. 8 the 28.2 kg m\(^{-3}\) surface does not intersect the Pacific–Antarctic Ridge (\( \approx 64\)\(^\circ\)S), but rather flows at least 500 m above the ridge crest. We observe a maxima in the potential vorticity \( (1.1 \times 10^{-12} \text{ m}^{-1} \text{ s}^{-1}) \) south of the ridge crest along both the P14 (\( \approx 64.5\)\(^\circ\)S) and P15 (\( \approx 66\)\(^\circ\)S) sections. The maxima is below the 28.2 kg m\(^{-3}\) surface for the P14 section (similar to the maxima illustrated in Fig. 4), however, the maxima and 28.2 kg m\(^{-3}\) surface coincide along the P15 line. The potential vorticity minima for the P14 section is difficult to observe in Fig. 8, however, it is clearly highlighted in Fig. 9a; the P15 section minima is less well defined as data unavailable at densities greater than 28.21 kg m\(^{-3}\). In addition, the GEOSECS station 282, located north of the crest near
the WOCE P14 section (Fig. 5), indicates a $\delta^3$He maxima near 28.2 kg m$^{-3}$ (Fig. 7). The P14 (and to a lesser extent, P15) minima in potential vorticity near the 28.2 kg m$^{-3}$ surface combined with the GEOSECS station 282 $\delta^3$He data imply hydrothermal plume sources ~900 km farther upstream than the region proposed in Winckler et al. (2010) (see yellow bars in Fig. 5 for their potential plume sources).

Near 28.2 kg m$^{-3}$, we also observe a strong deflection in the P16 potential vorticity curve, and less so for P18, P19 and S4P at a slightly lighter density (Fig. 9a). The upstream sections that is primarily associated with interaction of the density surfaces and the regional topography. We find no clear deflections in $\delta^3$He for the sections upstream of P16 (Fig. 9b), including at 28.2 kg m$^{-3}$ where we observe decreases in potential vorticity (Fig. 9a).

In contrast, the Pacific sections (P16, P18, P19 and S4P) show a clear elevation in $\delta^3$He (to ~10%) at 28.2 kg m$^{-3}$, followed by a sharp decline in $\delta^3$He at higher densities. The lower Indian $\delta^3$He signature (~8.5% near 28.05 kg m$^{-3}$) is derived from the Central Indian Ridge (see Fig. 11) and other tropical sources.

d. The Ross Gyre and Ross Sea

Recent high-resolution model- and data-based studies have indicated large eddy mixing and topographic roughness on the southern flank of the Pacific-Antarctic Ridge where the ACC and Ross Gyre interact (Thompson 2008; Lu and Speer 2010; Nikurashin and Ferrari 2011). Eddy mixing provides a means for the SPBW potential vorticity signals to be transported poleward from the Pacific–Antarctic Ridge, and across and between the ACC fronts. Recent data from the CLIVAR S4P cruise (Fig. 10) includes both repeated stations of the WOCE S4P zonal line along 67$^\circ$S, and three meridional sections in the Ross Gyre that are essentially poleward continuations of the P14, P15, and P16 lines (see Figs. 1, 5, and 8).

In the Central Ross Sea section, the SPBW potential vorticity signal is evident until it reaches the continental shelf around 74$^\circ$S, where deep waters are modified and then mix with shelf waters to form AABW (e.g., Orsi et al. 1999, 2002). The interannual variability in the properties of SPBW and Circumpolar Deep Waters and Ross Gyre strength are principal drivers in AABW production rates (Assmann and Timmermann 2005). From the Central and Eastern Ross Sea sections, we conclude that the SPBW originating at the Pacific-Antarctic Ridge flows poleward via the Ross Gyre. The modification of SPBW north of the continental shelf, and the formation of AABW on the shelf dilutes the potential vorticity signal (Western Ross Sea section; Fig. 10).

4. Discussion and conclusions

We use a combined PV and $\delta^3$He tracer analysis to identify the mechanisms modifying the properties and circulation of Southwest Pacific Bottom Water (SPBW) in the South Pacific and show the SPBW potential vorticity signature and large-scale pathways change minimally
over the past two decades. This provides a physical interpretation of the pathway of hydrothermal plumes originating from the Pacific–Antarctic Ridge, as described by Winckler et al. (2010).

Hydrothermal activity can be identified using several tracers, such as manganese, silicate, and germanium (e.g., Klinkhammer 1980; Mortlock et al. 1993). We analyzed silicate data available on South Pacific WOCE cruises; however, high silica also originates from other regions surrounding the Pacific–Antarctic Ridge, in the North Pacific, and Ross Gyre. Hence, the silicate found near the ridge vent sites is thus not necessarily indicative of plumes. Cholorfluorocarbons (CFCs) and oxygen have been repeatedly used to trace deep and bottom waters in the Southern Ocean (Orsi et al. 1999, 2002). However, LCDW and SPBW are both old, well mixed water masses with a low CFC signal, and are thus difficult to separate using CFCs. In the recent CLIVAR S4P Ross Sea sections, we identified distinct nutrient signals (oxygen, nitrate, and phosphate) that were evident but irrelevant to the SPBW potential vorticity signals (figure not shown).

Use of high resolution potential vorticity and helium data have provided a means of separating these deep water masses in the Pacific Ocean.

We also investigated the potential vorticity and $\delta^3$He along several other cruise lines around the global ocean and found no ridges where the two tracers were coincident (Fig. 11). The WOCE I03 line intersects the Central Indian Ridge at $65^\circ$E. Helium from the ridge flows northeastward and in a cyclonic path around the Ninety-East Ridge (e.g., Srinivasan et al. 2009; Drijfhout and Naveira Garabato 2008). We also observe the 15% $\delta^3$He signal at 8$^\circ$S on the I02 cruise line (figure not shown). Along the Juan de Fuca Ridge (40$^\circ$–45$^\circ$N; WOCE P17N), a $\delta^3$He signal of 24% can be observed; however, the spread of the plumes originating along this ridge are focused in their source region to within a degree or so of longitude (Veirs et al. 1999; Cannon and Pashinski 1997). In the South Pacific (WOCE P06; Fig. 11), hydrothermal plumes originating along the East Pacific Rise flow westward and counterclockwise over the ridge, then southeastward toward Drake Passage (Lupton 1998;
We observe the high $\delta^3$He signal (>24%) on both sides of the East Pacific Rise at 32°S (WOCE P06; Fig. 11). Johnson and Talley (1997) suggested that stratification and $\delta^3$He signals associated with hydrothermal plumes downstream of the East Pacific Rise are likely to coincide in the absence of vertical mixing. Along the Mid-Atlantic Ridge a distinct, but weaker, $\delta^3$He signal of ~6% has been observed along, and equatorward of, the WOCE A10 section (Fig. 11; Well et al. 2001; Rüth et al. 2000). Primordial helium (defined by anomalies relative to the background) plumes emanating from the Mid-Atlantic Ridge at 30° and 11°S spread across the western Atlantic basin (Rüth et al. 2000). In contrast, potential vorticity signals created above the ridge have been related to local mixing (Thurnherr et al. 2005). These examples illustrate the dependence of these two tracer signals on background fields as well as local processes.

After analyzing several cruise lines near other major ocean ridges, we infer that the SPBW potential vorticity coinciding with $\delta^3$He signals are likely unique to the Pacific–Antarctic Ridge. Several physical processes coincide along the Pacific–Antarctic Ridge to form the potential vorticity and $\delta^3$He signals and then transport them across the South Pacific (see Fig. 12):

- Hydrothermal vents along most of the ridge inject significant heat into the water column, destabilizing the stratification.
- The efficiency of eddy mixing is large and persistent along the entire Pacific Antarctic Ridge for the intermediate and deep ocean (Lu and Speer 2010).

FIG. 11. Four cruise lines that intersect prominent ridges with hydrothermal activity. Potential vorticity (color, $\times 10^{-12}$ m$^{-1}$ s$^{-1}$) overlaid with $\delta^3$He (black contours; %) for WOCE lines I03 (20°S), P17N (145°W), P06 (32°S), and A10 (30°S). Black dots denote sampling locations for $\delta^3$He.
The ACC overlies the vents along the Pacific–Antarctic Ridge. This sweeps them downstream, between the Polar Front and Southern Boundary. Strong westerlies driving the ACC induce a steep gradient in the density surfaces across the Pacific–Antarctic Ridge. Bottom waters originating in the Ross Sea flow equatorward to the ridge, carrying a low potential vorticity signal, and mix with lighter LCDW. The multiple hydrothermal sources along the Pacific–Antarctic Ridge (Winckler et al. 2010) provide abundant opportunity for deep waters to interact with the hydrothermal plume within the ACC along the 28.2 kg m\(^{-3}\) surface. Thus, different layers within the ACC can have vastly different origins.

Several questions remain unanswered. From the results presented here and in Winckler et al. (2010), it is evident that more detailed surveying of the Pacific–Antarctic Ridge and surroundings are required, to determine exact vent locations, source strengths and local spreading rates. Such high-resolution surveys have been performed extensively over ridges such as the Mid-Atlantic Ridge, Juan de Fuca Ridge, and East Pacific Rise (e.g., Lupton 1998; Rüth et al. 2000, and references therein), but would be valuable in this region, since decomposing the diapycnal and along isopycnal eddy mixing, deep water upwelling, hydrothermal activity, and ACC flow requires better information about the abyssal interactions with the midocean ridge.

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