Brazil Current volume transport variability during 2009-2015 from a long-term moored array at 34.5°S

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Key Points:

- The multiyear continuous absolute Brazil Current transport measured at 34.5°S has significant variability on daily to monthly time scales
- The baroclinic component accounts for the largest part of the absolute transport variance, but the barotropic variance is not negligible
- No meaningful seasonal cycle, interannual variability or trend is detected during the roughly six years of daily transport measurements

Abstract

The Brazil Current, the western limb of the subtropical gyre of the South Atlantic Ocean, is one of the major Western Boundary Currents of the global ocean. Here, we present the first multiyear continuous daily time series of Brazil Current absolute volume transport obtained using 6+ years of observations from a line of four pressure-recording inverted echo sounders (PIES) deployed at 34.5°S. The array was augmented in 2012 with two current meter-equipped PIES, and in 2013 with a moored Acoustic Doppler Current Profiler on the upper continental slope. The Brazil Current is bounded by the sea surface and the neutral density interface separating South Atlantic Central Water and Antarctic Intermediate Water, which is on average at a reference pressure of 628 ± 46 dbar, and it is confined west of 49.5°W. The Brazil Current has a mean strength of -14.0 ± 2.8 Sv (1 Sv ≡ 10^6 m^3 s^{-1}; negative indicates southward flow) with a temporal standard deviation of 8.8 Sv and peak-to-peak range from -41.7 to +20 Sv. About 80% of the absolute transport variance is concentrated at periods shorter than 150 days with a prominent peak at 100 days. The baroclinic component accounts for 85% of the absolute transport variance, but the barotropic variance is not negligible. The baroclinic and barotropic transports are uncorrelated, demonstrating the need to measure both transport components independently. Given the energetic high frequency transport variations, statistically significant seasonal to interannual variability and trends have yet to be detected.

Plain language summary

Western Boundary Currents are the most intense currents of the global ocean and are key to the redistribution of mass, heat, salt, and carbon throughout the globe. In the South Atlantic, the Brazil Current transports warm and salty waters off the South American coast towards the pole, and is a major driver of climate variability. This study presents, for the first time, multiyear continuous-in-time direct observations of the Brazil Current at 34.5°S. Roughly six years of daily measurements from moored sensors, together with high-resolution snapshots of temperature, salinity, oxygen and velocity collected during seven oceanographic cruises since 2009, provide the ability to characterize the daily to seasonal to year-to-year variability of the Brazil Current with unprecedented detail. These observations reveal strong and rapid changes in the Brazil Current on time scales as short as 14-60 days, allow quantification of the required sampling to resolve longer-term variability, and improve estimates of the Brazil Current transport. The Brazil Current variability is dominated by the east-west density variations in the water column although east-west differences in bottom pressure are not negligible. Understanding of the strength, structure, and time variability of the Brazil Current is needed to improve model representations of this important flow.

Key words: Brazil Current, meridional overturning circulation, volume transport, variability, transport array, observations

1 Introduction

The Brazil Current (BC) is one of the major Western Boundary Currents (WBCs). WBCs are essential components of the ocean circulation and understanding their short-term and long-term variability and the mechanisms that drive them is a fundamental element of oceanic and climate research (e.g., Imawaki et al., 2013; Archer et al., 2018; Todd et al., 2019; Palmer et al., 2019). While the existence of the BC has been known for nearly two centuries (e.g., Rennell, 1832), and the mean circulation patterns of the current have been understood for several decades (e.g., Stramma et al., 1990; Peterson & Stramma, 1991), there are still many open questions about the temporal variability of the BC and the mechanisms that drive those variations.

The southward flowing BC is the western limb of the subtropical gyre of the South Atlantic Ocean, carrying relatively warm and salty waters along the continental margin of eastern South America. The BC originates in the subtropics between roughly 10°S and 15°S where the westward flowing trans-Atlantic South Equatorial Current bifurcates (e.g., Stramma et al., 1990; Stramma, 1991; Rodrigues et al., 2007) in two branches: The northern branch forms the North Brazil Current and the southern branch forms the BC. After the bifurcation, the BC flows southward along the continental margin of South America, slowly increasing its southward time-mean transport from -0.8 Sv near 10-25°S to -70 Sv at roughly 35-38°S (negative transport indicates southward flow), attributed in part due to a tight anticyclonic recirculation cell in the southwest South Atlantic to the east of the mean BC flow in the upper 1400 m (e.g., Peterson and Stramma, 1990; Stramma et al., 1990). At 36-38°S, it encounters the northward flowing cold and fresh waters of the Malvinas Current, a northward branch of the Antarctic Circumpolar Current (e.g., Vivier and Provost, 1999; Spadone and Provost, 2009; Ferrari et al., 2017) and forms the Brazil-Malvinas Confluence (BMC), which meanders off towards the east (e.g., Gordon and Greengrove, 1986; Olson et al., 1988; Lumpkin & Garzoli, 2011). Along its southward path, the BC is located mostly over the upper continental margin, with mesoscale features impacting its mean flow and producing occasional separations from the shelf break (e.g., Campos et al., 2000; Calado et al., 2010; Soutelino et al., 2011; Lima et al., 2016). It has been reported that the BC separates from the shelf break for the last time at about 36°S (Olson et al, 1998; Goni et al., 2011).

Most of the available historical estimates of the southward BC transport are geostrophic relative to pre-defined levels of no motion and are based on hydrographic observations collected in the 1980's and 1990's between 12°S to 25°S (see Peterson and Stramma, 1991). Around 20°S the BC is located close to the continental shelf, with values around -6 Sv (e.g., Peterson and Stramma, 1990; Stramma et al., 1990). Historical estimates of BC transport near 23°S to 24°S have varied from -13.2 Sv to -4.1 Sv (Garfield, 1990; Campos et al., 1995; Stramma, 1989; Zemba, 1991). More recently, geostrophic estimates from historical ship data yielded a mean transport of -5.6 Sv above 500 m at 23-26°S (Biló et al., 2014). Near 33°S-35°S the historical hydrographic transports are typically around -24 to -18 Sv (Olson et al., 1988; Peterson and Stramma, 1991; Garzoli et al., 1993; Boebel et al., 1999).

A few historical continuous-in-time mooring records of the BC variability do exist, although in most cases the records are short (only a few months). These historical time series generally did not include the barotropic component of the flow. Müller et al. (1998) provided the sole historical transport time series observations to directly observe the flow over the Brazilian continental slope using nearly two-year long current-meter mooring arrays between 20°S and 28°S. They observed a southward deepening and strengthening of the BC, reaching down to 670 m at 28°S with a transport of -16 Sv west of 45°W. Several studies have quantified the baroclinic transport component using Inverted Echo Sounders (IES) between 35°S and 38°S, finding time-mean transports of -24 Sv to -10 Sv above 800 m over time periods of fifteen and eight months, respectively (Garzoli & Bianchi, 1987; Garzoli & Garrafo, 1989; Garzoli & Simionato, 1990; Confluence Principal Investigators, 1990; Garzoli, 1993). Although other studies based on repeated expendable bathythermograph (XBT) transects (e.g., Garzoli & Baringer, 2007; Garzoli et al., 2013; Lima et al., 2016; Goes et al., 2019), surface drifters (Oliveira et al., 2009), combined hydrographic data and quasi-isobaric subsurface floats (Rodrigues et al., 2010), satellite altimetry and temperature data (Goni et al, 2011), blended products with satellite altimetry and winds (Lumpkin & Garzoli, 2011), or on combinations of Argo floats and satellite altimetry (Schmid, 2014; Schmid & Majumder, 2018) have revealed some aspects of the BC time-mean and temporal variability, they were limited either in their spatial and/or temporal coverage. As a result, much remains unknown about the spatial structure of the BC and how it varies across a range of time scales.

There is great interest in observing the BC at 34.5°S, as at this latitude the BC is close to its transport maximum based on the limited observations to date at that latitude (e.g., Stramma 1989; Peterson & Stramma, 1991; Garzoli, 1993). Furthermore, on the large-scale context, this latitude marks the entrance to the subtropical South Atlantic, as it corresponds to the southern boundary of the African continent. For this reason, 34.5°S has been identified as a crucial latitude for examining the meridional transports of volume, heat, and salt due to the possibility that our current climate may be in a regime of multiple equilibria of the Atlantic Meridional Overturning Circulation (AMOC) and the need for observable transport metrics in the region (e.g., Dijkstra, 2007; Huisman et al., 2010; Drifjhout et al., 2011; Weijer et al., 2019). Observational and modeling results suggest that analysis of the boundary currents in the South Atlantic is essential to explore the local and remote forcing of AMOC fluctuations (e.g., Biastoch et al., 2008; Dong et al., 2009; Dong et al., 2011a,b; Rühs et al. 2015). While in the subtropical South Atlantic the warm upper limb of the AMOC primarily occurs at the eastern basin (e.g., Kersalé et al., 2019), the BC is unique as it carries mass and heat southward (ie. in opposite direction of the AMOC) and recent studies have shown that changes in the BC transport contribute to low frequency variability that has been observed in South Atlantic meridional volume and heat transports (e.g., Dong et al., 2009; Dong et al., 2015). It is therefore essential to accurately quantify the BC transports and variability to close the meridional basin-wide volume and heat transport budgets in the South Atlantic (e.g., Garzoli & Matano, 2011). Furthermore, the BC variability has often been poorly resolved and/or reproduced by different models (e.g., Palma et al., 2004) and there are many open questions about the response of the BC to intensifying Southern Hemisphere westerly winds under global warming (e.g., Yang et al., 2016, 2020; de Souza et al., 2019; Drouin et al., 2021), highlighting the importance of observing and understanding the characteristics and long-term variability of the BC.

In recent years, observations of the BC have increased as the AMOC observing network has expanded into-the South Atlantic (e.g., Garzoli et al., 2012; Meinen et al., 2013; Ansorge et al., 2014; Hummels et al., 2015; Meinen et al., 2018; Frajka-Williams et al., 2019; Kersalé et al., 2020; Herrford et al., 2021). One element of this AMOC observing network, the South Atlantic MOC Basin-wide Array (SAMBA) has been under development at 34.5°S since 2009, and is presently being used to observe daily variations in the AMOC (Meinen et al., 2013, 2018; Kersalé et al., 2020). The western boundary component of SAMBA (termed “SAMBA-West”) observes the flows of the BC in the upper ocean and the Deep Western Boundary Current (DWBC) below, and these observations provide the backbone of the present study between the continental shelf and 44.5ºW (Figure 1; Table 1). The initial studies using the SAMBA-West array data have concentrated their analysis on the deep limb of the flow carried by the DWBC and the abyssal flows (Meinen et al., 2012, 2017; Valla et al., 2019), and this is the first study focusing on the upper ocean currents on the western boundary.

The purpose of the present study is to examine the daily to interannual variability of the depth-integrated volume transport of the upper ocean, carried by the BC, by analyzing the SAMBA-West moored observations during a particularly well-observed segment of time between March 2009 and November 2015. These results will provide a better understanding of the strength, structure, and temporal variability of the upper branch of the WBCs in the South Atlantic by examining variations across a multitude of timescales with this unprecedented 6+ year data set.

2 Data and Methods

2.1 Overview of the SAMBA array in the western South Atlantic ("SAMBA-West")

This paper presents results from 6+ years of data from the SAMBA-West array between March 2009 and November 2015. The first component of the array was deployed in March 2009 and is still operating continuously (e.g., Meinen et al., 2012, 2017; Valla et al., 2018, 2019; Kersalé et al., 2020). The SAMBA-West array was designed to study the western boundary contributions to the AMOC, and it also represents the western cornerstone of the SAMBA array which measures the basin-wide integrated meridional volume and heat transport at 34.5°S (Meinen et al., 2013, 2018; Kersalé et al., 2020). The location of the array was chosen to be just north of the meander-window for the BMC, typically found around 38°S (e.g., Gordon & Greengrove, 1986; Goni et al., 1996, 2011; Lumpkin & Garzoli, 2011), to avoid the complicated variability of the confluence front and its energetic meanders (Perez et al., 2011; Meinen et al., 2012, 2017). Initially the array consisted of a line of three pressure-recording inverted echo sounders (PIES; Sites A, C and D) moorings and one current- and pressure-recording Inverted Echo Sounder (CPIES; Site B) mooring deployed along 34.5°S spanning about 640 km between the 1360-m isobath at 51.5°W (Site A; Figure 1) and the 4757-m isobath at 44.5°W (Site D; Figure 1). The initial CPIES was later replaced with a PIES in 2011, and in December 2012 two additional CPIES moorings were deployed between the westernmost two pairs of the existing PIES moorings (Sites AA and BB) to improve the horizontal resolution of the array (e.g., Meinen et al., 2017). Furthermore, in December 2013 a bottom-mounted upward looking Acoustic Doppler Current Profiler (ADCP) mooring was deployed to capture meridional flows on the upper continental slope, nominally at the 411-m isobath (Figure 1, Table 1). Note that the first CPIES deployed at Site B functioned improperly in 2010 and was lost during a recovery attempt in 2011, and as a result there is a gap of about two months at the start of the first deployment at that site and of nearly 15 months in 2010-2011 at the end of that first deployment. The nominal positions, water depths and deployment dates of the PIES, CPIES and ADCP moorings are given in Table 1 (see also Figure 1).

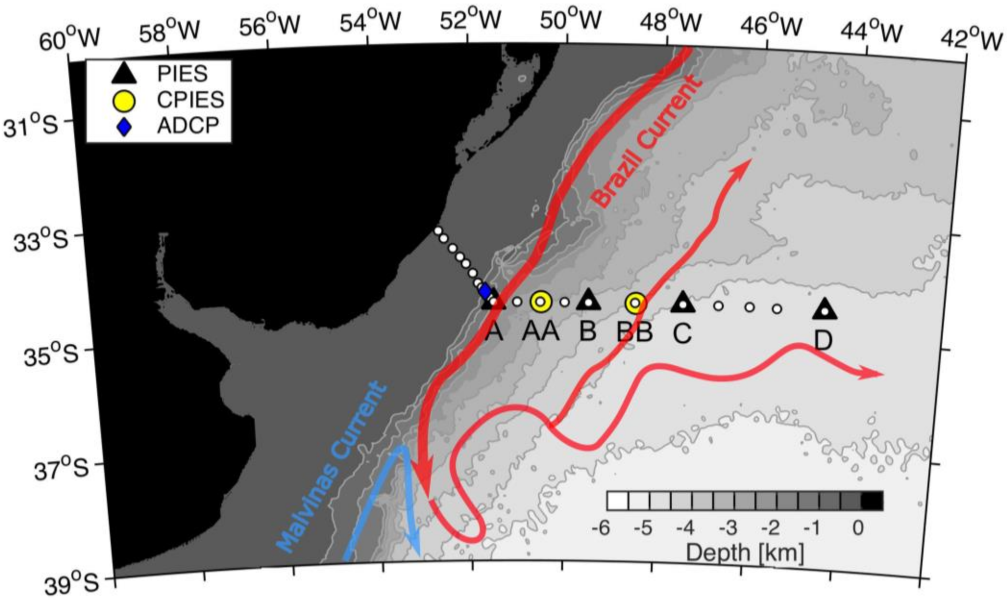

Figure 1. Configuration of the SAMBA-West array: PIES (black triangles), CPIES (yellow circles), ADCP mooring (blue diamond). PIES/CPIES site names on the transport line are displayed. White circles represent the nominal positions of the hydrographic CTD/O_2/LADCP stations (Table 2). Bathymetry (shaded background) comes from the Smith & Sandwell (2007) data set. The arrows on top represent the surface and sub-surface circulation of the Brazil Current (red) and Malvinas Current (light blue) based on the climatological time-mean dynamic topography (1993-2012) extracted from https://www.aviso.altimetry.fr/.

Table 1. Nominal positions, water depths and initial deployment dates of the instruments distributed across the SAMBA-West transport line at 34.5°S.

Description	Site Name	Longitude (West)	Latitude (South)	Water Depth [m]	Date of first deployment
ADCP		51°40.0’	34°19.0’	411	14 December 2013
PIES	A	51°30.0’	34°30.0’	1360	18 March 2009
CPIES	AA	50°31.2’	34°30.0’	2885	11 December 2012
PIES[a]	B	49°30.0’	34°30.0’	3535	18 March 2009
CPIES	BB	48°30.5’	34°30.0’	4140	12 December 2012
PIES	C	47°30.0’	34°30.0’	4540	19 March 2009
PIES	D	44°30.0’	34°30.0’	4757	20 March 2009

[a] The original CPIES deployed at Site B transmitted data poorly and/or incompletely during download cruises in August 2009, July 2010, and December 2010, so it was decided that it would be replaced with a new PIES during the next cruise in July 2011. The CPIES was lost during recovery, and as a result there is a gap of about two months at the start of the deployment and nearly 15 months (21March 2010-10 July 2011) at the end of the first deployment at that site.

Between 2009 and 2015 seven cruises in support of SAMBA-West have been conducted roughly every six to twelve months (Table 2, labelled “SAM” followed by cruise number). During each cruise, conductivity, temperature and depth (CTD) stations were occupied along the SAMBA-West line (white circles in Figure 1, Table 2). The SAMBA-West hydrographic sections, which are typically collected over 4-6 days, have a typical horizontal resolution of 20 km at the inner shelf and shelf-break and 45-90 km offshore. In addition to the basic CTD measurements, some of the cruises also collected quasi-continuous dissolved oxygen (O_2) profiles and lowered ADCP (LADCP) measurements of full-depth velocities (data processing is described in Valla et al., 2018, 2019). These hydrographic sections provide independent transport estimates as we will see in section 3.2, which can be compared with the PIES/CPIES derived transports. Three of the SAMOC-West cruises also occupied additional stations inshore of the westernmost PIES (Site A) on the continental shelf and shelf-break, providing an essential mean transport estimate inshore of the westernmost moored PIES at Site A, as will be shown in section 2.2.1. An additional shelf/upper slope CTD/LADCP section was taken in 2013 as part of the SubTropical Shelf Front (STSF) program (Berden et al., 2020; Charo et al., 2020a,b). Data from the high-density XBT transect AX18, which nominally crosses the South Atlantic along 35°S between South America and South Africa quasi-quarterly since 2002 (e.g., Garzoli & Baringer, 2007; Dong et al., 2009; Garzoli et al., 2013), are also used to compare against the baroclinic geostrophic estimates from the PIES/CPIES array and the CTD sections. Processing for PIES/CPIES and ADCP measurements is explained in the following subsections.

Table 2: Hydrography (CTD), dissolved oxygen (O_2), and LADCP data used in this study.

Cruise	Dates	Data type	Longitude range [West]
SAM02	17-25 August 2009	CTD	51°29.0'; 44°28.0'
SAM03	5-16 July 2010	CTD/O_2	51°29.0'; 44°27.0'
SAM04	20-29 December 2010	CTD	51°29.0'; 44°27.0'
SAM05[a]	2-12 July 2011	CTD/LADCP	51°40.0'; 44°31.0'
SAM07	2-12 July 2012	CTD/O_2/LADCP	52°50.0'; 44°31.0'
SAM08	1-16 December 2012	CTD/O_2	51°38.0'; 44°29.0'
SAM10[a]	4-16 October 2014	CTD/LADCP	51°31.0'; 44°31.0'
STSF	7-8 October 2013	CTD/LADCP	51°20.0'; 51°30.0'

Note. [a]LADCP section incomplete

2.1.1 PIES data processing and gravest empirical mode

The IES has been in use as an oceanographic tool for about 50 years (e.g., Rossby, 1969; Watts and Rossby, 1977). In its basic form the IES measures the travel time τ for a sound pulse (originally 10 kHz, since the year 2000 generally 12 kHz) to travel from the sea bottom to the surface and back. The basic processing of the raw hourly IES τ measurements has been well established for many years and are described in detail elsewhere (e.g., Tracey et al., 1997; Donohue et al., 2010; Chidichimo et al., 2014). Early applications of the IES in the Southwestern Atlantic utilized hydrography-derived relationships between dynamic height anomaly integrated between the surface and a fixed pressure (often 300 dbar) and simulated travel time to estimate the baroclinic transport of the upper ocean currents relative to an assumed level of no motion (e.g., Garzoli & Bianchi, 1987; Garzoli, 1993). In the late 1980s and early 1990s, a variant instrument integrating a bottom pressure sensor was developed (PIES; e.g. Watts et al., 1995). The processing of the raw hourly bottom pressure data, which includes a "response analysis" tidal correction following Munk and Cartwright (1966), has been presented in detail elsewhere and will not be repeated here (e.g., Watts and Kontoyiannis, 1990; Donohue et al. 2010). Later in the early 2000s a version of the instrument adding a single-depth acoustic current meter 50 m above the bottom was also developed (CPIES; e.g., Donohue et al., 2010; Bishop et al. 2012; Green et al. 2012). In essence the aforementioned papers illustrate how the hourly pressure and τ measurements are processed to produce a single daily value of τ and pressure at noon GMT each day. Subsequently, a 72-h low-pass filter is applied to the τ and pressure records at each site to remove tides.

Modern analysis of τ data involves the use of two-dimensional look-up tables that are created via the Gravest Empirical Mode (GEM) technique developed by Meinen and Watts (2000). The details of how τ data and bottom pressure measurements of a PIES (or CPIES) are combined with historical hydrography from the region via the use of the use of two-dimensional look-up tables that are created via the GEM technique to produce daily estimates of relative geostrophic velocity at the SAMBA-West array are presented in Meinen et al. (2012, 2013, 2017) and will not be repeated here. In brief, the GEM look–up tables quantify temperature (*T*), salinity (*S*), and/or specific volume anomaly (δ) as functions of pressure and τ calculated from hydrographic casts between the surface and a selected reference pressure (1000 dbar herein). The daily PIES-GEM estimated profiles are gridded with uniform 20 dbar pressure increments from the surface to the seafloor. The PIES-GEM δ profiles are then vertically integrated to give daily dynamic height anomaly (ϕ) profiles relative to an assumed level of no motion. Differencing the ϕ profiles at neighbouring PIES/CPIES sites yields daily full water column profiles of the component of the relative geostrophic velocity perpendicular to the line between the sites (e.g., meridional) via the standard geostrophic method (e.g., Gill, 1982). Similarly, differencing the bottom pressure measurements at neighbouring PIES/CPIES yields the temporal anomaly of the absolute geostrophic bottom velocity. However, the time mean absolute geostrophic bottom velocity cannot be determined due to the well-known leveling problem (Donohue et al. 2010; Meinen et al., 2012).

The products from a high-resolution numerical model are utilized to provide estimates of the time-mean bottom or reference velocities (the variability of the reference velocity comes from the bottom pressure observations). We use the time-average of 35 years of 3-day snapshots from the OGCM for the Earth Simulator (OFES) with 0.1° horizontal resolution and more than

50 vertical levels, provided by the Japan Agency for Marine-Earth Science and Technology (JAMSTEC) (e.g., Sasaki et al., 2008). Similar to previous studies in the region using data from the SAMBA-West array that have also selected this particular model run, the time-mean reference velocity was computed at 1500 dbar (Meinen et al., 2013, 2017, 2018; Valla et al., 2019).

Herein the time-mean meridional reference velocities used from the 35-year run from the OFES fields between 1980 and 2015 are -5.6, -6.0, 0.7, 0.7, and -0.2, cm s^{-1} between site pairs A-AA, AA-B, B-BB, BB-C, and C-D, respectively and -5.8 and 0.7 cm s^{-1} between site pairs A-B and B-C, respectively. Note, our results are not sensitive to using mean OFES 1500 dbar velocities for the same period as the observations (2009-2015) vs. the 35-year mean that we used in this study.

Throughout this paper, we use the term “baroclinic” to refer to the transport estimated from geostrophic velocity shears referenced to zero at the bottom, and the term “barotropic” to refer to the transport contribution from the bottom-reference velocity (non-sheared term). The “absolute” transport is the sum of the baroclinic and the barotropic terms. Note that all of the time variability associated with the baroclinic and barotropic terms originate from the PIESs travel time τ and bottom pressure measurements, respectively, and hence are independent from the model time-mean reference velocities. The time-mean of the vertically-sheared (baroclinic) term is also purely observational and independent from the model. The resulting time-mean transports are not hugely sensitive to the choice of the model to compute the time-mean meridional reference velocities (as shown in Meinen et al., 2017).

2.1.2 ADCP data processing

The data from the ADCP mooring deployed at 411 m depth on the upper continental slope in December 2013 (Figure 1, Table 1) were processed following standard procedures (e.g., Côté et al., 2011). The 150 kHz ADCP recorded hourly velocity on vertical bins of 16 m, with bins centered at depths between 23.73 and 343.73 m. The hourly data were averaged to obtain daily values, and subsequently a 72-h low-pass filter was applied to remove tides.

2.2 Absolute Brazil Current transport calculation

The calculation of a boundary current transport requires a choice of the vertical and zonal limits of integration. Along the SAMBA-West array, the neutral density surface (γ^n) of 27.1 kg m^3 separates the warm, salty, relatively low in oxygen Tropical Water (TW) and South Atlantic Central Water (SACW) from the relatively cold-fresh oxygen-rich Antarctic Intermediate Water (AAIW) located below (Valla et al. 2018; Figure 2a, b, c, d). Thus to obtain the bulk of the BC transport, we calculate transports in the depth layers occupied by the TW and SACW by evaluating the γ^n structure (Jackett and McDougall, 1997) across the hydrographic sections (Figure 2a, b, c; Table 2). We find the interface between SACW and AAIW on average at the 628 ± 46 dbar pressure level (variability represents one standard deviation; the shallowest and deepest positions observed during cruises are at 533 and 667 dbar, respectively), and we define this temporally-fixed lower bound for integrating the velocities to obtain transport. The resulting transport time series are not significantly different for modest (±100 dbar) changes to the 628 dbar pressure level (correlation coefficient of 0.99 and root-mean-square error of 1.5 Sv). Note that at this latitude the western boundary flow is mostly southward over the entire water column so it is not possible to establish more quantitative, time-varying, criteria to separate the BC from the DWBC, such as a zero-velocity crossing. For similar reasons, Meinen et al. (2017) chose 800 dbar as the upper bound to obtain the DWBC flow.

The definition of the horizontal limits of integration to estimate transports from in situ arrays is also not straightforward, as they strongly depend on the array's spatial resolution as well as on eddies/meanders passing by the array's laterally separated sites at each time step. Defining a static box within which a “total” (or “absolute”) BC transport will be calculated will certainly include transient transports from Rossby Waves and other features, which propagate into the integration domain. Even utilizing time-varying integration boundaries based on characteristics of the ocean flow, such as stream-coordinates, will still incorporate these other features when they superimpose on top of the main current (e.g., Meinen & Luther, 2016). As a result, the measured southward “Brazil Current” transport will fluctuate higher or lower, potentially reaching zero or even flowing northwards due to the influence of large eddies and other features (e.g., Meinen et al. 2017). For this study, the BC transport will be calculated within fixed boundaries determined from the available LADCP ship section (Figure 2e, Table 2), which as we will see later agree well with the PIES/CPIES time-mean velocity section.

The full-depth spatial structure of alongshore flow from the LADCP section (Figure2e) evidences southward flow between the shelf break and Site B (at 49.5°W) from the surface to the bottom, with the BC core at the ~100-400 dbar range near the continental slope. Offshore of Site B there is a northward recirculation. Thus, an integration domain between the coastline and Site B, and between the surface and 628 dbar (i.e., the γ^n = 27.1 kg m^3 surface) is selected. We will come back to the analysis of the horizontal extent of the BC in the following sections. The "absolute" BC transport represents the sum of the baroclinic and barotropic velocities integrated over this domain.

Note, we find that the wind-forced Ekman transport contribution to the BC in this region is very weak, and it has a very small temporal mean and standard deviation of -0.002 ± 0.225 Sv during 2009-2015 (not shown). Additionally, the SAMBA-West array location is close to the zero value of the wind stress curl (e.g., Schmid and Majumder, 2018). Thus, we focus on the geostrophic component of the BC transport in this study.

To obtain the total flow above 628 dbar associated with the BC, we compute transports through two cross-sectional areas and afterwards sum them: i) over the upper continental slope inshore of the 1360 m isobath between the coast and Site A (section 2.2.1); and ii) between Site A (1360 m isobath) and Site B (3535 m isobath) (section 2.2.2).

Note that the full details on the estimation of the total transport accuracy is provided in Appendix A. Throughout this paper and unless otherwise noted the reported velocity/transport/pressure variability represents one standard deviation from the time-mean and negative velocities/transports indicate southward flow.

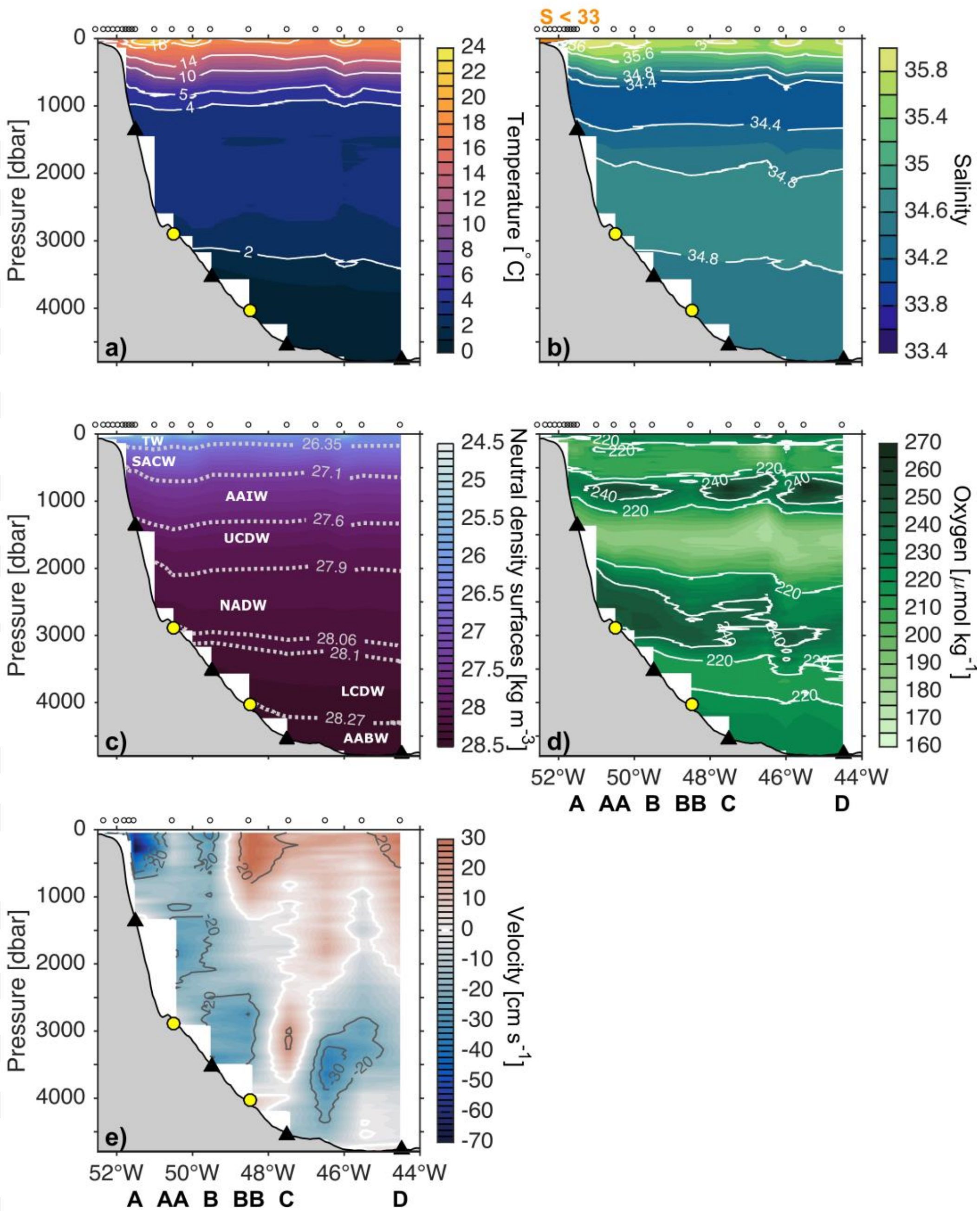

Figure 2. Average high-resolution hydrographic sections along the SAMBA-West line of (a) in situ temperature (°C), (b) salinity and (c) neutral density surfaces (γ^n ; kg m^{-3}) during four hydrographic cruises (SAM05, SAM07, SAM08, and SAM10; Table 2), (d) dissolved oxygen (µmol kg^{-1}) during SAM07 and SAM08, and (e) alongshore velocity (cm s^{-1}) derived from LADCP velocities collected during SAM07 cruise during July 2012 (white contours indicate zero flow; negative velocities, blue shading, indicate southward flow). A 50 m moving average is

applied in the vertical to the alongshore velocity profiles for stations deeper than 400 m in order to reduce noise in the LADCP data. Acronyms in panel c) indicate the key water masses after Valla et al. (2018): TW: Tropical Water; SACW: South Atlantic Central Water; AAIW: Antarctic Intermediate Water; UCDW: Upper Circumpolar Deep Water; NADW: North Atlantic Deep Water; LCDW: Lower Circumpolar Deep Water; AABW: Antarctic Bottom Water. White circles on top of the upper axis represent the nominal positions of the hydrographic CTD/O_2 stations (identical to Figure 1) in panels (a, b, c, d) and the positions of the hydrographic LADCP stations during SAM07 in panel (e). Gray shading is the bathymetry from the Smith & Sandwell (2007) data set. PIES and CPIES locations are indicated by the black triangles and yellow circles, respectively, on top of bathymetry of each panel. Labels at the bottom of panels d) and e) indicate the names of the PIES/CPIES sites.

2.2.1 Transport on the shelf and upper continental slope

Due to the more limited data inshore of Site A, several different approaches were applied to estimate the transport within that wedge. A preliminary first estimate of the time-mean flow inshore of Site A was determined by evaluating the alongshore transport from the three snapshot hydrographic sections that included absolute velocity measurements with an LADCP between the 400 m isobath and Site A. Integrating the LADCP profiles from three cruises over a distance of 29 km between the 400-m isobath and Site A (1360 m isobath) and from the surface to 622 m depth gives a mean and standard deviation of the transport of -2.4 ± 0.8 Sv. A second estimate of -2.7 ± 1.5 Sv was determined using all CTD data available (-3.0 ± 1.8 Sv for the three cruises that also collected LADCP observations) to estimate the velocity shear between the surface and the γ^n = 27.1 kg m^3 surface relative to the deepest common level (i.e., the bottom) between stations from each of the cruises that sampled between the 400 m isobath and Site A (Table 2; Figure 3b). This estimate is in good agreement with the LADCP mean estimate. The vertical structure of the CTD and the spatially-averaged LADCP shear profiles between station pairs also agree fairly well for three cruises with both types of data (see for example light green dotted and solid lines in Figure 3b).

Analyses of the velocity data collected with the moored ADCP at the 411-m isobath, only available between December 2013 and November 2015, provides further insights about the shelf and upper slope velocities. The time-mean flows from the ADCP (Figure 3a, blue line) are weak compared to the snapshot CTD relative velocities and the LADCP absolute velocities that are found for the stations at the 500-1000-m isobath range (Figure 3b). The standard deviations of the moored ADCP velocities are quite large (Figure 3a, blue shading), however, which suggest large meridional velocity variability at this location consistent with the large spread of velocities observed during each cruise (shallow profiles on top of blue profile, Figure 3a). The mean moored ADCP velocities instead agree well in magnitude with the CTD relative velocities (solid red, light green, magenta, light blue, and orange lines) and LADCP velocities (not shown) up on the shelf for depths shallower than 200 m, which are very likely unrelated to the BC. We note that high-resolution numerical models have previously estimated weak seasonal reversing flows of ~-0.4 to 0.3 Sv on the continental shelf inshore the 500 m isobath near 32°S (Palma et al., 2008).

The relatively small transports in the western wedge that are not sampled by the PIES/CPIES array (-2.4 Sv from LADCP; -2.7 Sv from CTD) agree fairly well with previously published estimates based on long-runs of the OFES model (-3.0 ± 1.6 Sv) and the Nucleus for European Modeling of the Ocean (NEMO) model (-4.6 ± 3.3 Sv) as well as from 18 XBT sections (-2.1 ± 2.5 Sv) in Meinen et al. (2013). The limited cruise snapshot data together with the moored ADCP data suggest that the largest portion of the flow not presently sampled by the PIES/CPIES array is concentrated within the ~25 km immediately west of Site A between the 500 m and 1000 m isobaths (Figure 3b), with large southward velocities reaching up to nearly -100 to -40 cm s^{-1} in the upper 400 m of the water column during individual cruises. Figure 3b also demonstrates that the vertical structure of the velocity profiles inshore of Site A has a very "baroclinic" structure which clearly varies over time. Although a small number of snapshots are not representative of the multi-year long-term mean wedge transport, the best available option for estimating the transport within this inshore wedge is to use the average of the cruise section data. Because there are more CTD sections than LADCP sections, we have elected to use the mean CTD estimate (-2.7 Sv) on the shelf-break.

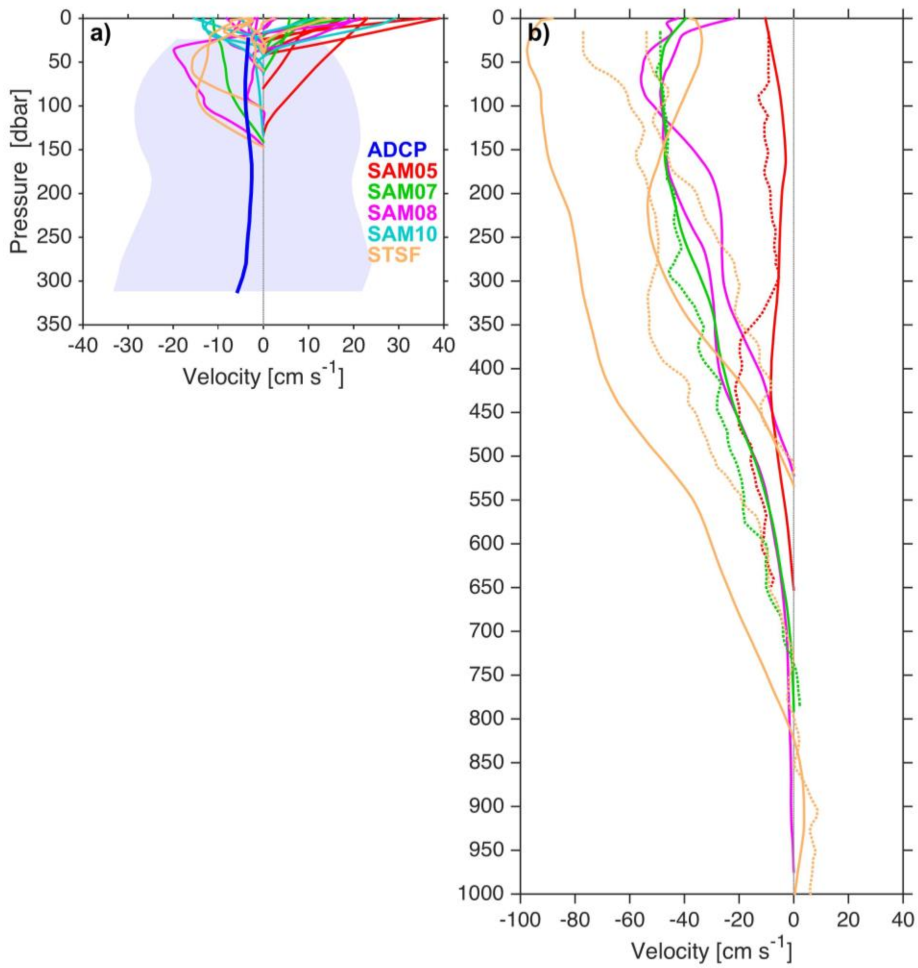

Figure 3. Alongshore velocity on the shelf and continental slope inshore of Site A. a) Time-mean velocity profile from the moored ADCP (blue line) and standard deviation (blue shading) are shown as well as the geostrophic velocity profiles from CTD station pairs (solid red, light green, magenta, light blue, and orange lines) for stations shallower than 400 m from the cruises indicated in the legend. Note the absolute velocity profiles from LADCP in this region are not shown for clarity. b) Geostrophic velocity profiles from CTD station pairs (solid red, light green, magenta, and orange lines) and spatially-averaged absolute velocity profiles from LADCP between station pairs when available (dotted red, light green, and orange lines) for sites between the 500 m isobath and Site A (at 1360 m) from the cruises indicated in the legend in a). No CTD/LADCP profiles are available in this region during SAM10. Negative velocities indicate southward flow. Note, the different velocity scale on the x-axes for a) and b).

2.2.2 Transport from PIES/CPIES

The transport within the PIES/CPIES array was integrated between pairs of moorings following well-established methods (e.g., Meinen et al. 2017, and references therein; section 2.1.1). Based on the meridional velocity section in Figure 2e the southward flow of the BC Current is concentrated west of 49.5°W. Sites A (at 51.5°W) and B (at 49.5°W) constitute the primary endpoints for this calculation because the records at those two sites span the longest time period (2009-2015).

Recall that in December 2012 two CPIES were deployed midway between Sites A and B (Site AA) and between Sites B and C (Site BB) (Figure 1, Table 1). Geostrophic flow is naturally integrating, so including an additional site between two neighbouring sites does not change the geostrophic flow integrated between the two original sites. In practice, however, transports determined using the central site or excluding the central site will not be identical in the presence of sloping bottom topography. Furthermore, the use of additional sites can reduce the size of unobserved bottom triangles along that slope. In order to test the impact of the increased horizontal resolution between Sites A and B during 2013-2015, we estimate transports between Sites A and B during those three years including Site AA. There is no significant change in the mean transport and its variability compared with the estimate excluding AA (correlation coefficient of 0.98 and root-mean-square error of 1.0 Sv) indicating that the velocity in the unobserved bottom triangles when using only Sites A and B for the calculation are very small. To test the transport contributions between Sites B and BB, we evaluate the transport from B to BB for each time step; interestingly the time-mean absolute transport between these two sites was positive (northward) during 2013-2015 (4.9 ± 7.6 Sv), which decreases the BC southward transport if included. The resulting time-mean absolute transport between Site B and Site BB in the BC depth layer when only the periods of southward flow in that span are averaged (with the condition that transports between Sites A and B are also southward) is only -1.0 ± 1.6 Sv, and would increase the southward BC transport only very slightly if included. Because Site BB is not always available during 2009-2015, we decided to only use Sites A and B as endpoints for our calculation recognizing that we may occasionally miss a minimal fraction of the BC transport offshore of Site B.

We determine the absolute transport BC transport as follows. First, we add a constant (time-mean) slope transport of -2.7 Sv from the average of the CTD sections (section 2.2.1), to the time-varying baroclinic PIES transport time series referenced to zero at the bottom (section 2.1.1), hereafter referred to as the baroclinic component of the BC transport. Next, we estimate the barotropic BC transport component as described in section 2.1.1. Finally, we combine the baroclinic and barotropic components to obtain the absolute BC transport. The resulting transport time series will be presented and discussed in the following section.

3 Results and Discussion

3.1 Brazil Current absolute velocity and transport variability

The spatial structure of the time-mean absolute velocity section from the PIES/CPIES data shows southward flow from the surface to the ocean bottom between Sites A and B (Figure 4a) with peak southward speeds reaching -12 cm s^{-1} near the surface. This southward flow includes both the BC above ~628 m (based on the density interface definition), and the DWBC below the BC base; as in previous studies with the SAMBA-West array, there is no obvious indication in the time-mean velocity field as to where one current ends and the other current begins (Meinen et al., 2012, 2017). The BC is concentrated west of 49.5°W (Figure 4a, e) in agreement with the hydrographic section (Figure 2e). When CPIES records at AA and BB are included, the peak speeds within the BC core are faster (compare Figures 4c and 4e) by approximately 4 cm s^{-1} between Sites A and AA, due to less horizontal averaging, but the integrated transports are very similar as is to be expected in a geostrophic calculation when velocity at the unresolved bottom triangles is small (section 2.2.2). Offshore of Site B (at 49.5°W) an evident northward recirculation can be observed in the full water column extending to 46°W, and further east of Site C the flow becomes southward again (Figure 4a, c, e) as also observed by Meinen et al. (2017) for the period 2009-2014. The temporal standard deviation of the absolute velocity profiles (Figure 4b, d, f) display large variability across the array from the surface to the bottom, with the largest variability between the surface and about 1000 dbar. When including AA and BB in the calculation, the temporal standard deviation reaches nearly 16 cm s^{-1} at the BC core (comparable to the northward recirculation region) (Figure 4f). While the standard deviation provides evidence of strong surface intensification, the subsurface variability of the BC is also substantial (e.g., Figure 4b) and it is further enhanced with the addition of data from two more moorings (Figure 4f).

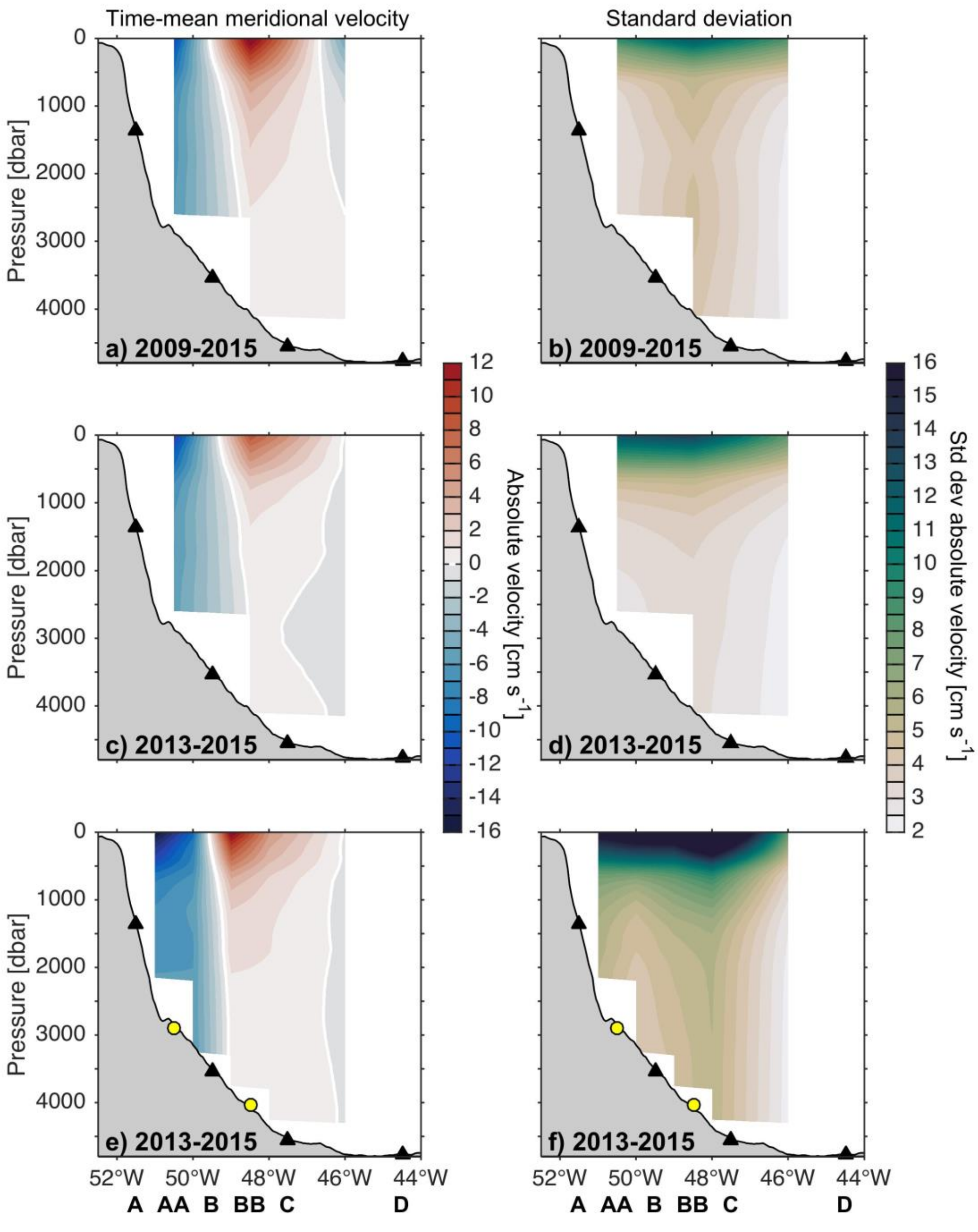

Figure 4: Time mean absolute meridional velocity between PIES/CPIES station pairs along the SAMBA-West array (left panels) and temporal standard deviation (std dev) of the absolute meridional velocity (right panels). The time-mean and temporal standard deviation of the

absolute meridional velocity are computed for (a, b) the full period of PIES measurements (2009-2015), (c, d) for the period when the horizontal resolution of the array was augmented with CPIES AA and BB but only using the PIES data (2013-2015), and (e, f) for the period when the horizontal resolution of the array was augmented with CPIES AA and BB using all available data (at Sites A, AA, B, BB, C and D) (2013-2015). White contours in left panels indicate zero flow. Negative velocity indicates southward flow. Labels at the bottom of panels e) and f) indicate the names of the PIES/CPIES sites.

The absolute, baroclinic, and barotropic BC transport records during 2009-2015 exhibit means and standard deviations of -14.0 ± 8.8 Sv, -7.4 ± 7.6 Sv, and -6.6 ± 3.1 Sv, respectively (Figure 5; Table 3). The Standard Error of the Mean (SEM) of the absolute transport is 1.1 Sv based on the estimated 64 degrees of freedom in the record (Appendix A). The three time series all have large temporal variability with respect to their time-mean value, consistent with what has been observed in the DWBC below/offshore of the BC (Meinen et al., 2017). There are several times when the absolute transport (black curve in Figure 5) is positive (northward) or close to zero, e.g., for the periods centered on 9 March 2010, 21 September 2012, 5 February 2014, 12 October 2014 and 31 May 2015. These periods of positive (or close to zero) absolute transport are mainly due to changes in sign of the baroclinic velocity component (compare black and red curves in Figure 5). The absolute transport time series exhibits a peak-to-peak transport range of 61.7 Sv, with maximum southward flow of -41.7 Sv on 6 December 2014, and maximum northward flow of 20 Sv on 12 October 2014. There is large short-term variability in the record, with changes of up to 20-30 Sv occurring over periods as short as 2-3 weeks, for instance between 24 November and 11 December 2009, 2 September and 22 September 2012, 15 June and 9 July 2013, 20 January and 4 February 2014, and 17 September and 7 October 2014 (black curve in Figure 5). Transport variations from peak-to-peak of ~40-45 Sv are observed over periods of 30-60 days, e.g., between 13 December 2013 and 4 February 2014, and 7 September and 12 October 2014, and even larger peak-to-peak changes of ~50-60 Sv are observed between mid October 2014 and early December 2014.

The baroclinic transport also exhibits a large peak-to-peak transport range of 50 Sv, with similar rapid changes observed over short periods (red curve in Figure 5). The barotropic transport peak-to-peak range of 24.2 Sv is smaller than for the baroclinic record but it is still quite large. The barotropic transport also has fast changes of approximately 15 to 20 Sv during short periods of time, for instance between 12 December 2012 and 2 January 2013, and 12 September 2014 and 7 October 2014 (green curve in Figure 5). As with the peak-to-peak ranges, the standard deviation of the barotropic transport is smaller than that of the baroclinic transport (3.1 Sv vs. 7.6 Sv, respectively). The large amplitude fluctuations in the three records persist even after 30-day low-pass-filtering the records using a sixth order Butterworth filter applied forward and backward to avoid phase shifts (Emery and Thomson, 2001; gray curves superimposed on black, red and green curves, Figure 5).

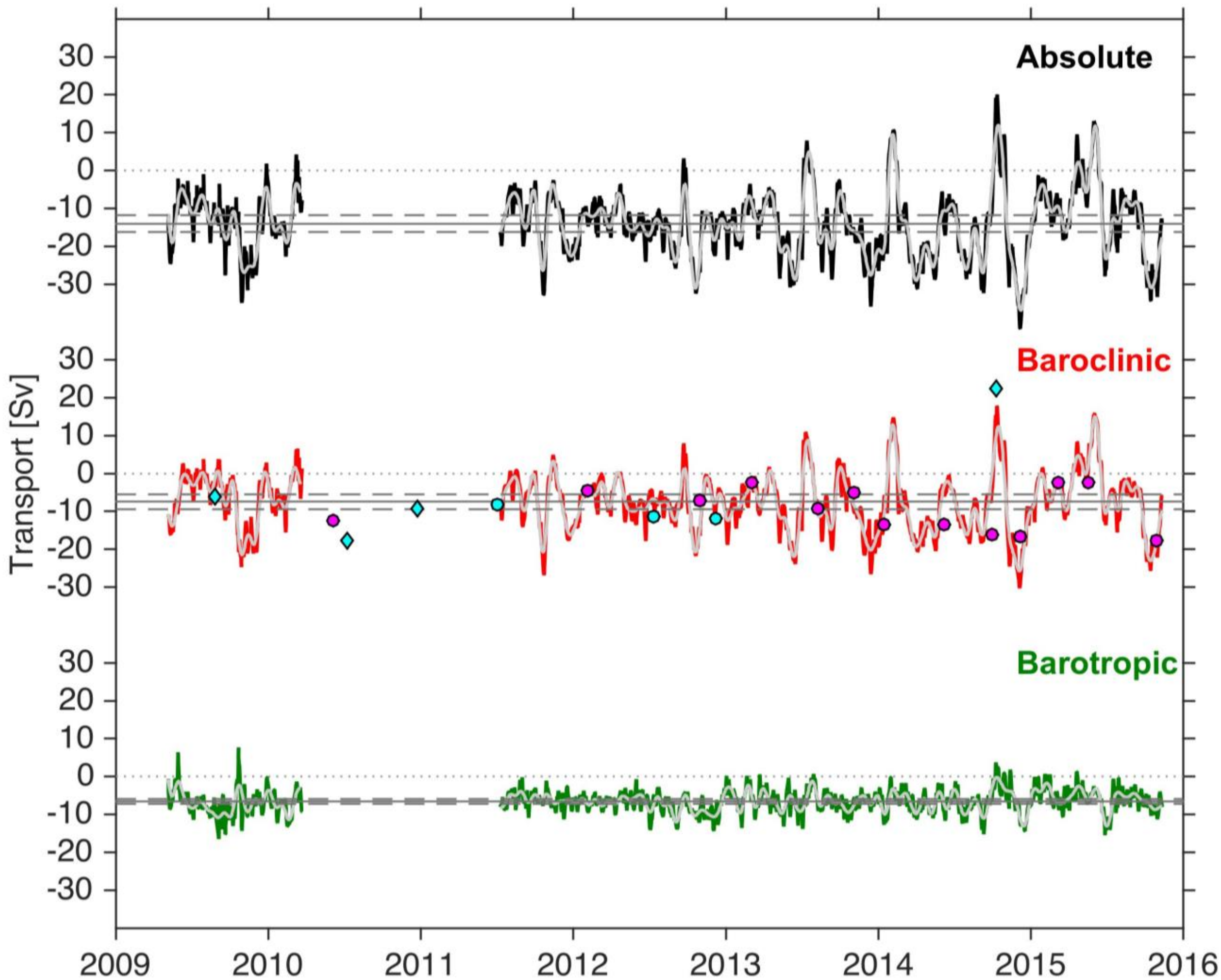

Figure 5. Time series of 72-h low-pass-filtered absolute (black), baroclinic (red), and barotropic (green) Brazil Current transport observed at 34.5°S during 2009-2015. The thin gray lines overlaid on top of each time series are the 30-day low-pass-filtered records. The gray horizontal solid and dashed lines indicate the time-mean and time-mean plus minus two standard errors of the mean (2*SEM; 95% confidence value), respectively. The cyan symbols indicate baroclinic

transports estimated from concurrent hydrographic transects during the SAMBA-West cruises (circles and diamonds represent cruises with and without data inshore Site A, respectively; Table 2) and the magenta circles indicate the baroclinic transports computed using the measurements at the AX18 XBT line nominally at 35°S. Negative transports correspond to southward flow.

Table 3. Basic statistics of the 72-h low-pass filtered absolute, baroclinic, and barotropic Brazil Current transport between 2009 and 2015.

	Mean ± Standard Deviation [Sv]							
	2009	**2010**	**2011**	**2012**	**2013**	**2014**	**2015**	**Average 2009-2015**
Absolute	-14.0 ± 7.4	(-12.1 ± 6.0)	(-12.8 ± 6.9)	-14.7 ± 5.7	-14.8 ± 8.0	-16.2 ± 12.0	-10.8 ± 9.8	-14.0 ± 8.8
Baroclinic	-7.0 ± 7.0	(-4.7 ± 4.5)	(-6.9 ± 6.7)	-7.4 ± 4.8	-8.5 ± 7.4	-9.5 ± 9.7	-5.1 ± 8.2	-7.4 ± 7.6
Barotropic	-7.0 ± 3.7	(-7.4 ± 3.0)	(-5.9 ± 2.1)	-7.3 ± 2.5	-6.3 ± 2.7	-6.7 ± 3.5	-5.7 ± 3.0	-6.6 ± 3.1

	Median [Sv]	**Min [Sv]**	**Max [Sv]**	**ITS [days]/ DOF**	**SEM [Sv]**
Absolute	-13.6	-41.7	20.0	12/64	1.1
Baroclinic	-7.0	-30.2	17.8	12/64	1.0
Barotropic	-6.5	-16.5	7.7	63/125	0.3

Note. (top) The mean and standard deviation of annual and average transports are given in columns 2-9. The mean annual values for years with less than eight months are given in brackets (years 2010 and 2011 have three and six months of available data, respectively). (bottom) The median, minimum and maximum values are given in columns 2-4. The integral time scale (ITS, days) and the degrees of freedom (DOF) are given in column 5 and the standard error of the mean (SEM) is given in column 6 (see Appendix A).

The correlation coefficient (r) between the baroclinic and absolute transport is high (r = 0.92; significant with 99% confidence, Appendix A). In contrast, the correlation between the baroclinic and barotropic components of the transport is quite low (r = 0.31) but significant with 95% confidence, and increases only slightly after applying a 30-day low-pass filter to both records to remove the high frequency signals (r = 0.42; significant with 95% confidence). The coherence spectra between the two records indicates that the time series are moderately correlated (0.40 < r < 0.50) for a narrow coherence period band centered at about 100 days, but are poorly coherent at other periods (not shown). The variability of the baroclinic component accounts for the largest fraction of the absolute transport variability (85%), confirming that the

BC is mostly driven by changes in the baroclinic density field, but also indicating that the variability of the barotropic component is significant, accounting for 15% of the variance of the absolute transport.

3.2 Comparison with other transport estimates

Historical BC transport studies have used different methods to define the depth of the base of the current and to determine its zonal extent, and most of these studies have not accounted for the barotropic component of the flow. Therefore, comparison among the different estimates is not straightforward. Furthermore, the BC has considerable transport variations along its southward path (e.g., Olson et al., 1988; Goni et al., 2011), with baroclinic values ranging from -1.5 Sv to -4.9 Sv to -13.2 Sv at 22°S, 24°S, and 34.5°S, respectively, all calculated using data collected during the same A09 cruise in 2009 (King and Hamersley 2010; Bryden et al. 2011). As such, the most useful comparison is with other measurements collected close to our mooring locations at 34.5°S. Near 37.5°S-38°S, a time-mean baroclinic transport of -10 Sv relative to and above 800 m was reported by Garzoli and Bianchi (1987) from eight months of data from two IES moorings. This value yields roughly one third larger southward flow than our time-mean baroclinic estimate of -7.4 Sv, but the difference is well within our standard deviation of 7.6 Sv. During the Confluence Program study in 1988-1990 (Garzoli, 1993), a much larger southward baroclinic transport was found, -24 Sv at 35.2°S and 36.5°S. Because these historical observations were collected a decade or more before our new results, it is difficult to know whether the observed differences are due to temporal variability between the two time periods or are the result of observational/methodological differences.

The highly spatially-resolved hydrographic sections taken along 34.5°S during the SAMBA-West cruises (Table 2), yield a mean and standard deviation of the transport of -6.6 ± 13.0 Sv (cyan circles and diamonds in Figure 5), which agrees well with the PIES baroclinic transport time series mean and standard deviation of -7.4 ± 7.6 Sv (red curve in Figure 4, Table 3). Note that these hydrographic CTD T and S profiles measured during the SAMBA-West cruises were not included in our GEM construction, thus allowing an independent comparison between the PIES/CPIES-derived estimates and the baroclinic transport estimates from hydrography. The even higher spatially-resolved XBT NOAA transect AX18 along 35°S (e.g., Garzoli & Baringer, 2007; Garzoli et al., 2013; Dong et al. 2015), which has been maintained since 2002, gives a mean and standard deviation for the baroclinic transport relative to and above 800 m between 2010 and 2015 of -9.5 ± 5.8 Sv (the XBT transect was not occupied in 2009). Most of these XBT transects were occupied on average during periods of relatively enhanced southward flow (magenta circles in Figure 5), which explains the slightly enhanced southward baroclinic transport and highlights again the need for continuous measurements to avoid aliasing.

Historical estimates of the absolute transport, including both baroclinic and barotropic components of the flow, are fewer in number and have included a combination of XBTs, Argo, models, and satellite altimetry. Goni and Wainer (2001) found -14.0 ± 7.0 Sv from a TOPEX/POSEIDON ground track crossing the BC near 35°S, in good agreement with our measurements. At 35°S a combination of *in situ* data from the AX18 XBT line and a model-based barotropic adjustment yielded an absolute transport of -19.4 ± 4.3 Sv (Garzoli & Baringer, 2007; Garzoli et al., 2013). More recently, meridional transports of the BC at 35°S in the upper 800 m were estimated as -12.6 ± 2.6 Sv at 34.5°S using Argo and altimetry data (Schmid and Majumder, 2018). While the temporal mean values from the XBT/model adjusted and Argo/altimetry products (-19.4 and -12.6 Sv, respectively) are relatively close to the value we find with the SAMBA-West data (-14.0 Sv), their transport standard deviations (4.3 and 2.6 Sv, respectively) are considerably smaller than what is found with the daily (8.8 Sv) or even with the 30-day low-pass filtered (8.2 Sv) SAMBA-West data. This is almost certainly due to the lower sampling rates of the XBT/model adjusted (quasi-quarterly transects) and Argo/altimetry (7-10 day repeats for altimetry & Argo) products. The lower sampling rates will miss the high frequency variability and rapid changes in the transport that are captured by the SAMBA-West data (section 3.1; Figures 5, 6). An additional disadvantage of the Argo-based product is that Argo data are not available on the continental shelf and upper slope, as floats are generally not deployed inshore of the 2000 dbar isobath to reduce the chances of float grounding (e.g., Riser et al., 2016). Thus Argo-based estimates will miss the portion of the BC inshore of the 2000 dbar isobath which as we have shown is a significant portion of the current (e.g., Figures 3b, 4e).

Our absolute time-mean transport estimate from SAMBA-West is not particularly sensitive to the choice of the reference level to compute the time-mean meridional reference absolute velocities from the high-resolution OFES run (section 2.1.1). Choosing different reference levels instead of 1500 dbar yields differences of -0.6 Sv (stronger southward flow) and

1.7 Sv (weaker southward flow) in the resulting time-mean absolute transport when choosing 1200 dbar or 2100 dbar, respectively. Similarly, if the time-mean reference velocity at 1500 dbar was 10% larger or 10% smaller the resulting transport estimates would differ by only -0.7 Sv (stronger southward flow) or 0.6 Sv (weaker southward flow). Recall the OFES model is only used to provide the time-mean reference velocity value of the non-sheared (barotropic) term, the variability of the barotropic component is purely observational as is the time-mean and temporal variability of the sheared (baroclinic) term (section 2.1.1).

3.3 Spectral distribution of energy of the observed Brazil Current transport

Band-pass filtering the time series into bands with periods shorter than 30 days, periods between 30 and 150 days, and periods longer than 150 days shows that nearly 70% of the absolute transport variance is associated with periods between 30 and 150 days, while periods longer than 150 days account for only 19% of the variance, demonstrating a large variability on monthly and shorter time scales (Table 4). Similarly, 70% of the baroclinic transport variance is associated with periods between 30 and 150 days, while for the barotropic transport a much larger fraction of the variance (40%) is associated with periods shorter than 30 days compared with what is found for the baroclinic term (11%) (Table 4).

Table 4. Variance (Sv^2) and % of variance explained for different period bands in the observed absolute, baroclinic, and barotropic Brazil Current transports.

	Variance explained (Sv^2) (Percentage)		
	3 to 30 days	30 to 150 days	More than 150 days
Absolute	9.2 (12%)	53 (69%)	15 (19%)
Baroclinic	6 (11%)	39 (70%)	11 (19%)
Barotropic	4 (40%)	5 (50%)	1 (10%)

Note. The statistics correspond to the period of continuous measurements between July 2011 and November 2015 (same period used to compute the spectra in Figure 6).

In order to evaluate more precisely how the energy in these records is distributed by time scale, the spectral frequency distributions of the transport time series were computed using the longest portion of the records with continuous measurements (2011-2015) via Welch's periodogram method using a 300-day-wide Hamming window allowing 150 days of overlap (Welch, 1967; Emery and Thomson, 2001). The spectrum of the absolute transport (Figure 6a) has a significant broad peak centered at 100 days and a much less prominent and noisier peak near 20 days. The spectrum of the baroclinic transport also has significant energy near 100 days and a secondary and noisier peak near 20 days (Figure 6b). The barotropic transport spectrum is noisier (Figure 6c) with several shorter period peaks (less than 150 days), with the most prominent peaks at 40-50 and 100 days. The large high-frequency content in all of these records highlights again the importance of continuous measurements with high temporal resolution to avoid aliasing.

To further investigate the origin of the observed transport fluctuations, the spectra of the geopotential anomaly ϕ and bottom pressure at Sites A and B calculated in the same manner are also shown (Figure 6d, e). Both ϕ records at Sites A and B have energy near 20 days, while ϕ at Site B exhibits a very distinctive broad peak centered near 100 days, thus indicating that density variations at Site B largely drive the variability of the absolute transport at this time scale. The 72-h low-pass-filtered records of the absolute transport and ϕ at Site B are significantly anticorrelated with 99% confidence (r = -0.86), implying that a linear relationship between the transport and ϕ at B explains nearly 74% of the absolute transport variance, while the 72-h low-pass-filtered records of the absolute transport and ϕ at Site A are significantly correlated with 99% confidence but the correlation decreases (r = 0.60). Thus, observations of the density field at Site B at 49.5°W are key to the calculation of the BC transport at this latitude, but both sites are relevant in setting the time scales of variability. The spectrum of ϕ at Site AA (at the 2885 m isobath midway between A and B; purple curve in Figure 6d) calculated for the period 2013-2015 also exhibits a peak near 100 days, albeit weaker than at Site B. The correlation between the ϕ records at Sites AA and B is low but significant with 95% confidence (r = 0.31), while the correlation between ϕ records at Sites A and AA is much lower (r = 0.11) and not significant. The absolute transport and ϕ at Site AA are also not significantly correlated (r = -0.21). The ϕ variability at Sites A and B is likely to be influenced by processes such as the variability of separation of the BC from the coast (e.g., Olson et al., 1988; Goni et al., 2011), meanders in the current itself, and eddies propagating into the mooring array and interacting with or superimposing on the current (e.g., Garzoli 1993; Meinen et al., 2017). The spectra of the bottom pressure at Sites A, AA and B have similar characteristics with an apparent common signal across the sites with noisy peaks for periods shorter than 60 days, peaking most prominently at 40-50 days, especially at Site B (Figure 6e). Strong bottom pressure variations at Site A at periods less than 200 days were also identified in Meinen et al. (2018).

In the literature, energetic short-term fluctuations (at periods of 20 to 50 days) have also been identified from observations of the baroclinic component of the BC transport near 37-39°S (Garzoli & Bianchi, 1987; Garzoli & Simionato, 1990). Near the continental slope, the most prominent signal observed was at 21 days, which they postulated was caused by a coastally trapped wind forced response. Garzoli and Simionato (1990) attributed the energy found in the band from 20 to 50 days to wave signals propagating at different periods with opposite directions, specifically a westward propagating wave associated with frontal displacements and an eastward propagating topographic Rossby Wave. Meinen et al. (2017) also noted large westward propagating Rossby Wave-like features within the SAMBA-West array.

The physical mechanisms responsible for the observed BC baroclinic and barotropic transport variability will be examined in a following study.

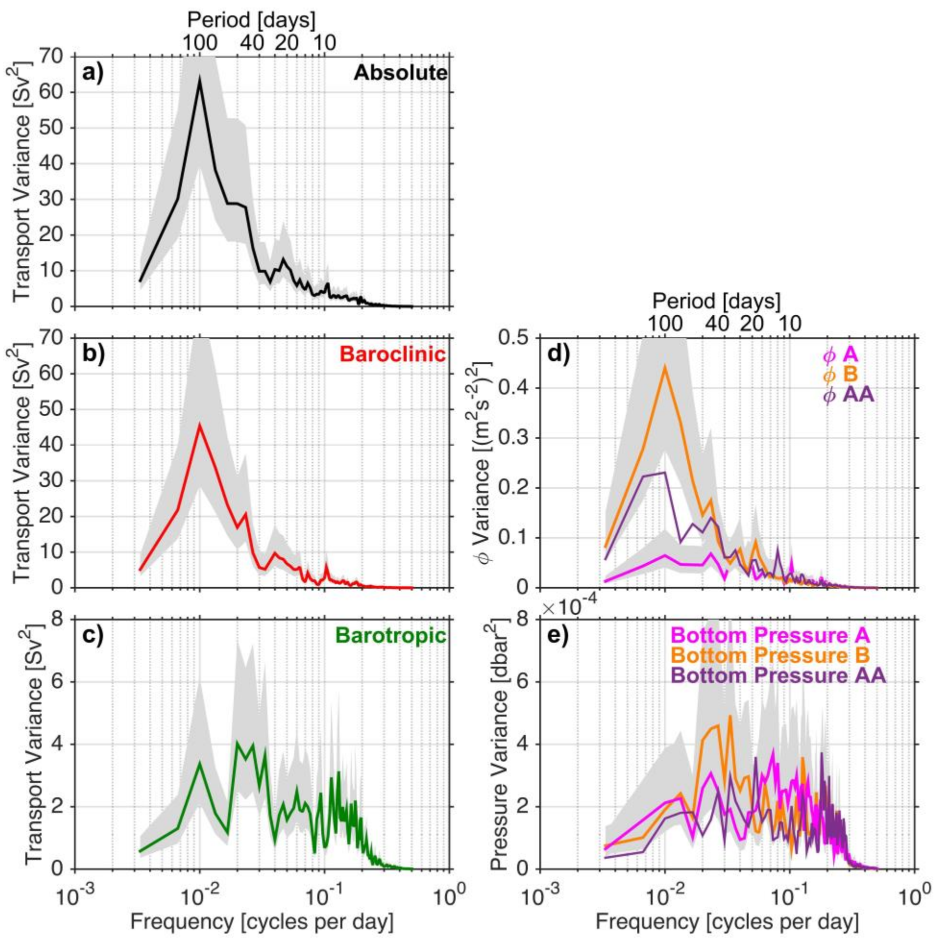

Figure 6. Variance preserving spectra of daily 72-h low-pass filtered a) absolute, b) baroclinic, and c) barotropic Brazil Current transport time series. Note the different scale on the y-axes for c). The variance preserving spectra of ϕ relative to the bottom integrated in the upper layer between the surface and the interface between TW/SACW and AAIW (on average at about 628 dbar) at Site A (magenta) and Site B (orange) are shown in d) and the variance preserving spectra of bottom pressure at site A (magenta) and site B (orange) are shown in e). Shaded confidence interval (95%) is also shown. The spectra of dynamic height (ϕ) and bottom pressure at Site AA (at 50.5°W midway between Sites A and B) for the period 2013-2015 are also shown in panels d) and c) respectively (purple lines; shaded confidence interval is not shown for clarity) to allow comparison. Recall that the time series at Sites A and B constitute the endpoints for the transport calculations. The spectra are computed based on Welch's periodogram method using 300-days-wide Hamming window and 150-days of overlap between consecutive data segments.

The meridional velocity measurements from the ADCP at the 411-m isobath on the upper slope (Figure1, Table 1) provide useful insights about the flow variability and its spectral distribution up on the shelf inshore of Site A. If we assume the moored ADCP velocities are representative of the mean flow inshore of Site A (i.e., from the coast to the 1360 m isobath) and integrate them over this domain we obtain a mean and standard deviation of –0.4 ± 2.0 Sv for the 72-h low-pass filtered record (blue line in Figure 7a). The calculated transport from the ADCP velocities is very small compared to the absolute geostrophic transport estimates in the upper 628 dbar from the PIES/CPIES between Sites A and AA and between Sites AA and B (black and gray lines, respectively, in Figure 7a). The mean moored ADCP transport value also yields much smaller southward flow than the -2.7 Sv determined as the average of the CTD snapshot sections (section 2.2.1), furthermore randomly subsampling the ADCP velocity profiles in a way that mimics the 4 CTD section snapshot realizations for 1000 iterations indicates that only a very low fraction of the random subsamples (about 7%) would produce a mean that is comparable to the CTD time mean of -2.7 Sv. However, the mean moored ADCP transport compares well with the time-mean ~-0.4 to 0.3 Sv seasonally reversing flow at the shallow wide shelf from models (Palma et al., 2008). Additionally, while it is difficult to know for certain how representative the moored ADCP velocity is for the region inshore of Site A, we can definitely note that the largest portion of the variance (70%) in the ADCP record is associated with periods shorter than 40 days (Figure 7b; the spectra is calculated via Welch's periodogram method using a 200-day-wide Hamming window allowing 100 days of overlap), which seems likely to be unrelated to the BC transport variability, as the latter has a very prominent spectral peak near 100 days (Figure 6a).

The ADCP record is short and perhaps a longer record would be able to reveal more pronounced variability on time scales longer than 40 days. However, we note that spectra of the BC transport records between Sites A-AA and AA-B for the identical time period as the ADCP record, 2014-2015 (Figure 7c), are able to show a clear peak at 100 days suggesting that different dynamics are governing the flow at the shelf/upper continental slope and in the BC region.

Given the roughly six-year length of our *in situ* transport records, estimates of the seasonal cycle and/or interannual variations can be made, with the understanding that the number of degrees of freedom is quite small. The annual climatology (average of each transport as a function of year day**)** of the observed 30-day low-pass filtered absolute transport from SAMBA-

West does not indicate a clear annual cycle for the period 2009-2015 for the absolute, baroclinic or barotropic components (Figure 8). The absolute and baroclinic transports exhibit very similar characteristics and amplitudes (Figure 8a, b), while the amplitude for the barotropic transport is much smaller (Figure 8c).

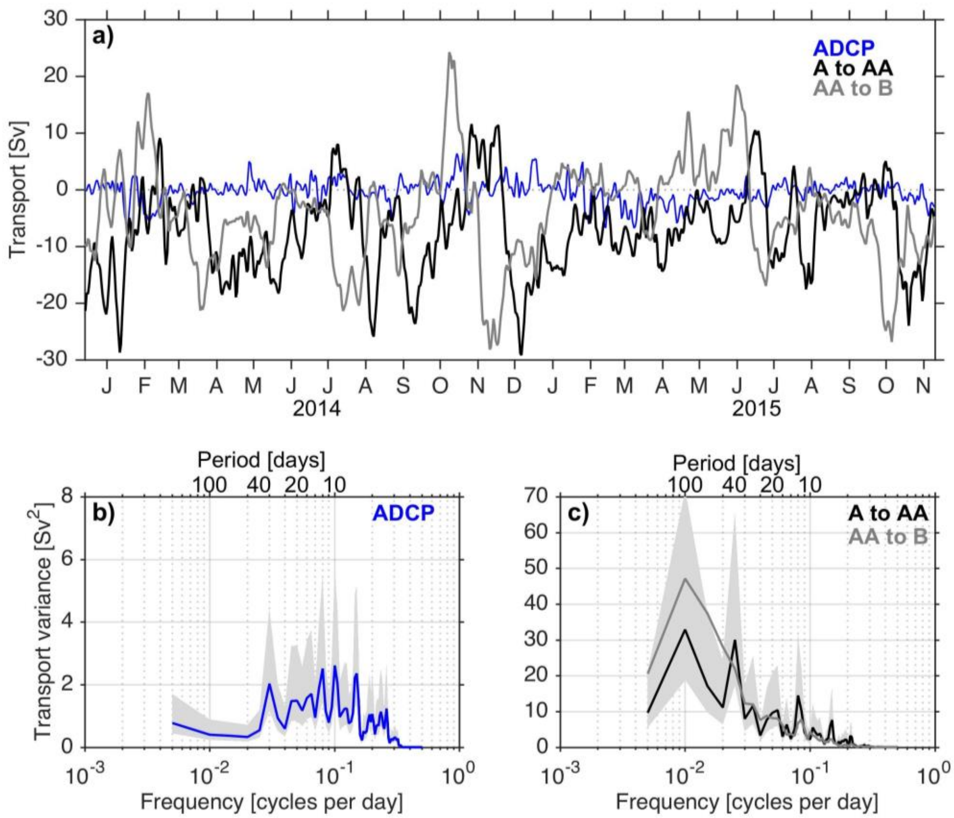

Figure 7. Transport variability on the upper continental slope. a) Estimated 72-h low-pass filtered transport from the ADCP deployed at the 411-m isobath integrated over the area inshore of Site A (blue). Also shown are the 72-h low-pass filtered PIES/CPIES-derived absolute geostrophic transports between Sites A and AA (black) and between Sites AA and B (gray) during 2014-2015 when the ADCP and CPIES at Site AA measurements overlapped. b) Variance preserving spectrum of daily 72-h low-pass filtered ADCP-derived transport. Shaded confidence interval (95%) is also shown. c) Variance preserving spectrum of 72-h low-pass filtered PIES/CPIES-derived absolute geostrophic transports between Sites A and AA (black) and between Sites AA and B (gray; shaded confidence interval is not shown for clarity) during 2014-

2015. Note the different scale on the y-axes for b) and c). The spectra are computed based on Welch's periodogram method using 200-days-wide Hamming window and 100-days of overlap between consecutive data segments.

Our measurements show the strongest southward flow anomalies towards the end of the year in November-December, and the weakest southward flow anomalies during the first half of the year. Given the large high-frequency variability of the geostrophic component of the BC transport, its seasonality does not emerge from 6 years of continuous-in-time data – note the strong northward flow anomaly in October 2014 (strong anomaly in gray line in Figure 8a, b, c; see also Figure 5 and Table 4) when the seasonal cycle is shifting toward stronger southward transport anomalies (black and red curves in Figure 8a, b). The main difference among the three records occurs during January and July. During January, the three records have small amplitudes however both the absolute and baroclinic transport exhibit southward flow anomalies while the barotropic transport exhibits northward flow anomalies. For the period centered at the beginning of July the barotropic transport exhibits its largest southward flow anomalies, while the baroclinic transport have near zero or northward flow anomalies. If we estimate the seasonal cycle instead as monthly means, we similarly find no clear seasonal signal, with monthly values in the absolute transport anomalies that are not statistically different from zero or from each other at even the 67% confidence level (not shown). The seasonal cycle of the BC has previously been analyzed using XBTs (Dong et al. 2014; Goes et al. 2019), as well as with synthetic products derived from Argo data, satellite altimetry, winds, and with numerical models (e.g., Matano, 1993; Combes & Matano, 2014; Schmid & Majumder, 2018). These studies consistently found a seasonal cycle with the BC intensifying during austral summer (January, February, March) and weakening during austral winter (June, July, August) at 35°S, consistent with the meridional displacements of the BMC at seasonal time scales (Olson et al. 1988; Goni & Wainer, 2001; Garzoli & Bianchi 1987; Garzoli & Garrafo, 1989; Saraceno et al., 2004; Lumpkin & Garzoli, 2011) associated with the seasonal variability of the basin-wide wind stress curl (e.g., Matano et al., 1993) or with the first mode of variability of the wind stress curl in a smaller domain in the western South Atlantic (Combes & Matano, 2014). However, we find that the directly locally-forced Ekman transport in the BC region at 34.5°S is very weak, and it has a very small temporal mean and standard deviation of -0.002 ± 0.225 Sv during 2009-2015 (not shown). As in previous studies the Ekman transport anomaly has a well-defined seasonal cycle at this latitude, which reaches its maximum (positive) transport in austral winter and reaches its minimum (negative) transport in austral summer (e.g., Dong et al., 2014). Thus, the Ekman component of the BC transport (computed for the BC region) has a clear seasonality, but it is very weak (peak-to-peak amplitude of only ~0.16 Sv) compared to the much larger monthly peak-to-peak BC geostrophic transport amplitude of ~9 Sv. The standard deviation of the Ekman transport (of $\pm$ 0.225 Sv) is well below the scale of the standard deviation of the observed absolute or baroclinic transport signals (of 8.8 and 7.6 Sv, respectively) indicating that the observed geostrophic transport fluctuations cannot be caused by locally-forced changes induced by wind stress variability at any time scale. As a result, the annual climatology of the geostrophic absolute or baroclinic transports with the addition of the Ekman transport component yield similar results as in Figure 8a and 8b, respectively (not shown). Furthermore, there is a good correspondence between the calculated annual cycle of the BC and the DWBC below (Meinen et al., 2017), which is perhaps not surprising given the low shear between the two southward flows (e.g., Figure 4e). However, neither the BC or the DWBC show a meaningful annual cycle signal, possibly as a result of the aliasing of much stronger amplitude high frequency "noise" when attempting to extract the seasonal cycle. These results, combined with the low percent variance of absolute BC transport signals with periods longer than 150 days in Table 4, reveal that it is likely that more years of continuous measurements are needed to determine whether a robust seasonal cycle exists.

The annual means of the transport records for each individual year are given in Table 3. Note that years 2010 and 2011 have only three months and six months of data, respectively, and hence are excluded from this analysis. Years 2009 and 2015 have eight and ten months of data, respectively, and were judged usable for this analyses. During years 2009, 2012, and 2013, the annual mean absolute transport is markedly steady with only small variations compared to the record-length time-mean value of -14.0 Sv (-14.0, -14.7, and -14.8 Sv, respectively). During 2014 the strongest southward flow is observed (-16.2 Sv) and during 2015 the weakest southward flow is observed (-10.8 Sv), differing from the time-mean by -2.2 Sv (stronger southward flow, due to enhanced southward baroclinic transport, Table 3) and 3.2 Sv (weaker southward flow, due to weaker baroclinic and barotropic transport, Table 3), respectively. However, the 95% confidence limits of the annual means calculated as 2*SEM for each year-long segment are large (SEM is ~3.0 Sv in 2014 and 2015). Additionally, fitting a linear temporal trend to the absolute BC transport for the longest continuous segment (2011-2015) yields 0.08 ± 1.78 Sv/year (-0.05 ± 1.52 Sv/year and 0.13 ± 0.57 Sv/year for the baroclinic and barotropic transports, respectively), not statistically significant with 95% confidence for 64 DOFs following the methods of Bendat and Piersol (1986). Thus, there is no statistically significant trend or deviations from the long-term mean in the yearly averages between 2009 and 2015.

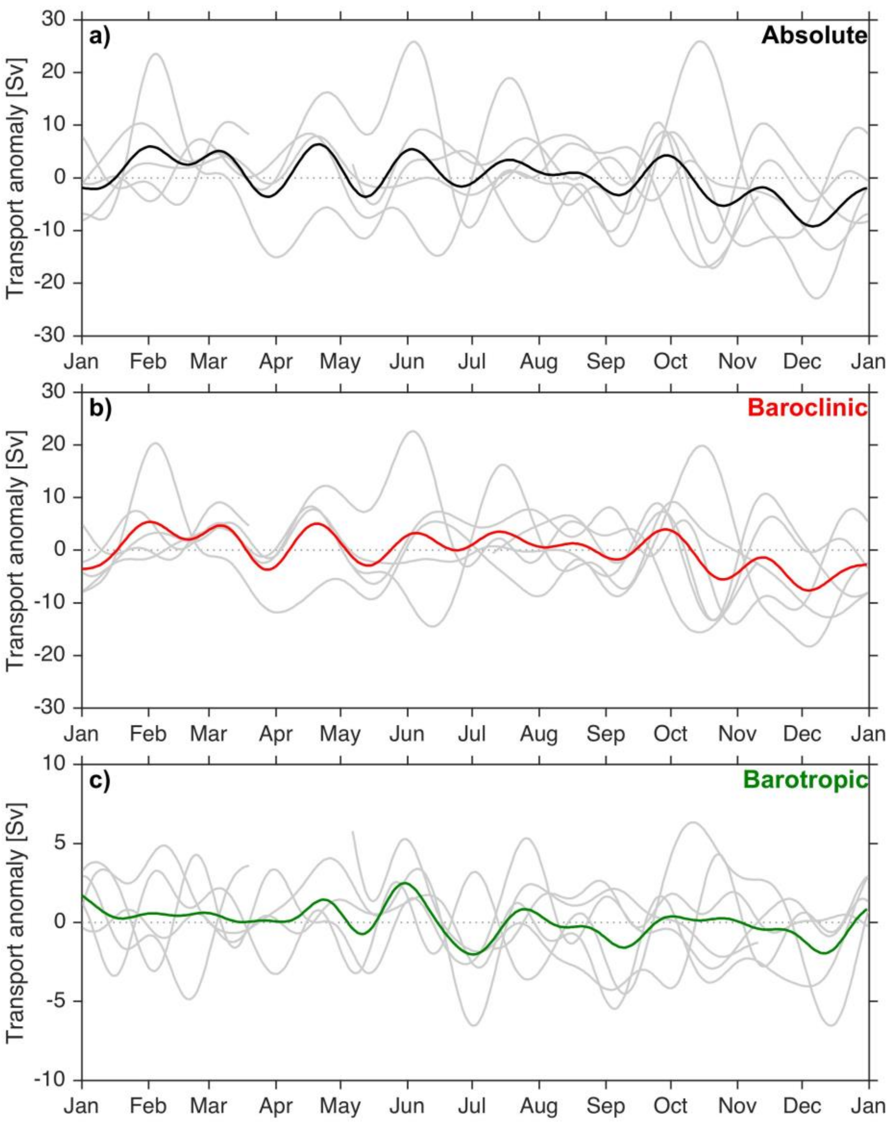

Figure 8. Annual climatology (average for each year day) of 30-day low-pass filtered a) absolute, b) baroclinic, and c) barotropic Brazil Current transport anomalies. Transport anomalies are computed relative to the record-length mean. Light gray lines represent transport

anomalies for each individual year; the thick black, red and green lines are the temporal averages of all individual years in a), b, and c) respectively. Note the different scale on the y-axes for c). Negative values indicate southward flow anomaly.

To investigate the accuracy to which the monthly means and annual means can be estimated from the available 6+ years long BC transport record, we estimate the errors in monthly and annual means performing a Monte Carlo-style analysis (e.g., Emery and Thomson, 2001; Meinen et al. 2010). For the monthly averaging, we randomly selected a month from the transport time series and then we randomly subsampled estimates of daily observations within that chosen month. Subsamples ranging from one to twenty days were averaged and the root-mean-square (RMS) error was computed between those averages and the average from the complete month. A similar procedure is applied to obtain the RMS error of the annual mean estimates. The results for 1000 iterations (Figure 9) indicate that, to consistently obtain a monthly value accurate to within one SEM (of 1.1 Sv, Appendix A) at least 25 daily observations within that month are needed. In order to obtain an annual mean value accurate to 1.1 Sv (or better), at least 50 randomly sub-sampled daily observations are required during that year. These results again point to the need of continuous-in-time daily measurements to resolve monthly to interannual time scales in the BC transport.

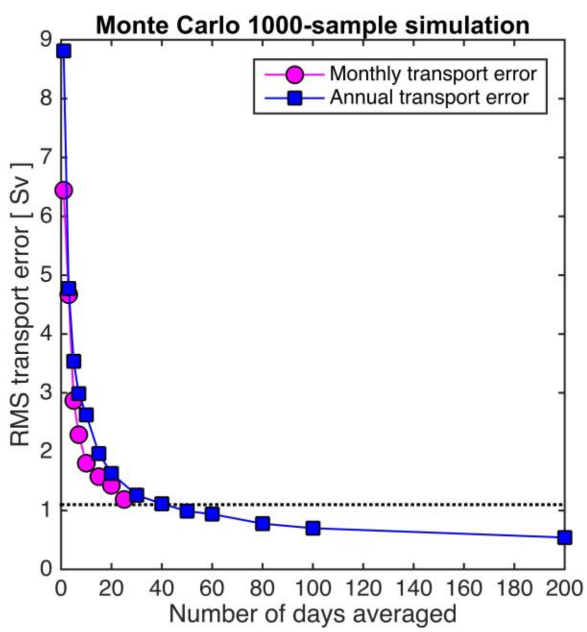

Figure 9. Root-mean-square (RMS) error in the estimates of monthly (magenta) and annual (blue) mean absolute Brazil Current transport versus number of samples considered, after a

Monte Carlo-style analysis with 1000 iterations. The horizontal dotted line indicates the Standard Error of the Mean (SEM) for the absolute transport record (1.1 Sv; Appendix A).

4 Summary and Conclusions

The absolute Brazil Current (BC) transport at 34.5°S, integrated between the sea-surface and the mean-pressure of the density interface between TW/SACW and AAIW, during 2009-2015 has a mean strength of -14.0 Sv, a large standard deviation of 8.8 Sv, and a large peak-to-peak range of 61.7 Sv. Transport variations of 20-30 Sv occur in periods as short as 2-3 weeks and larger variations of about 40-60 Sv occur in periods of 30-60 days, illustrating the dynamic short-term variability in the record and the necessity of continuous-in-time daily observations to avoid aliasing. The daily transport values for the BC are estimated to be accurate to within 2.8 Sv (Appendix A). The long-term PIES/CPIES array at 34.5°S has been shown to capture the majority of the variability of the BC, centered mainly in the area between 51.5°W and 49.5°W. High-resolution hydrographic section data have been used to estimate the mean related transports near the shelf/shelf-break to the west of 51.5°W (which are not currently sampled by the PIES/CPIES array) and data form a bottom moored ADCP provided important insights about upper slope and shelf velocities. Our results indicate that additional time series observations at the shelf-break between the 500 and 1000 dbar isobaths would improve the overall estimation of the BC transport by the array. A new CPIES was deployed in mid 2019 west of PIES A at 875 m (Site 0A) and funding has been awarded to deploy a tall dynamic height mooring additionally equipped with current meters and biogeochemical sensors at the western boundary wedge currently not sampled by the PIES/CPIES array.

The results presented here reveal that the largest part of the variance (~80%) in the absolute BC transport is concentrated at periods shorter than 150 days. The spectral distribution of energy has a well-defined peak near 100 days and secondary (and noisier) peaks between 20-50 days, consistent with earlier studies of the DWBC at this same location (e.g., Meinen et al., 2017). The baroclinic component of the BC transport accounts for the largest fraction of the absolute transport variance (85%), but the barotropic variance contribution (15%) is not negligible. The baroclinic and barotropic transports are uncorrelated on a daily basis, and are only marginally correlated after smoothing with a 30-day low-pass filter. This highlights the need to measure both transport components independently in order to accurately describe the flow variability. No statistically significant seasonal cycle is found and there is no statistically significant trend or variability in the annual averages during the 6+ years of measurements. Our analyses demonstrate that to consistently obtain a monthly value accurate to within 1.1 Sv (one standard error level) at least 25 daily observations within that month are needed, and at least 50 randomly sub-sampled observations throughout a year are required to obtain an annual mean value accurate to 1.1 Sv (or better). It is likely that more years of measurements are necessary to clearly detect significant longer period (seasonal to interannual) signals and trends given the large amplitude of the variability on short time scales.

Overall, the results presented here strongly highlight the strength of a continuous observing system to capture the variability of the BC transport at subseasonal, seasonal, and interannual time scales. A future study will address the physical mechanisms that drive the meridional volume transport fluctuations observed here.

Appendix A. Error analysis

Transport Accuracy Estimates

The transport accuracy estimates closely follow the methods in Meinen et al. (2013). The sources of the transport errors are classified as random or potential bias respective to the time variability. A detailed explanation of the sources of each error is given in Meinen et al. (2013), here we focus on the specifics of the BC estimate.

There are several sources of uncertainty that affect the τ accuracy estimated from PIES/CPIES. To convert the τ accuracies into transport accuracies, we use the linear relationship between the baroclinic streamfunction (a.k.a. Fofonoff potential or potential energy anomaly, χ) and τ at 1000 dbar (τ_{1000}). Here the vertical integration domain for χ is chosen to be consistent with the layer occupied by the BC (ie. between the surface and approximately 628 dbar, $\chi^{0\text{-}628}$). There is a tight relationship between τ and $\chi^{0\text{-}628}$. The slope for a linear fit is -0.6 10^5 J m^{-2} ms^{-1}. The χ accuracy is converted into a transport error bars by using the local Coriolis parameter f = -8.2605 10^{-5} s^{-1} and a constant density of 1030 kg m^{-3}.

The random sources of the geostrophic velocity or transport (baroclinic relative to an assumed level of no motion and barotropic or reference velocity) affect the time-variability of the transport and can be listed as follows: 1) the accuracy of PIES/CPIES-measured τ (0.17 ms;

following Chidichimo et al., 2014, revised after Donohue et al., 2010); 2) the scatter in the calibration relationship to convert PIES/CPIES τ actual depth on the fixed 1000 dbar level (0.2 ms or 0.5 Sv; as estimated in Meinen et al., 2013); 3) the accuracy of the GEM lookup tables calculated as the rms scatter about the fit between $\chi^{0\text{-}628}$ and τ_{1000} (1.65 10^5 J m^{-2}); 4) additionally, for the calculation of the reference velocity accuracy the sole random source of error are the pressure gauges (0.01 dbar; Donohue et al., 2010). These four random errors amount to 1.9, 0.5, 1.8, and 0.7 Sv, respectively (independent of one another) and are combined via the square-root of the sum of the squares method yielding a total random error of 2.8 Sv for the 72-h low-pass filtered BC transport estimates.

There are two potential sources bias errors affecting the time-mean transport: 1) the offset due to the calibration of τ into the equivalent τ_{index} from CTDs (1.8 ms or 1.9 Sv from Meinen et al., 2013); 2) the accuracy of the time-mean reference velocity between Sites A and B, estimated conservatively as the difference of time-mean reference velocity at 1500 dbar between two sources: the 35-year run of OFES (1980-2015) described herein (-0.058 m s^{-1}) and a 26-year run (1992-2018) of the Estimating the Circulation and Climate of the Ocean, Phase 2 (ECCO2) ocean state estimate (https://ecco.jpl.nasa.gov/) (-0.027 m s^{-1}). The approximate difference between these estimates amount to 0.031 m s^{-1}, which considering a width of 1.8315 x 10^5 m and a vertical layer of about 628 m translates into a potential transport bias error of 3.5 Sv. Adding more models would refine/reduce the estimate of this source of bias in the future, but this is beyond the scope of this study. As these potential bias sources of errors are independent of one another, they are also combined via the square-root of the sum of the squares, yielding 3.9 Sv.

Degrees of freedom, integral time scales, standard error of the mean

The integral time scale (ITS) for the absolute transport record is determined by the first zero crossing of the autocorrelation function of the daily transport time series (Emery and Thomson, 2001). Evaluating the longest segment with continuous measurements (between July 2011 and November 2015, almost 4.5 years of data) of the absolute transport record, results in an ITS of 12 days, thus every 24 days there is an independent "degree of freedom" (DOF) for estimating both the average and the statistical accuracy of the average. This yields about 64 DOFs and a Standard Error of the Mean (SEM) of 1.1 Sv. If we also include the first one-year segment in the analysis, the DOFs increase to 76 and the resulting SEM is 1.0 Sv. To be conservative we keep the estimate for the longest period with continuous measurements. The ITS after applying a 30-day low-pass filter is slightly longer (14 days), thus the time series have fewer DOFs (56). The statistical significance of the correlation coefficients is calculated following the methodology described in Emery and Thomson (2001). For 60 DOFs, correlations (r) for 72-h low pass filtered records with cutoff $r = |0.250|$ and $r = |0.325|$ are significantly different from zero at the 95% and the 99% confidence level, respectively. For 50 DOFs, correlations with cutoff $r = |0.273|$ and $r = |0.354|$ are significantly different from zero at the 95% and the 99% confidence level, respectively.

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Data: The PIES/CPIES and hydrographic data used herein from the South Atlantic MOC Basin-wide Array (SAMBA) can be found at:
https://www.aoml.noaa.gov/phod/SAMOC_international/samoc_data.php
The STSF hydrographic and LADCP data are available at https://doi.org/10.7910/DVN/IHWLL3 and https://doi.org/10.7910/DVN/JJNFKG, respectively.

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