Ágata Piffer Braga

Description and Dynamics of the Santos Bifurcation

A thesis presented to the Instituto Oceanográfico of Universidade de São Paulo, in partial fulfillment of the requirements for obtaining the degree of Master in Science, Oceanography Program concentration area of Physical Oceanography

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2019

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I would like to dedicate this work to my parents who always gave me unconditional love and unbreakable support. They raised me teaching that the process of learning is a life time objective, and that we should never stop perusing it. I hope this work can give a clue on how I value this lesson.

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"Our lives are full of bifurcations..."

Me, at some point in the last month of my masters.

Contents

Contents	i
List of Figures	iii
List of Tables	vi
List of Abbreviations	vii
Abstract	viii
Resumo	1
1 Introduction	1
1.1 Preamble	1
1.2 Ecological and Economic Relevance	2
1.3 Water Masses	3
1.4 Current Patterns	6
1.5 Vertical Shear and Vertical Modes	10
1.6 Hypotheses	15
1.7 Objectives	15
2 Study Area and Datasets	16
2.1 Study Area	16
2.2 Cruise OSE 1	17
2.3 Mercator Products	18
2.4 Mercator Output Validation	20
2.4.1 Geostrophic Stream Function	20
2.4.2 Velocity derived Stream Function	24
3 The Mean Circulation Pattern	28
3.1 Horizontal Perspective	28
3.2 Vertical Perspective	32
3.3 Volume Transport Intercomparison	37
3.4 Annual and Intraseasonal Variability	38

4	Dynamical Mode Analysis	45
4.1	Brunt-Väisälä and Deformation Radii	46
4.2	Mean Circulation Modal Structure	52
5	Summary and Conclusions	64
	Bibliography	66

List of Figures

1.1	Stramma and England (1999) scheme	3
1.2	Prevailing stratification scheme	4
1.3	(a) Water masses and currents scheme. (b) Current system scheme adapted from Soutelino et al. (2013)	5
1.4	(a) Boebel et al. (1999b) floats trajectories, (b) Boebel et al. (1999b) mean velocities	7
1.5	Stommel (1965) scheme	8
1.6	Rodrigues et al. (2007) bifurcations	9
1.7	Upper panel: Legeais et al. (2013) trajectories of the floats. Lower panel: Legeais et al. (2013) averaged float velocities	11
1.8	(a) Silveira et al. (2004) vertical velocities, (b) Silveira et al. (2004) velocities components, baroclinic and barotropic, (c) Silveira et al. (2004)	12
1.9	Scheme of the south Atlantic circulation at intermediate levels	13
2.1	Bathymetric map from the Brazilian southwestern coast.	17
2.2	OSE 1 stations map. The cruise was constructed between 21 March 2000 to 06 July 2000, therefore comprehending the austral late fall and early winter.	18
2.3	A and C: OSE1 objective mapped geostrophic ψ for 200 m and 900 m. B and D: Mercator objective mapped geostrophic ψ for 200 m and 900 m. . .	22
2.4	OSE1 objective mapped geostrophic ψ for 200 m and 900 m. B and D: Mercator objective mapped velocity derived ψ for 200 m and 900 m. . . .	25
2.5	A and C: Mercator objective mapped geostrophic ψ for 200 m and 900 m. B and D: Mercator objective mapped velocity derived ψ for 200 m and 900 m. .	26
3.1	Mercator objectively mapped ψ (upper panel) and velocity field (lower panel) at 900 m. The green dot is the SB location.	29
3.2	Mercator objectively mapped ψ for 200, 400, 600, 900, 1500 and 3000 m. .	30
3.3	Mercator Contour map of ψ at 900 m and bathymetric contours. The ψ (bathymetry) contours are displayed in black (grey)	32

3.4	Map of the location of sections used to describe the vertical current patterns. In green, the E section, in magenta the S section, in light blue the N section and in yellow the AI section.	33
3.5	Cross-transect velocity, volume and heat transport for the N section. The isopycnals interfaces between water masses are presented in cyan with their respective σ_θ values. Southward velocities are in blue.	34
3.6	Cross-transect velocity, volume and heat transport for the S section. The isopycnals interfaces between water masses are presented in cyan with their respective σ_θ values. Southward velocities are in blue.	35
3.7	Cross-transect velocity, volume and heat transport for the E section. The isopycnals interfaces between water masses are presented in cyan with their respective σ_θ values. Westward velocities are in blue.	36
3.8	Along-transect velocity for the AI section. The isopycnals interfaces between water masses are presented in cyan with their respective σ_θ values. Southwestward velocities are in blue.	37
3.9	Seasonal plots for the 900 m horizontal field. The green dot represents the reference position of the SB for the mean scenario.	39
3.10	AAIW volume transport.	40
3.11	AAIW heat transport	41
3.12	Volume and heat transport as percentage from the North and South to the East.	41
3.13	AAIA+UCDW Transport.	41
3.14	AAIA+UCDW Heat Transport.	42
3.15	Transport Percentage from the North and South to the East.	42
3.16	Hovmoller for the a zonal line at 29°S.	43
3.17	Hovmoller for the AI Section line extended. The black dashed line is the 29°S latitude, which is the SB main latitude.	44
4.1	Exemplification of the smoothing of the N^2 profile at one open ocean grid point.	47
4.2	Modal decomposition and recomposition for the sSEC at 43°S, 28.75°W: a) N^2 ; b) The first three dynamical modes; c) Modal decomposition of zonal velocity profile for $i = 0, 1, 2$; d) Modal recomposition truncated to $i = 0 - 1$, $i = 0 - 1 - 2$ and $i = 1 - 2$	48
4.3	Map of the zeroth deformation radius for the study area.	49
4.4	Map of the first deformation radius for the study area.	50
4.5	Map of the second deformation radius for the study area.	51
4.6	Barotropic, first and second baroclinic modal decomposition for the ψ field at 900 m. The green point indicates the reference location of the SB at 900 m.	53

4.7	Modal recomposition for ψ field at 900 m by: barotropic and the first baroclinic modes (upper panel), barotropic and first two baroclinic modes (central panel), and the first two baroclinic modes (lower panel).	54
4.8	Modal decomposition for the AI Section. The plots in each panel shows modal Amplitude values along the section. The lower plots correspond to the product of the modal amplitude to the corresponding vertical dynamical mode. The dashed black line point out the zero velocity contour for the total velocity.	56
4.9	Velocity modal recomposition for the AI Section. BT+BC1, upper left, BT+BC1+BC2, upper right, and BC1+BC2 lower panel. The dashed black line point out the zero velocity contour for the total velocity.	57
4.10	Modal decomposition for the East Section. The plots in each panel shows modal Amplitude values along the section. The lower plots correspond to the product of the modal amplitude to the corresponding vertical dynamical mode. The dashed black line point out the zero velocity contour for the total velocity.	58
4.11	Velocity modal recomposition for the East Section. BT+BC1, upper left, BT+BC1+BC2, upper right, and BC1+BC2 lower panel. The dashed black line point out the zero velocity contour for the total velocity.	59
4.12	Modal decomposition for the North Section. The plots in each panel shows modal Amplitude values along the section. The lower plots correspond to the product of the modal amplitude to the corresponding vertical dynamical mode. The dashed black line point out the zero velocity contour for the total velocity.	60
4.13	Velocity modal recomposition for the North Section. BT+BC1, upper left, BT+BC1+BC2, upper right, and BC1+BC2 lower panel. The dashed black line point out the zero velocity contour for the total velocity.	61
4.14	Modal decomposition for the South Section. The plots in each panel shows modal Amplitude values along the section. The lower plots correspond to the product of the modal amplitude to the corresponding vertical dynamical mode. The dashed black line point out the zero velocity contour for the total velocity.	62
4.15	Velocity modal recomposition for the South Section. BT+BC1, upper left, BT+BC1+BC2, upper right, and BC1+BC2 lower panel. The dashed black line point out the zero velocity contour for the total velocity.	63

List of Tables

2.1	Values of the Pearson Correlation Coefficient and the Normalized Root Mean Squared Error, for 200 m and 900 m, between the OSE 1 and Mercator scenarios.	24
2.2	Comparison of the maximum and mean velocities (in m s^{-1}) obtained from the ψ fields for the OSE 1 and Mercator scenarios.	24
2.3	Values of the Pearson Correlation Coefficient and the Normalized Root Mean Squared Error, for 200 m and 900 m, between the OSE 1 and Mercator by the VectOA.	25
2.4	Values of the Pearson Correlation Coefficient and the Normalized Root Mean Squared Error, for 200 m and 900 m, between the Mercator ψ geostrophic and velocity derived.	26
3.1	Transports for each water mass during Autumn (Aut), Winter (Win), Spring (Spr), Summer (Sum) and annual mean (Mean). All values are presented in Sverdrups ($1 \text{ Sv} = 10^6 \text{ m}^3 \text{ s}^{-1}$).	40
4.1	NRMSE and EV for 900 m horizontal field between the total Mercator velocity and the reconstituted velocities sets.	52

List of Abbreviations

BC	Brazil Current
SB	Santos Bifurcation
SEC	South Equatorial Current
IWBC	Intermediate Western Boundary Current
DWBC	Deep Western Boundary Current
DWBC	Deep Western Boundary Current
NBUC	North Brazil Under Current
OA	Objective Analysis
VectOA	Vectorial Objective Analysis
CDM	Classic Dynamic Method
AAIW	Antarctic Intermediate Water
SACW	Atlantic Central Water
NADW	North Atlantic Deep Water
SCW	Superior Circumpolar Water
CTD	<i>Conductivity, Salinity and Depth</i>
LADCP	<i>Lowered Acoustic Doppler Current Profiler</i>
LSM	<i>Level Set Method</i>
VMADCP	<i>Vessel Mounted Acoustic Doppler Current Profiler</i>
XBT	<i>Expendable Bathythermograph</i>
LaDO - IOUSP	Laboratório de Dinâmica Oceânica do Instituto Oceanográfico
OSE 1	Oceano Sudeste 1(Southeast Ocean 1)
BT	Barotropic mode
BC1	First Baroclinic
BC2	Second Baroclinic
N	North Section
S	South Section
E	East Section
AI	Along Isobath Section
SPP	São Paulo Plateau
RGR	Rio Grande Rise

Abstract

In this study we describe the Santos Bifurcation mean pattern and investigate its modal composition. Also we analyze and compare the local water masses transports for the Santos Bifurcation split, in terms of annual and intraseasonal means. The Santos Bifurcation is crucial to the understanding of the Brazil Current system and it is located in a important oil and gas exploration blocks domain, what increases even more the demands of environmental control in the region. In order to answer our questions we validated the Mercator Global model output with quasi-synoptic data from the OSE Brazilian Navy cruise. The results shown virtually constant position of the bifurcation occurrence over time, although the split transport presented variations. During May and June the northern branch volume transport is higher, representing around 68% of the sum. During the October and November months, the split reaches its most discrepancy, when the southern branch is responsible for more than 74% (more than 2/3) of the sum. In the other months, despite the transport of the southern branch is higher, it corresponds to 50-60% of the sum. Regarding the modal composition we have found that the south South Equatorial Current is essentially barotropic, and its first and second baroclinic modes have less than 3.5% of explained variance. On the other hand, in the Brazil Current system the baroclinicity increases equatorward, changing from 2.84, South of the bifurcation, up to 34%, north of the bifurcation, of explained variance. This is the first tridimensional description of the Santos bifurcation, and its transport variability.

Keywords: Santos Bifurcation, Intermediate Water Bifurcation, Brazil Current, Intermediate Western Boundary Current, South Equatorial Current, Antarctic Intermediate Water, Brazil Current System.

Resumo

Neste estudo descrevemos o padrão médio da Bifurcação de Santos e investigamos sua composição modal. Também analisamos e comparamos os transportes das massas de água locais para na partição da Bifurcação de Santos, em termos das médias anuais e intra-sazonais. A Bifurcação de Santos é crucial para o entendimento do sistema da Corrente do Brasil e está localizada em um importante domínio de blocos de exploração de petróleo e gás, o que aumenta ainda mais as demandas de controle ambiental na região. Para responder às nossas perguntas, validamos a saída do modelo global Mercator com dados quase sinóticos do cruzeiro OSE realizado pela Marinha do Brasil. Os resultados mostraram uma posição praticamente constante da ocorrência da bifurcação ao longo do tempo, embora a partição do transporte tenha apresentado variações. Durante maio e junho, o transporte de volume do ramo norte é maior, representando cerca de 68% da soma. Durante os meses de outubro e novembro, a divisão atinge sua maior discrepância, quando o ramo sul é responsável por mais de 74% (mais de 2/3) da soma. Nos outros meses, apesar do transporte do ramo sul ser maior, corresponde a 50-60% da soma. Com relação à composição modal, descobrimos que a parte sul da Corrente Sul Equatorial é essencialmente barotrópica, e que o primeiro e segundo modos baroclínicos têm menos de 3,5 % da variância explicada. Por outro lado, no sistema da Corrente do Brasil a baroclinicidade aumenta no sentido do equador, passando de 2,84, ao sul da bifurcação, até 34 %, ao norte da bifurcação, de variância explicada. Esta foi a primeira descrição tridimensional da Bifurcação de Santos e de sua variabilidade temporal de transporte.

Descritores: Bifurcação de Santos, Água Intermidiária Antártica, Corrente do Brasil, Sistema da Corrente do Brasil, Corrente Intermidiária de Contorno Oeste, Corrente Sul Equatorial.

Chapter 1

Introduction

1.1 Preamble

The northern portion of the South Atlantic subtropical gyre, named South Equatorial Current (SEC), carries water from the Benguela Current region crossing the South Atlantic basin towards the Brazilian coast. The SEC reaches the Brazilian continental margin as four different branches (Figure 1.1): The Northern SEC (nSEC), the Equatorial SEC (eSEC), the Central SEC (cSEC) and the Southern SEC (sSEC). The sSEC, in particular, approaches the coast as a broad flow and hits the continental slope at different latitudes at different depths (Stramma and England, 1999; Boebel et al., 1999b; Silveira et al., 2000; Rodrigues et al., 2007). Stramma and England (1999), Boebel et al. (1999b) and Wienders et al. (2000) have documented this pattern by observational data, showing the sSEC arrival at 14°–16°S in the surface layer (0–150 m), at 20°–24°S in the thermocline level (150–500 m), and at 26°–28°S in the intermediate layer (600–1300 m). As the sSEC reaches the continental margin at different latitudes with depth, so does the northern limb of the South Atlantic Subtropical Gyre. The SEC is intrinsically related to the climate equilibrium once it is the major contributor to the Meridional Overturning Circulation (MOC) (Talley, 2003; Ganachaud, 2003; Lumpkin and Speer, 2003) and also, the leading mass exchange from the subtropical cells (STC) to the tropical portion (Malanotte-Rizzoli et al., 2000; Zhang et al., 2003).

When the sSEC reaches the coast it splits into two portions, bifurcating, one flowing poleward and another flowing equatorward. These splits occur in all the sSEC arrivals. However our interest regards in to investigate the bifurcation from the southern impinging branch of the sSEC at sub-pycnocline, or intermediate level, renowned as Santos Bifurcation (SB) (Stramma and England, 1999; Boebel et al., 1999b). The SB is an important limit for the Brazil Current (BC) system, where its presence marks the change from a highly baroclinic to a stronger barotropic regime. The southern portion of the bifurcation, that turns pole-wards, incorporates the portion of water brought by the intermediate

sSEC jet, thickening the BC from 400 m to 1300 m (Rocha et al., 2014). Moreover, the northern portion, which flows equator-wards, generates the Intermediate Western Boundary Current (IWBC). This northern is confined between the BC and the Deep Western Boundary Current (DWBC), creating a multi-layer system with inverse flowing currents, such as predicted in the seminal work of Stommel (1965).

The SB plays a key role on the understanding of the BC system and the South Atlantic ocean circulation. Despite the SB has been studied by several authors (Stramma and England, 1999; Boebel et al., 1999b; Silveira et al., 2000; Schmid et al., 2000; Rodrigues et al., 2007; Schmid and Garzoli, 2009; Legeais et al., 2013; Schmid and Majumder, 2018), it is still incipient knowledge and it is required further investigation in terms of its dynamics, three-dimensional structure and temporal variability. In the following paragraph we intent to bring to light the state of the art of the SB studies and consequently formulate our scientific questions and hypotheses. .

1.2 Ecological and Economic Relevance

Its domain lies at the São Paulo State's continental margin, which is an economic pole in Brazil. The state resides the busiest container port in Latin America and Brazil's leading port in container traffic, the Port of Santos¹. Furthermore the region lays in one of the most profitable oil and gas exploration blocks domain, within the Santos Basin (23°S – 28°S), also referred as Santos Plateau. This scenario highly demands an environmental control and knowledge of the area (Biló et al., 2014a). Our study area also encompasses the Rio Grande Rise, a geomorphological feature of more that 3000 km² and rising more than 3000 m from the abyssal plain. Since 2009 the Brazilian Geologic Service (CPRM) research this geological feature, which revealed to be a great potential mineral source. In 2015, the International Seabed Authority (ISBA) and the CPRM signed an environmental study program to explore the Cobalt-rich Ferromanganesifers crosts (PROERG). This agreement guarantees to Brazil 15 years of exclusive rights to research the area ².

This scenario magnified the interest of knowing better all the aspects of this region, including its water masses circulation and patterns. The BC system is extremely important to the understanding of the ocean circulation, mass and heat transport towards the pole. The BC variability and mesoscale activity exceeds its analogous northern hemisphere counterpart, the Gulf Stream, in prior to the separation from the coast (Soutelino et al., 2013). The BC eddies are determinant for the larval distribution of benthic organisms (Yoshinaga et al., 2010). Sumida et al. (2004) also relates the presence of nutrient rich Antarctic Intermediate Water (AAIW) to the development of deep corals and associated

¹<http://web.archive.org/web/20070512181621/http://www.gfmag.com/index.php?idPage=315>

²<http://www.cprm.gov.br/publique/Noticias/CPRM-Comemora-a-Primeira-Expedicao-na-Elevacao-do-Rio-Grande-1495.html>

rich fauna. As well as recent geological study Schattner et al. (2016) revealed 984 elliptical pockmarks on the Santos Basin, exactly under the BC–DWBC transition, which owe their formation and geometry to gas exudation and to the gradients of this currents velocities. The depth-dependent bifurcation pattern has an analogue episode in the North Equator Current (NEC) in the Pacific Ocean, as described by Qu and Lukas (2003), therefore the investigation and understanding of the SB can aggregate to the learning of this analogue structure and, by all means, other patterns around the world ocean.

Figure 1.1: Scheme of the SEC branches at intermediate depth from Stramma and England (1999)

1.3 Water Masses

Prior to reviewing the previous works about the SB, it is desirable to characterize the water mass conformation of the region, once they are one of the eminently tracers of the circulation. The classical concept of water mass formation assumes the existence of water sinking (deep convection) sites around the globe. These locations are regions of important sea-atmosphere interactions, where the exchanges between the ocean and air impel these local waters to become more dense. As these waters sink they preserve their thermodynamical properties, once temperature and salinity are considered conservative

properties, and that they only erode slowly during their paths across the ocean basins or by mixing with another to other places.

In the case of surface water masses, their direct exposure to rainfall and solar radiation provides them a wide range of salinity and temperature values. The consequence of this interaction of different types of water is an overlap of increasing density layers, formed by their primary patterns and proportions of their mixtures, outcoming a vertical and horizontal stratified water column. More recently in the oceanographic history, the identification of the water masses has been improved thanks to the instrumental technological progress. New instruments obtain better accuracy and precision, therefore more accurate measurements of the conservative properties (Temperature and Salinity). The water mass analyses incorporates new tracers for the identification of the water masses. Semi-conservative tracers, such as dissolved oxygen, nitrate, phosphate and silicate, can also contribute to the understanding of the transports and circulations of these water masses (Emery, 2001; Zemba, 1991).

The South Atlantic Ocean receives water from various locations as the North Atlantic, the Pacific, the Weddel Sea, the Sub-Antarctic Convergence and the Subtropical Convergence, as well as a small portion of the Indian Ocean brought by the Current of the Agulhas. The scheme of Figure 1.2 simplifies the water mass stratification encountered in the surroundings of the SB. In the upper layer (between 0 m and around 150-200 m), the BC core region, we observe the Tropical Water (TW), a warm and salty in constant exposure to the ocean-atmosphere interface. Right underneath the TW, we can detect the South Atlantic Central Water (SACW), which displays uniform properties throughout its range where the T-S curve appears as a straight line between points of 5° C, 34.3 and 20° C, 36.0, as analogous of the curves of Indian (ICW) and Western South Pacific Central Water (WSPCW) (Miranda, 1982; Stramma and England, 1999; Tomczak and Godfrey, 1994; Zemba, 1991). Actually, the SACW is not only formed by waters that are subducted of in the subtropical convergence, but also receives important contribution of the ICW, which is brought from the Agulhas Current intrusions leaked edge (Stramma and England, 1999). The eddy separation region and the Agulhas Current incorporates the contribution into

Figure 1.2: Prevailing stratification scheme within our study domain, where each color represents a water mass sided by its mainly depth. TW stands for Tropical water, SACW for South Atlantic Central Water, UCDW for Under Current Deep Water and NADW North Atlantic Deep Water.

the T-S, unaffected its curve yet adding volume to the waters around the 13° C. Reaching depths of 550 m, the SACW is also transported by the BC into the pycnocline layer. The isopycnal $27.1 \sigma_0$ marks the transition between the SACW and the Antarctic Intermediate Water (AAIW), that as an intermediate water mass, which receives also the contribution of the Indian Ocean, joining the South Atlantic AAIW via the Agulhas Current leakage. The South-Atlantic-formed AAIW originates at the Sub-antarctic Front (Stramma and England, 1999). The AAIW core can be recognized by a subsurface oxygen maximum and a salinity minimum. The low-salinity is found at a depth of about 300 m near the Subantarctic Front (45° S), descending northward to 900 m at 30° S in the subtropical gyre, center and rising again to 700 m near the equator (Tsuchiya et al., 1994). The AAIW is the main water mass carried by the intermediate sSEC and at the SB, which leads our attention to it as a preferred tracer of our features of interest.

Figure 1.3: (a) Scheme of the water masses carried by each current following the concept of Stommel (1965). (b) Schematic representation of the Santos Basin current system, adapted from Soutelino et al. (2013).

Below the AAIW, we find the Upper Circumpolar Deep Water (UCDW) following the AAIW circulation pattern in South Atlantic. It is transported towards the North Atlantic carried by the IWBC in the northern portion of the SB. The UCDW as a whole is a large body of fresh and oxygen-poor water, that associated with the Antarctic Circumpolar Current (ACC) and is found over a large density range. An oxygen minimum associated with the UCDW is found in the South Atlantic lying below the oxygen maximum of the AAIW and extending northward to about 22° S at 25° W (Tsuchiya et al., 1994). North of 22° S, the intense, shallow equatorial oxygen minimum obliterates the signature of below the UCDW. In the tropics the isopycnal $32.15 \sigma_1$ limits the UCDW from the

North Atlantic Deep Water underneath (NADW). The NADW is carried by the DWBC and presents a much lower nutrient content compared with the UCDW. Its vertical extent is 1300 m, separating the BC in the southern portion of the SB from the DWBC, until 3900 m (or the ocean floor, if in the surroundings of the São Paulo Plateau).

1.4 Current Patterns

In order to summarize the circulation configuration of the BC system as well as the water masses transported, we can inspect at Figures 1.3a and 1.3b. Inside our study domain (22.5°S - 33°S , 50°W - 33°W) we have from the surface up to about 560 m the BC carrying TW and SACW continuously towards the pole (Stramma and England, 1999; Silveira et al., 2004; Biló et al., 2014a). Its velocity magnitude can exceed the 0.5 m s^{-1} and transports between 5-7 Sv (Biló et al., 2014a). Between 560 m and 1200 m, we can observe the nearly zonal flow coming from the east, the sSEC branch (Boebel et al., 1999b). Its AAIW transports is 12 Sv of AAIW (Boebel et al., 1999b). Upon reaching the Santos Basin and colliding with the São Paulo Plateau, it splits into two branches. The northern branch goes equatorward, generating a counter-current constrained between CB and the DWBC. the IWBC 6 Sv of the AAIW and 2-3 Sv of the UCDW are transported (Stramma and England, 1999). According to Boebel et al. (1999b), its average maximum velocity is 0.3 m s^{-1} . This IWBC vertical extent can exceed and 1000 m depth and its core is located at 900 m (Silveira et al., 2004). The southern branch flows poleward being incorporated by the BC, the intermediate BC, and adds about 9 Sv to the BC transport, according to Boebel et al. (1999b). The thickened BC goes from 200 m until 1200 in this region, laying on the top of the DWBC transporting the NADW. Thus, the total transport pole-wards becomes more than 12 Sv, with maximum local speeds up to 0.7 m s^{-1} .

Heading to a brief review on some of the previous works regarding the SB, we begin from the original descriptions of this ocean circulation feature. Stramma and England (1999) based on the World Ocean Circulation Experiment (WOCE) datasets and supported by numerical modeling, the main characteristics and paths of the water masses in the South Atlantic are presented as summary sketches for the domain of all water masses (See Figure 1.3a for the AAIW level). In the same year we also have the description of the SB in a Lagrangian approach by Boebel et al. (1999a). Gathering neutrally-buoyant floats and current meters data (Figure 1.4) the study measurements reveals the BS at 29° S, firstly hypothesized by Reid (1994). They portrayed the intermediate flow arriving with the sSEC and splitting into two branches: 9 Sv goes south and 4 Sv flows north. They also emphasize the formation of the IWBC and its relevant impact in the calculation and modeling of the anomalous northward basin wide heat flux found in the South Atlantic, since it introduces a shallow level of no motion to the BC.

Figure 1.4: (a) Summary of the floats trajectories by Boebel et al. (1999b). (b) Space-time averaged mean velocities based on a total of 170 float years of velocity data, obtained by Boebel et al. (1999b).

The observed IWBC jet presents a width of about 30 km and 400 m vertical thickness with the average velocity being 0.3 ms^{-1} . Near the upper latitudinal limit of 4°S of the IWBC, part of the current veers east, indicating a possible mechanism of injection of AAIW into the equatorial regime. The portion that does not undergo the retroflexion continues to contour the continental margin and culminates in feeding the Northern Brazil Current (NBC). Outside the western boundary layer, the flow is less intense and reduced to less than 0.1 ms^{-1} without any preferential direction. According to those authors, the intersection of the IWBC and the NBC the physical properties of the AAIW blend and result in the degradation of the AAIW original characteristics.

To better understand the pathways and the dynamics of the Antarctic Intermediate Water circulation in the subtropical South Atlantic, Schmid et al. (2000) explored two models of ventilated thermocline and results of the Semtner–Chervin Parallel. The choice of the classical Sverdrup and ventilated thermocline theories are due to the belief that the directional interaction of the currents formed by the wind and the thermocline circulation, as shown in Figure 1.5. The idea of a combination of a Sverdrupian gyre modified by the Atlantic Meridional Overturning Circulation (AMOC) within the western boundary layer was first hypothesis was raised by Stommel (1965) when comparing analogously with the Gulf Stream system. This supports the importance of the SB as resulting from a baroclinic bifurcation fed by both wind-forced circulation where the opposite direction

to the intermediate thermoaline circulation modifies the BC vertical structure. Their Lagrangian current measurements suggest a gradual increase of the westward transport from 18 Sv at 30° W to 29 Sv at 36° W before they fall back to 19 Sv at 40° W and 13 Sv at 42° W, a consequence of the SB. Comparing Schmid et al. (2000) model results with this lagragian observations and hydrographic data. Their simulations suggest that 90% of the intermediate circulation composition is derived from the westbound transport of the South Atlantic turnaround. Only 50% of the intermediate transport wind forced. Therefore, they concluded that the intermediate circulation is partialy of the wind-driven subtropical gyre system, extending to more than 1300-m depth, which includes AAIW and UCDW.

Figure 1.5: Scheme of the Stommel (1965) schematic of the subtropical western boundary current of the Atlantic model, adapted from Silveira et al. (2000)

We invite the reader to compare the pattern predicted by H. Stommel to that presented in Figure 1.5. In like manner, Rodrigues et al. (2007) applied a reduced-gravity, primitive equation Oceanic General Circulation Model (OGCM) simulation and observational data validation in order to investigate the sSEC bifurcations, in terms of its latitudinal depth dependence and seasonal variability. The Figure 1.6 presents the poleward shift of the sSEC with increasing depth. From the observational data, the annual mean dynamic height and geostrophic flow relative to 1000 dbar shows the sSEC bifurcation in 0–100 m at around 14° S, in 200 m at 18.6° S, in 400 m at 21° S and in 600–800–900 m at 23.6° S, 25.5° S and 26° S, respectively. Therefore, The bifurcation latitude varies by 7° in the top 100 m, by about 3° from 100 m down to 400 m, by less than 1° between 400 m and 600 m, and 2° below 600 m. This variability was related to the local wind forcing, where local positive (negative) wind stress curl produces an anomalous anticyclonic (cyclonic) circulation, whose southward (northward) component near the western boundary causes

the sSEC bifurcation to occur at lower (higher) latitude during the austral spring/summer (winter) months.

Figure 1.6: Annual mean dynamic high ($10^{-1} \text{ m}^2 \text{ s}^{-2}$) and geostrophic velocities for 0 m, 100 m, 200 m, 400 m, 600 m and 800 m (SB level), from Rodrigues et al. (2007).

Schmid and Garzoli (2009) verified the main features of the AAIW circulation from all Argo data available at the time and simultaneous observations of velocity at 800 to

1100 dbar, together with salinity and pressure fields. At this work the impinging sSEC jet was called the Benguela Current Extension (BECE) which resembles the pattern in figure 1.1. The mean SB latitude is found to be at about 28.5°S . One of the addressed questions was the quantification of the split of BECE into the two branches. The observed mean southward transport in the Brazil Current was about 5.2 Sv and the mean northward transport in the IWBC was about 2.2 Sv just north of the bifurcation. The sum of these two transports is 7.4 Sv, which corresponds well with the transport of 7.6 Sv of the BECE near the western boundary, at 45°W . Therefore they concluded that, in mean terms, two-thirds of the water that enters the bifurcation region recirculates in the subtropical gyre.

1.5 Vertical Shear and Vertical Modes

Fourteen years later of the original description of the SB, the study of Legeais et al. (2013) corroborated the findings of Boebel et al. (1999b), using subsurface float data deployed during the French effort DBE/SAMBA repeatedly between 1994 and 2003 (Figure 1.7). These new measurements also added to the region characterization on a mesoscale context, both meridionally and perpendicularly to the continental slope. The isopycnal levels used, $\sigma_0 = 27.0 \text{ kg m}^{-3}$ (AAIW–SACW interface), $\sigma_1 = 31.0 \text{ kg m}^{-3}$ (AAIW core) and $\sigma_2 = 36.83 \text{ kg m}^{-3}$ (AAIW–UCDW interface), agreed with the Boebel et al. (1999b) values. The SB is detected between $25\text{--}28^{\circ}\text{S}$ along the coast, between 46°W and 41°W . The mean velocity calculated for the sector between the slope and 39°W and between 28°S and 21°S was 0.23 m s^{-1} , being only 23% slower than the estimated 0.3 m s^{-1} value by Boebel et al. (1999b). The BC in the southern branch of the SB is characterized by an increase in velocity, from 0.1 m s^{-1} to 28°S to 0.25 m s^{-1} in 36%, with a maximum of 0.75 m s^{-1} (41%). In the north of SB, the IWBC appears with regular pattern over the Santos Plateau (from SB up to about 20°S), with an average velocity of $0.1\text{--}0.15 \text{ m s}^{-1}$ and a maximum velocity of 0.3 m s^{-1} . At the Vitória-Trindade ridge, from 20°S to 15°S , the velocity decays to 0.2 m s^{-1} , due to the highly irregular bathymetry, which destabilizes the flow (Boebel et al., 1999b). This instability generates intense mesoscale activity, resulting in the formation of meanders and vortices, in addition to providing important water exchange to the oceanic interior. At the near-equatorial end, north of 10°S , the IWBC width is around 50 km.

Figure 1.7: (a) Trajectories of the floats that marked the Santos bifurcation, the southward branch in blue and northward green, the magenta trajectories correspond to float immersions greater than 1050 dbar, from Legeais et al. (2013). (b) Averaged float velocities, where blue arrows correspond to averaged velocities less than 5 cm s^{-1} and the yellow arrows to averaged velocities greater than 5 cm s^{-1} , according to Legeais et al. (2013).

Figure 1.8: (a) Vertical velocities section for the TRANSCOBRA transect. (b) Vertical sections of the velocities components, baroclinic and barotropic, for the TRANSCOBRA transect. (c) Silveira et al. (2004) POM baroclinic absolute velocity Vertical section of baroclinic absolute velocity computed by the numerical model POM.

Silveira et al. (2004), presented for the first time the vertical velocity structure of the system BC-IWBC, north of the SB, at 23°S (22.5°S is the northern limit of our domain). He verified the structure of superposed currents with opposite directions, the BC with 400–500 m, maximum surface and a volume transport of 6 Sv and the IWBC reaching 1200 m, with a shift narrow core at 800 m exceeding 0.3 ms^{-1} . To answer how baroclinic this configuration is, they computed the barotropic component and subtracted from the total velocities. They obtained that the velocities at this region are 75–80% baroclinic. To verify this finding, they employed a cross-shelf circulation version of the Princeton Ocean Model (POM) to obtain an absolute baroclinic BC-IWBC field from hydrographic data. By using this method, they substituted the use of the Classical Dynamic Method (CDM) which requires reference level (which produces relative velocity fields). In Figure 1.8b, we observe the calculated baroclinic component, which was found to be very similar to the

total velocity (Figure 1.8a). In Figure 1.8c, we display the absolute baroclinic velocities estimated with the numerical model POM for the same latitude as the other panels.

The water column average eddy kinetic energy accounts for 30–60% of the IKE. Instabilities of the mean flow may give rise to these mesoscale fluctuations. Indeed, the linear stability analysis suggests that the BC System is baroclinically unstable between 22°S and 28°S . In particular, the model predicts southwestward-propagating fastest growing waves (190 km) from 25.5°S to 27.9°S and quasi-standing most unstable modes (230 km) at 22.7°S . These modes have vertical structures roughly consistent with the observed eddy field.

Figure 1.9: Scheme of the south Atlantic circulation at intermediate levels (800-1100dbar, AAIW layer) from Schmid and Garzoli (2009). AC = Agulhas Current, ACC = Antarctic Circumpolar Current, BC = Brazil Current, BEC = Benguela Current, BECE = Benguela Current Extension, IWBC = Intermediate Western Boundary Current, MC = Malvinas Current, NICC = Northern Intermediate Countercurrent, S = stagnant region (poorly ventilated, cross-hatched), SAC = South Atlantic Current, SICC = Southern Intermediate Countercurrent, ZE = Zappioli Eddy.

Rocha et al. (2014) presented an investigation on the downstream changes in the BC vertical structure of four current meter mooring records at the latitudes: 22.7°S , 24.15°S , 25.5°S and 27.9°S placed around the 1299 m isobath. The dataset depict the thickening of the BC. Poleward its vertical extension increases from 350 m at 22.7°S to 850 m at 27.9°S . The BC transport is increased by at least 4.3 Sv (70%) from the 25.5°S

mooring (5.73 Sv) to the 27.9°S mooring (10.02 Sv). Moreover, the SB is associated with a substantial increase in the barotropic component of the BC System. On average, the water column Integrated Kinetic Energy (IKE) is highly baroclinic: 70% on the 22.7°S and 24.15°S moorings. The BC system becomes 54% barotropic at the 27.9°S mooring.

1.6 Hypotheses

Based on the aforementioned works revised in this chapter, while we can develop a first picture of the SB, the comprehension of its tridimensional pattern and space-temporal variability still lack. According to Schmid and Garzoli (2009), in mean terms and between 800-1100 m (AAIW layer), the SB represents the split of the sSEC jet (or BECE) volume, where two-thirds flows southward, and one-third northward. In this we intend to address the temporal variation of the SB branches' transport distribution encompassing the entire intermediate layer, and therefore including the UCDW. Silveira et al. (2004, 2008) and Rocha et al. (2014) have addressed the issue of the vertical structure of the current system along the Southeast Brazilian continental margin. They showed that the BC-IWBC mean jet is virtually baroclinic north of 24°S and the BC thickening with a robust barotropic component. However, no information about the sSEC jet modal composition and the role of the São Paulo Plateau in altering the BS modal structure, and therefore, changing the BC vertical shear and baroclinic instability properties. In the present work we examine the modal structure of the whole SB system and seek for the understanding of the role of the topographic features in the geostrophic pressure mode composition.

We address the two issues raised in the previous paragraphs by testing two hypotheses:

- The sSEC jet splits into two branches, where the southern portion carries two times the volume transport of the northern one, in terms of annual and monthly means.
- The sSEC jet is mostly barotropic and the BC has a increasingly barotropic influence as it goes south, from 22.5°S down to 33°S.

1.7 Objectives

In order to test and demonstrate our hypotheses we aim to describe the mean tridimensional Santos Bifurcation velocity structure and its dynamics, in conjunction with its volume and heat transports, in terms of intra- and seasonal variability. Therefore we need to specially accomplish:

- The Mercator's products validation into the study area.
- Obtain the major volume transports northern and southern of the SB, together with the transport for the sSEC jet.
- Compute the dynamic mode composition and analyze its configuration into the Santos Bifurcation.

Chapter 2

Study Area and Datasets

2.1 Study Area

The study area behold the Southwestern Brazilian continental margin. The latitude bounds go from 33°S to 22.5°S, and the longitude 50°W to 33°W. The eastern boundary of our area of interest is at the open sea, east of the Rio Grande Rise (RGR), where the depth reaches more than 4000 m (Figure 2.1). The RGR is a large near circular seamount of about 3000 km² and 1500 km in diameter. It rises from the ocean floor more than 3000 m tall (around 1000 m from the surface) (Mohriak et al., 2010). Located about 1.500 km offshore it belongs to the international water zone. Another distinctive geological feature in our domain is the SPP (between 23° S and 28° S) Kumar and Gamboa (1979). The plateau have more than 600 km at its widest portion. Its main level is 2,000 m to 3,500 m depth, and underlies the adjacent shelf and slope, within the Santos Sedimentary Basin. Most of the plateau surface is irregular. The southern portion of the plateau is constitute by the São Paulo ridge, an east-west-trending ridge at approximately 28°S. The eastern boundary is delimited by the 3,500 m isobath, and the northern is limited by the Brazilian upper slope, where the 200–500 m isobaths abruptly turn east-west. Its western limit is set by the 2,000 m isobath. The Brazilian continental margin has a divergent feature, originated at the Atlantic opening during the lower cretaceous (Biló et al., 2014a). At our interest area the continental shelf varies from 50 km to 230 km and mean depth of 150 m. The current sedimentary processes at the shelf are known to be linked at the hydrodynamic processes for instance the Rio da Prata plume transport in the inner shelf and the BC "floor polishing effect" at the outershelf, until 100 m (De Mahiques et al., 2010).

Figure 2.1: Bathymetric map from the Brazilian southwestern coast.

Given those facts the São Paulo Plateau it is a relevant topographic feature within our domain. Its structure comprise 212.35 km^2 between the continental slope's base and the depth's marks of 3200–3600 m. At the Campos and Santos Basins it takes place of the continental foothills (Zembruski and Chaves, 1979), reducing the declivity to around 1:4000 - 1:5000. The presence of this plateau offsets the DWBC offshore the given its variations over the continental slope and foothills (Godoi, 2005).

2.2 Cruise OSE 1

The objectives established for the present work requires spatial and temporal coverages in order to characterize annual and shorter-term variability as well as estimate the geostrophic mode composition of the current system. Unfortunately, in our case *in situ* observations are insufficient, especially in terms of temporal coverage, what lead us to resort to models outputs. We believe that some of the available information can be employed in an alternative manner. In order to assure the model represent the features of interest, we validate the model results into our study area. Primarily, we compare a extensive spatial coverage of hydrographic cruise data set, which was provided by the Brazilian Navy, on partnership with the Institute of Oceanography, of the University of São Paulo (IOUSP) Ocean Dynamics Laboratory (LaDO).

Figure 2.2: OSE 1 stations map. The cruise was constructed between 21 March 2000 to 06 July 2000, therefore comprehending the austral late fall and early winter.

The dataset is from the Southwestern Ocean 1 Expedition (Oceano Sudeste 1–OSE1) which provides high resolution quasi-synoptic data, collected by the Brazilian Navy between 21 March 2000 to 06 July 2000. We employ here the temperature and salinity data from the Conductivity Temperature Depth Profilers (CTD). A standart pre-processing for the hidrographic data is applied to the CTD profiles:(i) Spike removal; (ii) Bin averaging in 1 dbar bins; and (iii) Window averaging. This procedure has been regularly applied to most of the studies conducted by the LaDO, e.g. Silveira et al. (2008), Calado et al. (2006), Biló et al. (2014b) and Namiki et al. (2017).

2.3 Mercator Products

The Mercator Ocean GLORYS2V4 is a global ocean reanalysis provided by the Copernicus-Marine Environment Monitoring Service (CMEMS), previously known as GMES (Global Monitoring for Environment and Security). The simulations are performed by the modelling framework NEMO V 3.1 Nucleus for European Modelling of the Ocean (Madec, 2008) in configuration ORCA025_LIM. The ocean component of NEMO has been developed from the *Océan Parallélisé* (OPA) model, release 8.2, described in Madec et al. (1997).

This model has been used on a wide range of applications, both regional or global, as a forced ocean model and as a model coupled to a sea-ice and/or atmospheric models. It has been continuously tested and improved in a multitask way. It includes realistic heat fluxes and initialization with the Levitus data sets. The numerical solutions were improved in

terms of a the barotropic equation with a preconditioned conjugate gradient method, different treatments of static instabilities, and iterative method to compute the vertical diffusion. In the global ocean configuration, OPA 6.0 (1990), some of the major additions were the inclusion of the variable bottom topography and islands, the generalization of the curvilinear formalism and cyclic boundary conditions. In 1992, the model was rewritten (OPA 7.0) in order to provide one routine for each term in the momentum, tracer equations, implicit treatment of vertical diffusion and a turbulent closure.

Its results are obtained by the primitive equations, i.e. the Navier-Stokes equations along with a nonlinear equation of state which couples the two active tracers (temperature and salinity) to the fluid velocity. The code also uses scale considerations, such as: (1) spherical earth approximation; (2) thin-shell approximation; (3) turbulent closure hypothesis (the turbulent fluxes are expressed in terms of large-scale features); (4) Boussinesq hypothesis; (5) hydrostatic hypothesis (this removes convective processes from the initial Navier-Stokes equations, and so convective processes must be parameterized instead); (6) incompressibility hypothesis. An ocean is bounded by complex coastlines, bottom topography at its base and an air-sea or ice-sea interface at its top. Thus, different boundary conditions are applied according to those extents. Surface layers, western boundary currents, equatorial currents, or ocean fronts are strong dynamical features, which requires a higher resolution in its domains to be satisfactorily represented. Regarding coastal dynamics and the North Pole singularity a solution consists of introducing an appropriate coordinate transformations, so it is important to solve the primitive equations in various curvilinear coordinate systems. At the Mercator projection, a tensorial formalism, suited to any multidimensional curvilinear coordinate system, employs three-dimensional orthogonal grids on the sphere with preservation of the local vertical.

The outputs are given in a resolution of 0.25° ($1/4^\circ$), an eddy permitting interval, in a regular grid. It is composed of 75 vertical levels with 1 meter spacing near the surface increasing until 200 meters when deeper than 3000 m, reaching 5900 m. The outputs are available from the period of 1 of January 1993 until the present, although we are employ until only 31 December 2015, once periods later than this are still going under reanalysis. The reanalysis is fulfilled by a data assimilation process, which techniques incorporate Satellite (TOPEX/POSEIDON and ERS-1 satellites) and *in situ* data (CORA4 data base). This merge corrects the model fields to be as close as possible to the real ocean (observations), beyond contributing to the model physic's learning, improving its forecast skills. The validation applied at the reanalysis by the own CMEMS, consists in a series of diagnostic methods specified to each variable as described in Garric and Parent (2013). This quality analysis has shown great results for most of the global ocean and it's major features, however as our desired area is more restrict and particular we also prescribed an validation process detailed at section 2.4. From the provided products we employed the variables temperature, salinity, meridional and zonal velocities, under a daily mean

output for each of the 20 years offered. Considering our area limits (33°S–22.5°S and 50°W–33°W) with the 1/4° model resolution, we have a horizontal grid of 2967 regularly spaced grid points.

2.4 Mercator Output Validation

As previously mentioned, there is no observational data available within the area to achieve our objectives, what lead us to call upon the Mercator global simulation outputs. In order to validate and evaluate the Mercator products for our region specifically, ensuring its representativeness of our scenario, we firstly compared the stream function (ψ) fields obtained from the OSE 1 dataset with the Mercator outputs. As we only embodied temperature and salinity parameters from the cruise data, we estimate the geostrophic velocities by the CDM. This first technique is explained in the Subsection 2.4.1. Secondly, we compared the ψ fields estimated by the CDM with the ψ fields obtained by the Vectorial Objective Analyses (VectOA). This second validation was necessary once we favored to work with the velocities computed by the Mercator model itself. Our preference is due its no requirement of a level of no motion, what makes it further accurate and adjusted to the variation of the no motion level. This second procedure is described in the subsection 2.4.2.

2.4.1 Geostrophic Stream Function

From the available profiles of temperature and salinity, of the OSE 1 cruise, we obtained the estimation of the geostrophic velocity fields by the CDM. The CDM was formulated at beginning of the XX century by Arrhenius and Sandström (1903), and consist of the vertical integration of the thermal wind equation, what results the geostrophic velocity approximation (2.1). The thermal wind equation states that an horizontal movement is induced by the pressure difference between two horizontal positions, due to their respective density vertical structure.

$$v_g(p') - v_g(-p_0) = \frac{g}{f_0} \int_{-p_0}^{z'} \frac{\partial \delta}{\partial x} dp', \quad (2.1)$$

Where p_0 is the isobaric reference level or the level of no motion, δ is the specific volume anomaly, f_0 mean Coriolis parameter. The resultant velocity will always be normal to the pairs of stations a and b (or grid points). To the application of this theory in real data it is needed a discrete formulation by finite sum, where the z axis are pressure levels (p)

$$v_g(p') - v_g(-p_0) = \frac{1}{f} \frac{\Delta \phi_b - \Delta \phi_a}{\Delta x}, \quad (2.2)$$

the $\Delta\phi$ is the geopotential given by:

$$\Delta\phi = \int_{p_0}^p \delta dp \quad (2.3)$$

Once the geostrophic velocities are horizontally non-divergent, we can define a stream function ψ as:

$$u_g = -\frac{\partial\psi}{\partial y}, \quad (2.4)$$

$$v_g = \frac{\partial\psi}{\partial x}. \quad (2.5)$$

Under the condition that $u_g, v_g = 0$ at $p = p_0$ and employing Equation 2.2, we define the geostrophic streamfunction as

$$\psi = \frac{\Delta\phi}{f}. \quad (2.6)$$

A limitation of the CDM calculation is the selection of the reference level, as Biló et al. (2014a) discussed. Despite the improvement in the referencing techniques, those still gather some artful details that retains the task's complexity. The inadequate choice of the level and its reference velocity can lead to inaccurate estimates of the currents. In our case, aggravated by our domain extension, the biggest obstacle is the fluctuation of the level of no motion. The currents' system changes substantially with latitude, predominantly at the Santos Bifurcation region, as reported in Chapter 1. For the sake of simplicity, we choose to select a unique level of no motion, for the northern and southern portions of the system. To choose p_0 we followed the works Biló et al. (2014a); Rocha et al. (2014) and Silveira et al. (2004) which worked with velocities observations in this area. Looking at the northern part of the bifurcation, the current system vertical profile presented 2 velocity inversions. The BC occupies the upper 550 m of water column (flowing southward), the IWBC from 550 m to around 1800 m (flowing northwards), and the DWBC from about 1800 m to the ocean floor (also flowing southward). On the other hand, the southern part has only one flow direction, generated by the fusion of the BC layer to the DWBC layer. Here the BC occupies the water column from the surface until 1800 m. Under this scenario, we can distinguish the reference levels: 550 m and 1800 m. Despite this second level is common between the 2 portions of the Santos Bifurcation, after several tests we concluded that this level is inadequate to represent the IWBC, which becomes not represented in the section maps. However, if we place the reference at 550 m, we will obtain the BC and the IWBC currents well represented, which are crucial to our investigation. Therefore considering that our bifurcation occurs at intermediate depths (originating the IWBC), we have a satisfactory result opting for the level of no motion at 550 m. This reference level has also been applied by Biló et al. (2014b).

After the inference of the ψ , we applied Neumann boundary condition in the interpolation scheme to map ψ . In other words we restrict the ψ to be constant along the solid boundaries. The interpolation algorithm employed is the Objective Analysis based in the Gauss-Markov theorem (Carter and Robinson, 1987). The *a priori* correlation length assign was 1.5° (assuming isotropy), and sampling normalized random error variance of 0.05. The result is shown on the left panels of Figure 2.3 (A and B).

Figure 2.3: A and C: OSE1 objective mapped geostrophic ψ for 200 m and 900 m. B and D: Mercator objective mapped geostrophic ψ for 200 m and 900 m.

For the Mercator ψ field construction, we utilize the model grid points closer to the OSE 1 stations, and for each one of the selected grid points we select the data exclusively from the same dates of the corresponding OSE 1 station (from 05/00 to 07/00). We first linearly interpolate the temperature and salinity from the 75 provided depths by the Mercator products to 1 m depth interval, for the ψ fields estimation by the CDM. We then objective analyze the Mercator geostrophic ψ field following Equation 2.2, similarly to what was done for the OSE1.

Panels A and C of Figure 2.3 consist on the resulting ψ fields for the OSE 1 scenario, at 200 and 900 m depth respectively, and the panels B and D of Figure 2.3 are the results for the Mercator analogous scenario, at the same depths, all obtained by the CDM. These two levels were chosen in order to illustrate the validation, once the 200 m has a well know current pattern what could help the understanding of the obtained patterns and its authentication, and the 900 m once it is our interest level. In the 200 m fields we can observe two prominent negative regions, one at $25^\circ\text{S} - 42^\circ\text{W}$ and another at $23^\circ\text{S} - 40^\circ\text{W}$. Along the coast we observe a positive band and also two positive prominent regions at around $28^\circ\text{S} - 46^\circ\text{W}$ and $27^\circ\text{S} - 39^\circ\text{W}$. Despite the slight differences in the eddy field border, the bulk position and intensity of the cyclons and anticyclons are maintained, which entails the same resulting current pattern, as a meandering BC flowing south along the coast. The ψ values in both settings are similar among the corresponding features.

For the 900 m level, we observe the dipole pattern of the positive north and south negative, with a positive band along the coast. Again, the ψ values agrees well in terms of bulk position and intensity. Both sets present the same current patterns which may reflect the sSEC impinging jet near the SPP. The sSEC enters the OSE1 area at 28°S , 40°W . The qualitative analysis seems to favor the validation of the Mercator product. However, a quantitative comparison is also needed. For this purpose, we follow Pinardi and Robinson (1987) and we calculate the Pearson Correlation Coefficient (PCC) and the Normalized Root Mean Squared Error (NRMSE). The PCC is given by

$$r = \frac{\sum_{i=1}^n (x_i - \bar{x}) - (y_i - \bar{y})}{\sqrt{\sum_{i=1}^n (x_i - \bar{x})^2} \cdot \sqrt{\sum_{i=1}^n (y_i - \bar{y})^2}} = \frac{\text{cov}(x, y)}{\sqrt{\text{var}(x) \cdot \text{var}(y)}} \quad (2.7)$$

where r denotes the PCC, x and y are the two variables to be compared (OSE 1 and Mercator fields, in our case), cov is the covariance and var the variance. the index i refers to the grid point, n is the total number of points and $\bar{x}, \bar{y}$ are the mean fields. The results measure of the linear correlation between two variables x and y . Its value range from -1 to +1, where 1 is a perfect positive linear correlation, 0 is no linear correlation, and -1 is a perfect negative linear correlation. In other words, with this evaluation we can verify if the fields are in phase, or how the features match in terms of horizontal pattern. By this estimation, we can evaluate, in terms of percentage, how the mercator ψ amplitude, or values, differ from the observed ones. The NRMSE is computed by:

$$\text{RMSE} = \sqrt{\frac{\sum_{n=1}^N (\hat{x}_i - x_i)^2}{N}} (\sum_{n=1}^N \hat{x}_i^2)^{-1}, \quad (2.8)$$

were N is the total number of data points, $\hat{x}_i$ is the observation value (OSE 1 ψ) and x_i is the model value (Mercator ψ).

By inspecting the PCC values (Table 2.1), we can verify an high positive correlation, which indicates a high level of agreement between the patterns of the OSE1 and the Mercator results in both compared levels. The NRMSE shows around 10% of difference in the Mercator ψ values, or 90% of the amplitudes similarity geostrophic. We can also observe a very favorable comparison in the range of velocities, showed in Table 2.2, where the absolute maximum and the mean values of the velocities are in the same order of magnitude.

Table 2.1: Values of the Pearson Correlation Coefficient and the Normalized Root Mean Squared Error, for 200 m and 900 m, between the OSE 1 and Mercator scenarios.

	200 m	900 m
PCC	0.84	0.78
NRMSE	0.11	0.10

Table 2.2: Comparison of the maximum and mean velocities (in m s^{-1}) obtained from the ψ fields for the OSE 1 and Mercator scenarios.

	OSE 200 m	MER 200 m	OSE 900 m	MER 900 m
Maximum Velocities	0.17	0.17	0.09	0.07
Mean Velocities	0.03	0.03	0.02	0.01

2.4.2 Velocity derived Stream Function

The geostrophic ψ validation was successful, meaning that the Mercator output reproduce the temperature and salinity horizontal gradients, and therefore the geostrophic velocities. We should also ask whether the model velocities also reproduces the geostrophic field exhibited in Figure 2.3. In order to do so, we employed the so-called Vectorial Objective Analysis (VectOA) (Biló et al., 2014a; Silveira et al., 2000) to map ψ from velocity fields. The technique was constructed based on the fact that if there is a relationship between ψ and (u, v) , as shown in Equation 2.5, there should also be a relationship among the spatial correlation function. In an OA scheme, the weight function is proportional to the autocorrelation function of the data via the Gauss-Markov theorem. Therefore we present next the PCC and NRMSE for the comparison between the OSE field with the Mercator velocity derived field, in Figure 2.4 and Table 2.3. Also we present the comparison between the Mercator objective mapped geostrophic ψ and the Mercator objective mapped velocity derived ψ , to observe the variation caused by the different methods itself, at Figure 2.5 and Table 2.4.

Figure 2.4: OSE1 objective mapped geostrophic ψ for 200 m and 900 m. B and D: Mercator objective mapped velocity derived ψ for 200 m and 900 m.

Table 2.3: Values of the Pearson Correlation Coefficient and the Normalized Root Mean Squared Error, for 200 m and 900 m, between the OSE 1 and Mercator by the VectOA.

	200 m	900 m
PCC	0.79	0.70
NRMSE	0.21	0.15

Figure 2.5: A and C: Mercator objective mapped geostrophic ψ for 200 m and 900 m. B and D: Mercator objective mapped velocity derived ψ for 200 m and 900 m.

Table 2.4: Values of the Pearson Correlation Coefficient and the Normalized Root Mean Squared Error, for 200 m and 900 m, between the Mercator ψ geostrophic and velocity derived.

	200 m	900 m
PCC	0.82	0.73
NRMSE	0.19	0.14

The PCC and NRMSE between the OSE field with the Mercator velocity derived field showed a less phase correlation and more amplitude disparity, as expected once they are obtained by different methods, however their values still reveal high correlation and low error between them. Even though the comparison between the Mercator objective mapped geostrophic ψ and the

Mercator objective mapped velocity derived ψ present a better correlation and less error, relative to the OSE versus the Mercator by Vectoa, once both derive Mercator products. The Mercator products satisfactorily represent our region of interest, regarding temperature, salinity (therefore density as well) and absolute velocities, both in terms of pattern and intensity. Thus we consider the Mercator products validated and the remainder of the document will employ the global model reanalysis to pursue our objectives.

Chapter 3

The Mean Circulation Pattern

Succeeding the validation, we investigated the mean circulation pattern, both from horizontal and vertical perspectives, in pursuance of a three dimensional description of the SB. The set up is built by calculating the mean of the Mercator velocity products from the total time period available (from January 1993 until December 2015). The mean is estimated for each grid point, the provided grid of $1/4^\circ$ resolution, inside 22.5° – 33° S and 50° – 33° W, our study area.

3.1 Horizontal Perspective

In this section, we analyze the horizontal configuration of the ψ fields, computed from the mean velocities using the VectOA method. In order to show the SB structure, we display ψ for levels ranging from 200 m until 3000 m. Our major goal with this analysis is to identify the presence and position of the SB according to the change in depth. We define as the bifurcation position the point where the zero streamline reaches the respective isobath limit perpendicularly, with at least 1° separation distance.

At the 900 m depth is the salinity minimum of the water column which is used to delimit the AAIW core. Therefore, we pre-establish this level as our reference level. At this depth we can observe in Figure 3.1 the SB occurring at 46.8° W – 28.7° S. We mark the 900 m BS location with a green dot (Figure 3.1) in all horizontal plots as a referral to help the comparison between the maps.

We observe that intermediate sSEC at 900 m reaches the continental margin virtually zonally, at about 29° S. The sSEC split occur at 44° W. The southern branch smoothly turn south and flows poleward bordering the continental margin. On the other hand, the northern branch performs a cyclonic loop as it flows equatorward. This cyclonic loop was first presented by Boebel et al. (1999b) and confirmed by Legeais et al. (2013).

Figure 3.1: Mercator objectively mapped ψ (upper panel) and velocity field (lower panel) at 900 m. The green dot is the SB location.

Moreover, the equatorward flow associated with the IWBC gets organized as jet north of 27° S. For instance, this is consisted with the mooring data analysis conducted by Rocha et al. (2014) and Biló et al. (2014a). These works indicated that the IWBC is fully formed at 25° S.

Figure 3.2 presents the ψ for each of the water mass domains. At 20 m (upper left panel) we can see the BC flowing poleward along the continental margin, and reaching more than 0.5 m s^{-1} in its core, a magnitude which agrees with the literature (Silveira et al., 2000; Rocha et al., 2014; Biló et al., 2014a). At this level the BC has approximately 1 km wide. Also, we can note the

sSEC branch approaches the coast from the east, at around 29° S, as a broad weak flow. At 400 m (upper right panel), we notice the weakening of the BC north of 27° S, antagonistically to its southern portion that remains well established and intense. At this depth, we also note the sSEC approximation centered at 29° S. Unlike the TW field, we notice that the BS occurs at more northern latitude, around 25° S.

Figure 3.2: Mercator objectively mapped ψ for 200, 400, 600, 900, 1500 and 3000 m.

At the upper 400 m, the sSEC pattern obtained is consistent to the Stramma and England (1999) description. At the depth range from 600-1200 m (central panels and the lower left panel,

Figure 3.2) the observed pattern suggests that the sSEC jet is intensified at intermediate depths. Deeper than 900 m the jet starts to lose strength until it vanishes around 2000 m. It can reach 5 cm s^{-1} at 900 m. Therefore the SB is present within the AAIW and UCDW domains. Deeper 1200 m the DWBC start to reveals itself flowing poleward countouring the SPP from offshore, as the IWBC flows on top of it. The DWBC appears since the northern latitude of our study area. At 3000 m (lower right panel, Figure 3.2), the DWBC establishes with approximately 0.05 m s^{-1} surrounding the bathymetry.

From the perspective raised in this section we verify that the SB position is virtually constant within the intermediate layer, around the reference point ($46.8^\circ\text{W} - 28.7^\circ\text{S}$). At shallower layers (400-600 m) we can verify the weakening of the BC, northern than 28°S and no presence of the SB. At 400 m we can find a weaker and shallower bifurcations, yet it is not completely pictured into our domain. This feature was reported by Rodrigues et al. (2007) as one of the sSEC shallower bifurcations. As presented in Chapter 1, one of the assumptions matured from earlier studies about the sSEC pattern is its latitudinal shift to the south as depth increases. This shift gives rise to the different northern bifurcations of the sSEC.

Bathymetry Influence

The bathymetry of our region of interest appears to be determinant to the formation and shaping of the SB and the IWBC origin site. As mentioned in the Section 3.1, the SB and its northern branch cyclonic excursion seem to be introduced by the SPP. Furthermore the presence of this topographic feature can also be the cause of the SB displacement to the south, disconnecting it from the shallower bifurcations. From the horizontal perspective presented at the Section 3.1, in the 600-1200 m depth range, we can identify the sSEC jet. Deeper than this, the DWBC countouring the plateau region from its offshore border. Certainly, the explanation of this phenomenon relies on Potential Vorticity conservation.

In Figure 3.3, we observe the 900 m ψ contour in black, containing the sSEC impinging jet and the SB originating the IWBC and the south BC thickening. Underneath the ψ contour is the bathymetry chart in shades of blue (higher depths) to yellow (lower depths), contoured by 500 m spaced isobaths in gray (1500, 2000, 2500, 3000, 3500, 4000 m). In this map we can qualitatively evaluate how coincident are the stream lines at the SB level to the bathymetric contours. The sSEC jet is channeled by the Rio Grande Rise elevation and the SPP around 29° S . It is clear its countouring pattern in the plateau, yielding to the anticyclonic loop described by Boebel et al. (1999b) originally. The IWBC remains close to the slope and velocities are low on the top of the SPP.

Figure 3.3: Mercator Contour map of ψ at 900 m and bathymetric contours. The ψ (bathymetry) contours are displayed in black (grey)

3.2 Vertical Perspective

In this section, we present a different perspective for the mean current system analysis. We planned four transects, as displayed in the Figure 3.4, to produce velocity sectional representations along them. Our intent is to capture the main component patterns of the system. The cyan transect is drawn to capture the equatorward arm of the SB. The magenta transect aims to capture the poleward branch of the SB. The green addresses the impinging sSEC jet, focusing in the portion that composes the SB. Lastly, the yellow transect exhibits the along-1900-m isobath current pattern, where we intend to follow the direction changes of the western current system through the western boundary. We refer to these transects as North (N), South (S), East (E) and Along Isobath (AI), respectively. For each section we present the mapping of the rotated mean velocity component. We also present the isopycnal interfaces (Cyan lines), and provide the volume and heat transports for each water mass. The rotation of the u and v velocity components is necessary once we want to analyze the velocities crossing the section. Only on the AI section, we use the along-sectional velocity component. In order to rotate the velocities, we estimate the angle between an approximated straight line to the transect and the model standard zonal and meridional velocities. The rotation was made using the equation 3.1.

Figure 3.4: Map of the location of sections used to describe the vertical current patterns. In green, the E section, in magenta the S section, in light blue the N section and in yellow the AI section.

We follow the literature to choose the isopycnal which represents the 50% of mixture between two continuous water masses Stramma and England (1999); Mémery et al. (2000). We present the isopycnal values as density anomaly: potential density minus one thousand (σ_θ). The σ 's subscriptions refer to the level pressure reference (0 - Atmospheric level, 1 - 1000 m and 2 - 2000 m.). For the water masses constrained into our system we have the values of: The TW, from the surface until the isopycnal $25.6 \sigma_0$, the SACW, from the $25.6 \sigma_0$ until $26.9 \sigma_0$ (Biló et al., 2014a); the AAIW right below until $32.2 \sigma_1$, followed by the UCDW until $37.0 \sigma_2$, and finally the NADW until 4000 dbar (Stramma and England, 1999).

$$\begin{aligned}
 \vec{v} &= u + iv \\
 \vec{v}_{rot} &= e^{i\theta} \vec{v} \\
 v_{rot} &= \text{Im}(\vec{v}_{rot}) \quad u_{rot} = \text{Re}(\vec{v}_{rot}),
 \end{aligned} \tag{3.1}$$

$\vec{v}$ is the velocity vector field, $\vec{v}_{rot}$ is the rotated vector field, the θ is the rotation respective angle for each transect, v_{Rot} is the cross sectional velocity component and u_{rot} is the along sectional velocity component.

At the North section, we can verify the BC flowing south (Blue shading), and occupying the upper 500 m depth, from the coast until 200 km offshore. Its core is located around 100 km from the coast. The BC jet has approximately 100 km of width, where velocities reach $>0.35 \text{ m s}^{-1}$. The BC transports TW and SACW at this latitude. The TW (SACW) volume transport is 2.3 Sv (5.2 Sv), while heat transport is 0.22 PW (0.36 PW). The total BC volume (heat) transport is, therefore, 7.49 Sv (0.58 PW). Those values agrees with Silveira et al. (2004), Biló et al. (2014a), Schmid and Garzoli (2009), Zemba (1991) and Müller et al. (1998) descriptions. Right underneath the BC, we find the IWBC (Red contours), from 500 until 2000 m depth. Its core is located next to the slope. The overall width is around 120 km. Its velocity magnitude can reach 0.08 m s^{-1} . Here, the water masses transported are the AAIW and UCDW, coming to 5.41 Sv and 1.64 Sv (0.09 PW and 0.02 PW), respectively. The IWBC total volume and heat transport are 7.05 Sv and 0.11 PW towards the equator. The maximum speed value is consistent with Legeais et al. (2013) observations at this latitude, the other characteristics cohere to the Silveira et al. (2004); Biló et al. (2014a); Schmid and Garzoli (2009); Zemba (1991); Müller et al. (1998) documentations. As we are over the SPP, we do not expect to see the DWBC, which is pushed offshore as Figure 3.2 indicates and Silveira et al. (2008) speculated.

Figure 3.5: Cross-transect velocity, volume and heat transport for the N section. The isopycnals interfaces between water masses are presented in cyan with their respective σ_θ values. Southward velocities are in blue.

At the S section we verify the southern branch of the SB, where the BC has thickened by the sSEC jet input flowing poleward. The whole water column flows poleward. The BC is present from the surface until 1500-2000 m depth, which it merges with the DWBC, until reaching the ocean floor. At the S transect location, the shelf breaks at about to the shore, thus the BC core appears around 30 until 100 km offshore, reaching 500 m depth and 0.26 m s^{-1} . The DWBC core

is placed about 2700 m depth, reaching 0.08 ms^{-1} . At this scenario the BC transports the TW, SACW, AAIW, and UCDW, with 1.8 Sv (0.17 PW), 10.91 Sv (0.67 PW), 6.49 Sv (0.13 PW) and 2.46 Sv (0.03 PW), totalizing 21.66 Sv (1.0 PW). Alongside the DWBC carries the NADW with 4.25 Sv (0.05 PW).

Figure 3.6: Cross-transect velocity, volume and heat transport for the S section. The isopycnals interfaces between water masses are presented in cyan with their respective σ_θ values. Southward velocities are in blue.

The E section exhibits the impinging sSEC jet, which sets the SB (Blue contours show the negative jet, approaching the Brazilian coast). The elevation at the southern end of the E section is related to the top portion of Rio Grande Rise, and the northern elevation is the SPP. The sSEC jet occupies the entire water column, exhibiting two cores: a shallower one from 200 until 1200 m, and a deeper one from 2500 until 3000 m. The shallower core is wider, 100 km, and the jet shape spreads for more than 500 km. The deeper nucleus has around 35 km, and the jet is constrained into 350 km wide. Both cores reaches around 0.045 ms^{-1} . The shallower core is mainly within the AAIW portion of the water column; the deeper within the NADW domain. All the water masses from these region are carried by the jet, TW, SACW, AAIW, UCDW, NADW, with 0.3 Sv (0.03 PW), 6.48 Sv (0.36 PW), 10.22 Sv (0.19 PW), 5.55 Sv (0.07 PW) and 7.75 Sv (0.08 PW), totalizing 30.3 Sv (0.73 PW). This values agrees with the previous works Schmid and Garzoli (2009); Rodrigues et al. (2007); Legeais et al. (2013); Zemba (1991); Müller et al. (1998).

Figure 3.7: Cross-transect velocity, volume and heat transport for the E section. The isopycnals interfaces between water masses are presented in cyan with their respective σ_θ values. Westward velocities are in blue.

The AI section displays the currents configuration, from the surface until the ocean floor, bordering east the 1900 m isobath. At this setup, different from the other sections, we utilize the along-section velocity component. Thus, in the upper 500 m, we observe the BC flowing poleward, heading to the left side of our panel (Blue shadings).

Below 500 m, we can see that part flows poleward, the southern portion (left) from the read line, thickening the BC and the other part flows equatorward, at the northern portion (right), originating the IWBC. The conglomerated patterning on the shading is caused by the average signal of meandering of the currents along the transect. The dashed red line marks the 900m level. The black line contours represent the zero velocity band, which highlights the velocity field inversions, as we focus here on observing the velocity inversion configuration. North of 28°S , the velocity inversion between BC and IWBC indicates the use of 550 m as level of no motion for the northern portion of the system (north of the SB) is adequate. Furthermore, it becomes clear that the velocity which marks the SB occurs around 500 m at 28.2°S and migrates south with depth until 1200 m, where it is virtually constant around 29.9°S until the zero velocity line intercepts the ocean floor. At this location the ocean floor is the southern limit of the SPP. We can see the plateau occupying 400 km distance from the beginning of the section line (29.9°S), up to more than 1300 km northward, where it merges with the curve of the continental margin at the Rio de Janeiro state (23°S). From the SB down to the southern limit of the domain (29.9°S - 32.5°S) we observe the bathymetry drop until 3000 m.

Figure 3.8: Along-transect velocity for the AI section. The isopycnals interfaces between water masses are presented in cyan with their respective σ_θ values. Southwestward velocities are in blue.

3.3 Volume Transport Intercomparison

Now we should compare the transports obtained for the 23-year mean volume transport from the Mercator products to those found in the literature. Schmid and Garzoli (2009) calculated the transport for what they considered as "the intermediate layer", i.e., from 800 to 1100 dbar in their climatology-building study using ARGO floats.

However, confirming studies based on the WOCE data base (Stramma and England, 1999; Mémery et al., 2000), our analysis shows the intermediate layer is composed of the AAIW combined with the UCDW. The circumpolar water indeed follows the motion of AAIW and do thick the IWBC at BS northern branch as well as the BC at southern branch (please, return to Figures 3.5 and 3.6 for details). Therefore we proceed with our analyzes taking into account these the AAIW (~ 550 m) and UCDW lower limit (~ 1300 m).

Schmid and Garzoli (2009) observed a mean southward transport of about 5.2 Sv for the intermediate portion of the BC at the southern flank of the BS, and a mean northward transport of about 2.2 Sv carried by the IWBC, just north of the bifurcation. Their sum is 7.4 Sv, which corresponds almost exactly with the transport of 7.6 Sv they estimated for the sSEC at 45°W . The authors then concluded that two-thirds of the water entering the bifurcation region recirculates in the subtropical gyre (going poleward). An earlier estimation conducted by Schmid et al. (2000) stated that three-quarters of the sSEC veers south. Müller et al. (1998) estimated the (700m-1450m) AAIW-UCDW transport from two WOCE current meter mooring transects as being 3.4 Sv flowing southward at 29°S and 1.3 Sv at 24°S . Núñez-Riboni et al. (2005) computed a mean

volume transport of about 4 Sv for the BC and 1 Sv for the IWBC. The differences (about 25%) are probably due to a combination of temporal variability and a much larger set of observations available for the Schmid and Garzoli (2009) work. Nevertheless, both works converge in the sense that the southern BS branch receives most of the volume transported by the sSEC.

Taking in consideration our mean estimates for the AAIW domain, we get 5.4 Sv for the North section, 6.5 Sv for the South section, and 10.2 Sv for the sSEC jet. If we add the transports of the North and South branches, 11.9 Sv. As described in earlier sections, we do not include the sSEC portion which diverges and countours the Rio Grande Rise from the south because it does not participate directly to the BS feature, but it indeed fattens the BC towards the confluence. This branch carries about 2.5 Sv within the study area. Contrarily to the earlier observational studies, we obtain from the validated MERCATOR reanalysis that the Southern (Northern) BS branch carries 54% (46%) of the sSEC transport poleward (equatorward).

3.4 Annual and Intraseasonal Variability

As we have discussed in this work, there very few studies addressing the AAIW flow in the South Atlantic Ocean. There is virtually no study conducted described the variability of the AAIW and/or the intermediate layer transport near the southwestern boundary. Certainly, this absence is related to the observational basis being not enough. However, there is no numerical modeling work to best knowledge of this author todate which addresses the time variations of the AAIW transport near the South Brazilian continental margin. Solely, the work by Rodrigues et al. (2007) dealt with the variation of the time variations of the BS locus.

In this section, we ask very simple but fundamental questions about the BS temporal variability:

- Is there an annual cycle to the BS location?
- Can seasonal or intraseasonal variations on the sSEC northern and southern branches be detected in the model output?

To address the first question, we compute a seasonal climatology of the Mercator output series and present the horizontal panels for ψ at 900 m (Figure 3.9). We there observe a displacement of less than 1° from the SB main locus, which reasonably agrees with the low variability found by Rodrigues et al. (2007) for the 500-1000m depth layer.

In order to the second question, we compute the seasonal transports for the East, North and South sections. We present the results on Table 3.1. the North and South transports add up the sSEC transport for each season: 11.4 Sv (autumm), 11.4 Sv (winter), 11.2 Sv (spring), and 13.3 Sv (summer). Interestingly enough, their proportions varies. During the Autumm and Summer, the Northern branch transport is 85% and 79% of the Southern branch transport, respectively. For the winter and spring, the North transport is only 61% and 45% of the Southern transport value, respectively. In other numbers, during theses seasons, makes the Northern branch transport is approximately, $1/3$ and $1/2$ of the Southern branch transport.

Figure 3.9: Seasonal plots for the 900 m horizontal field. The green dot represents the reference position of the SB for the mean scenario.

The Figures 3.10 and 3.11 illustrates the climatological run monthly values for the volume and heat transports within the AAIW portion of the water column. In those maps, we display the South, North and East section transports in red, blue and black, respectively. The green dashed line is the sum of the North and South section AAIW transports. In analyzing this graph, we first observe that the dominant variation in the Eastern Section transport is semi-annual and not annual. There are clearly three maxima with values around 11 Sv in March, June-July and November. We also observe that the transport sum of the North and South sections is overall higher than the curve for the Eastern Section. The offset is about 1.7 Sv, a consequence of not having included the southern portion of the sSEC which countours the Rio Grande Rise.

The major contribution in the sum comes from the South Section. Interestingly enough, there is a swap in May and June (end of autumn and beginning of the winter) in terms of the branch contribution. During this period, the North section transport is higher, representing around 68% of the sum. The variation in terms of proportions is clearer seen on Figure 3.12. This pattern can be possibly related to the BC transport minimum in austral winter at 24, 35, and 38°S, as described by Schmid and Majumder (2018), once the decrease of the BC transport would result

Table 3.1: Transports for each water mass during Autumn (Aut), Winter (Win), Spring (Spr), Summer (Sum) and annual mean (Mean). All values are presented in Sverdrups ($1 \text{ Sv} = 10^6 \text{ m}^3 \text{ s}^{-1}$).

<i>EAST</i>	Aut	Win	Spr	Sum	Mean
TW	0.3	0.1	0.2	0.5	0.3
SACW	6.4	6.3	6.6	6.9	6.5
AAIW	10.1	10.2	10.2	10.6	10.2
UCDW	5.5	5.5	5.3	5.2	5.6
NADW	8.0	7.8	7.3	7.6	7.8

<i>NORTH</i>	Aut	Win	Spr	Sum	Mean
TW	2.2	2.2	2.6	2.3	2.3
SACW	5.2	6.5	7.7	5.3	5.2
AAIW	5.3	4.3	3.5	5.8	5.4
UCDW	1.6	1.4	1.3	1.8	1.6
NADW	-	-	-	-	-

<i>SOUTH</i>	Aut	Win	Spr	Sum	Mean
TW	1.6	0.3	1.8	2.3	1.8
SACW	11.0	13.1	12.1	11.1	10.9
AAIW	6.2	7.1	7.8	7.4	6.5
UCDW	2.3	2.6	2.7	2.9	2.5
NADW	4.1	4.4	4.4	4.7	4.3

Figure 3.10: AAIW volume transport.

in a decrease of the South branch transport and in an increase of the IWBC transport. During the October and November months, the split reaches its highest amplitude, where the South section is responsible for more than 74% of the total value.

We now combine the transports of the AAIW and UCDW to obtain the intermediate layer transports. The agreement between the East transport, or the sSEC transport, and the sum of the North and South branches transports improves, as we can see at Figures 3.13 and 3.14. We credit this improvement to the better fit of this two water masses limits with the thickened BC

Figure 3.11: AAIW heat transport

Figure 3.12: Volume and heat transport as percentage from the North and South to the East.

Figure 3.13: AAIA+UCDW Transport.

and IWBC. This configuration maintains the North and South ratios, shown in the Figure 3.15.

Figure 3.14: AAIA+UCDW Heat Transport.

Figure 3.15: Transport Percentage from the North and South to the East.

Propagating Signals

Aiming to Investigate a little bit further of the intraseasonal variability of the SB, we analyzed the Hovmoller diagram of the main currents of the system in order to observe their time evolution. To do so we opted to inspect the velocities of the extended AI Section (from 20°S to 40°S) and a new drawn zonal line at 29°S, from 46°W. to 5°W. The AI line was used in attempt to contemplate the BC and IWBC currents, and the zonal line to picture de sSEC jet. At the zonal line we could capture a well-established propagating eastward signal, shown in Figure 3.16. To obtain this structure wavenumber and frequency ($k-\omega$) we resort to the Fast Fourier in a two dimensional approach (FFT2D). The FFT2D a two dimensional traditional FFT, which is a mathematical technique that decomposes a given signal, by a discrete Fourier Transform method, into its composite of Harmonics. In this manner we can convert the obtained senoids space-time into their $k-\omega$ energy spectrum. Therefore the obtained a k of 442.75 km and 122 days ω , thus a phase speed of 0.42 ms^{-1} . This signal resemble the signals described at upper depths by Majumder et al. (2018), however our analyzes were short and need more robustness that can be regarded in future works. On the other hand, for the AI Section we obtained mainly

a dual regime, divided exactly at the SB region, as we presented in Figure 3.17. The attempt to filtrate and obtain this regimes principal hamonic compositions was not successful, what is also a goal to future works. It is interesting to remember that the depth can act as a bandpass filter, however the refraction and reflection can originate interferences in the signal identification.

Figure 3.16: Hovmoller for the a zonal line at 29°S .

Figure 3.17: Hovmoller for the AI Section line extended. The black dashed line is the 29°S latitude, which is the SB main latitude.

Chapter 4

Dynamical Mode Analysis

In this section, we discuss about the orthonormal dynamic mode composition in order to uncover, map the individual modes contribution and seek a dynamical tracer to the SB. To obtain the modal configuration of our fields, we follow Houry et al. (1987), by solving the eigenvalue formulation for the long hydrostatic baroclinic wave equations, and under we also need to assume inviscid, hydrostatic and under rigid lid and flat bottom approximations. We can assume this hydrostatic motion thanks to the larger horizontal proportions of our features of interest [$O(10^5 \text{ m})$], compared to its depth layer size [$O(10^3 \text{ m})$]. Under the variable separation process, the momentum equations can be represented into independent expressions of the vertical structure and the horizontal. The velocity components u and v are decomposed into a infinite series of orthogonal functions related to the vertical structure (the eigenvectors or the vertical dynamic modes), composing base functions, which are multiplied by respective amplitude functions dependent of the time and horizontal variations, as demonstrated by the equations 4.1, 4.2, 4.3, 4.4 and 4.5.

$$\psi = \psi_i(x, y, t)\mathbb{F}_i(z) \quad (4.1)$$

$$\begin{aligned} u(x, y, z, t) &= \mathbb{U}_i(x, y, t)\mathbb{F}_i \\ v(x, y, z, t) &= \mathbb{V}_i(x, y, t)\mathbb{F}_i \end{aligned} \quad (4.2)$$

$$W(x, y, z, t) = \mathbb{W}_i(x, y, t) \frac{1}{N^2(z)} \frac{\partial \mathbb{F}_i(z)}{\partial z}, \quad (4.3)$$

$\mathbb{F}_i$ are the i th-mode vertical mode (or eigenvector), the $\psi_i, \mathbb{U}_i, \mathbb{V}_i$ and $\mathbb{W}_i$ are i th-mode stream function, zonal velocity, meridional velocity and vertical velocity amplitudes. In such way, we obtain equations of Sturm-Liouville for the amplitude and vertical structure evolution.

To obtain the dynamical modes $\mathbb{F}_i$ we solve the vertical structure equations 4.4 and 4.5, assuming rigid lid ($z=0$) and flat bottom ($z=-H$).

$$\frac{\partial}{\partial z} \frac{1}{N^2(z)} \frac{\partial \mathbb{F}_i(z)}{\partial z} + \lambda_i \mathbb{F}_i(z) = 0 \quad (4.4)$$

$$\frac{\partial \mathbb{F}_i(z)}{\partial z} = 0 \text{ @ } z = 0, -H \quad (4.5)$$

λ_i are the eigenvalues, f is the Coriolis parameter under β plane approximation ($f = f_0 + \beta y$, where f_0 is the Coriolis value for the field central latitude and β is the meridional variation of f , or $\frac{\partial f}{\partial y}$) and H is the local depth. It is possible to find infinite solutions for $\mathbb{F}_i$ ($i=0,1,2,3,\dots,\infty$), each related to λ_i , which is real and discrete. The Barotropic mode $\mathbb{F}_0$ is given by a single value through the entire water column ($\mathbb{F}_0(z) = 1$) and N^2 is the Brunt-Väisälä frequency, given by:

$$N^2(z) = \frac{-g}{\rho_0} \frac{\partial \rho(z)}{\partial z} \quad (4.6)$$

The ρ is the water density, g is gravity acceleration and N^2 is the Brunt-Väisälä frequency. Therefore in order to obtain the dynamic modal composition we only require the vertical profile of the local (at each grid point) $N^2(z)$ and the Coriolis parameter. We detail this process at the Subsection 4.1.

4.1 Brunt-Väisälä and Deformation Radii

As demonstrated in Equation 4.4, in order to obtain the modal composition, we need to compute the local Brunt-Väisälä frequency (Equation 4.6), for each grid point into our domain. To infer N^2 , we calculate the density vertical profile using the temperature and salinity vertical profiles provided by the Mercator products. After obtaining the N^2 profiles we smooth them using a hamming window. Figure 4.1 exhibits the "raw" and the "smothed" N^2 profile for one open ocean grid point.

The i th eigenvalue (λ_i) obtained by solving Equations 4.4 and 4.5 in the formulation 4.5, are proportional to the inverse of the i th deformation radius squared:

$$\lambda_i = f_0 R D_i^{-2}, \quad (4.7)$$

despite for the first mode (mode 0, Barotropic corresponded). Under the rigid lid approximation, the zeroth (or barotropic) deformation radius coming out from the solution of the Sturm-Liouville problem (Equations 4.4 and 4.5), $R D_0 = \inf$. In other words the barotropic deformation radius is much larger than the baroclinic ones. We opt to calculate the appropriate $R D_0$ by:

$$R D_0 = \sqrt{\frac{gH}{f_0}}. \quad (4.8)$$

Figure 4.1: Exemplification of the smoothing of the N^2 profile at one open ocean grid point.

On account of this interest we construct deformation radius maps for the first three modes: barotropic ($i = 0$), first baroclinic ($i = 1$) and second baroclinic ($i = 2$), following the equations 4.7 and 4.8. Therefore we computed the modal composition for each grid point, obtaining the Deformation Radii (intern) for all the dynamic modes computed (Baroclinic), at each of the points. For the Barotropic Deformation Radius estimation we, likewise at each grid point, calculated the equation 4.8 employing the bathymetry of the respective grid point (i.e. maximum depth, H).

Figures 4.3, 4.4 and 4.5 exhibit the Rd_i distribution over the study area for $i = 0, 1, 2$ at every $0.5^\circ \times 0.5^\circ$. Each box in the maps correspond to an average of 4 grid points of the original Mercator product. The "local" approach to the Rd analysis is presented on climatological works (Houry et al., 1987) and/or on more dynamical investigations (Tulloch et al., 2011). The Rd_i maps are important as guidance for interpreting the dynamics due to changes in local depth (due to topography features) and in Stratification. They are also useful for simpler application. This estimate, follows Flierl (1978):

$$[\mathbb{U}_i, \mathbb{V}_i] = \frac{1}{H} \int_{-H}^0 [u, v] \mathbb{F}_i dz. \quad (4.9)$$

The truncated modal recomposition is obtained employing Equations 4.1, 4.2 and 4.3, as we shall see, in the next section, under the approximation $i = 0, 1, 2, 0 - 1, 0 - 1 - 2; 1 - 2$. See Figure 4.2 for example.

Figure 4.2: Modal decomposition and recombination for the sSEC at 43°S, 28.75°W: a) N^2 ; b) The first three dynamical modes; c) Modal decomposition of zonal velocity profile for $i = 0, 1, 2$; d) Modal recombination truncated to $i = 0-1$, $i = 0-1-2$ and $i = 1-2$.

Figure 4.3: Map of the zeroth deformation radius for the study area.

Figure 4.4: Map of the first deformation radius for the study area.

Figure 4.5: Map of the second deformation radius for the study area.

4.2 Mean Circulation Modal Structure

After obtaining the modal decomposition for each point of the domain, we arrange the results into 900 m horizontal panels and into transects for the respective sections we worked on Section 3.2. Subsequently, we estimated the NRMSE recalling the same formulation we implemented at the Mercator validation Section 2.4, now following Silveira et al. (2000). This estimation allow us to understand the relative importance of each mode, where the obtained value can be interpreted as the percentage of the variance of the total velocity which is represented by each mode. For a more intuitive analyzes we subtracted the result from one and multiplied by one hundred $[(1 - NRMSE) \times 100]$. In doing so, we obtain the Explained Variance (EV), for the individual modes decompositions or a combination of them as a truncated decomposition. Therefore we could determine the modes which we should focus our analyzes. The formula of the NRMSE is given by Equation 2.8, this time $\hat{x}_i$ being the total velocity component for the section (or ψ of the total velocity for the horizontal field), and x_i the velocity reconstitution of each particular mode or their cumulative sum (or the ψ recomposition). Results of this analysis is shown on Table 4.1. The first tree modes suffice to nearly reproduces the total velocity structure. Moreover, we present the calculated modes fields settings for each modes individually and their accumulative sum: the barotropic (BT), first baroclinic (BC1), second baroclinic (BC2), the BT and BC1 combined and the BT, BC1 and BC2 combined.

Table 4.1: NRMSE and EV for 900 m horizontal field between the total Mercator velocity and the reconstituted velocities sets.

	BT	BC1	BC2	BT+BC1	BT+BC1+BC2	BC1+BC2
900 m	56.9%	42.7%	53.6%	72.8%	95.0%	59.5%
AI	49.4%	48.8%	35.9%	74.6%	85.3%	53.2%
East	60.8%	3.2%	2.7%	70.0%	80.6%	6%
North	21.1%	34.0%	1.5%	75.5%	82.1%	36.1%
South	67.1%	2.8%	1%	77.2%	81.9%	3.8%

Figure 4.6: Barotropic, first and second baroclinic modal decomposition for the ψ field at 900 m. The green point indicates the reference location of the SB at 900 m.

Figure 4.7: Modal recomposition for ψ field at 900 m by: barotropic and the first baroclinic modes (upper panel), barotropic and first two baroclinic modes (central panel), and the first two baroclinic modes (lower panel).

Figure 4.6 presents the BT, BC1 and BC2 components of the mean ψ field at 900m. It is clear that the BT flow is linked to the SPP in the northern half of our study. There is a weak BT cyclonic circulation on the top of the plateau. The BT ψ field retains the flow divergent of the total ψ at 46.8°W-28.7°S, our SB reference site. South of the SB, as expected, we observe an intense flow going polewards, which accounts for the BC. As described by Silveira et al. (2004, 2008); Rocha et al. (2014) the BC south of the SB can be up to 70% barotropic. Our EV value is about 57% in mean terms.

Considering now the BC1 configuration, the central panel of Figure 4.6, we do not observe a bifurcation system. The BC1 ψ field is composed by a very broad and weak jet feeding a narrow northward jet attached to the continental slope. Better formed at the northern half of our domain, the jet coincides with the IWBC site. The BC1 amplitude map presents a large spread of the interior streamlines centered 26.5°S 43°W. This region corresponds to the SPP. The BC2 ψ field (lower panel) displays a broad bifurcating jet. The bifurcation axis is more tilted (45°W relative to the parallels) in this modal amplitude map than in the total ψ map; but retains the SB reference site. On the other hand, there are only very weak currents next to the slope, suggesting that the BC2 probably is more relevant to the open ocean rather than to the western boundary currents. Subsequently inspecting the BC2 field, third panel in figure 4.6, we again detect the SB. Receiving broader and weaker contributions from east, it occurs around our reference point. South of the SB we do not have the BC signal. North of the SB we recognize an equatorward flow, which intensifies in a jet pattern northern of 25°S, adjacent to the 900 m isobath. This location was shown highly Baroclinic at Silveira et al. (2004); Rocha et al. (2014) studies.

The BT and BC1 combined explains 73% of the total streamfunction variance, where the field, first panel at Figure 4.7, remarks the sSEC, the SB, the BC (south of the SB) and the IWBC (North of the SB) well established. Adding also the BC2, second panel at Figure 4.7, the reconstitution comes to explain 95%, intensifying the IWBC and the westward flows in the ocean interior. When combining the BC1 with the BC2, as a proxy to the baroclinic component, third panel of Figure 4.7, we can not identify the SB.

Sectional perspective

We now present and discuss the sectional perspective, beginning with the AI section (Figure 4.8). At this section we have higher EVs for both BT and BC1 modes, each explaining almost 50% of the variance each. Together with the individual mode plots, we also displayed their respective graphs of amplitude to assist the evaluation of the region where the modes have greater influence (higher absolute values of amplitude). Also for the section plots we exhibit the zero contour of the total velocity, as a dashed black line, to indicate the inversion of the BC-IWBC and the SB location. In Figure 4.8, we clearly see that the BT map tracks the BC axis approximately. We have observed at the horizontal panels which agrees with Silveira et al. (2004) and Rocha et al. (2014) findings. Moreover, north of the SB, we do not see the BC flowing poleward. It is replaced by an equatorward circulation which promotes the SB.

Figure 4.8: Modal decomposition for the AI Section. The plots in each panel shows modal Amplitude values along the section. The lower plots correspond to the product of the modal amplitude to the corresponding vertical dynamical mode. The dashed black line point out the zero velocity contour for the total velocity.

The BC1 scenario, in Figure 4.8 show the vertical inversion BC-IWBC and no SB. The mode influence is minimum around the SB region, but the BC1 amplitude value is as intense as the BT one south of the SB site. This also agrees with Silveira et al. (2004) and Rocha et al. (2014): both work highlights the BC1 influence north of the SB. Even though the BC2 presents the SB, it is not dominant in this section. North of 23°S , BC2 reaches its high at absolute values of amplitude. This influence accounts for the high baroclinicity indicated at this region by Silveira

et al. (2004) and Rocha et al. (2014), and the pycnoclinic bifurcation described by Rodrigues et al. (2007).

When combining BT+BC1 we obtain the SB and the BC-IWBC inversion, explaining more than 74% of the variance of the total velocity field. Adding the BC2 together with the first two modes, the EV percentual, increases to more than 85%, once the IWBC achieves its higher velocities north of 23°S. The BC1+BC2 truncation does not present the SB.

Figure 4.9: Velocity modal recombination for the AI Section. BT+BC1, upper left, BT+BC1+BC2, upper right, and BC1+BC2 lower panel. The dashed black line point out the zero velocity contour for the total velocity.

The East section velocity decomposition is majorly barotropic (Figure 4.10, where the BC1 and BC2 modes only explains about 3.5% of the total variance. The BT section, shows a usual bell shaped amplitude, typical of geostrophic jets. The notable intermediate level core observed in the sSEC is due to the BC2 structures, while BC1+BC2 explain the site of the sSEC jet.

Figure 4.10: Modal decomposition for the East Section. The plots in each panel shows modal Amplitude values along the section. The lower plots correspond to the product of the modal amplitude to the corresponding vertical dynamical mode. The dashed black line point out the zero velocity contour for the total velocity.

Figure 4.11: Velocity modal recomposition for the East Section. BT+BC1, upper left, BT+BC1+BC2, upper right, and BC1+BC2 lower panel. The dashed black line point out the zero velocity contour for the total velocity.

The North section is explained mostly by the BC1, 34%, followed by a nearly 21% EV related to the BT mode. In Figure 4.12, we observe a higher amplitude over the shelf and the shelf break, where the BC occupies the whole water column. We identify the IWBC and high BC1 amplitude next to the slope with the BC flowing poleward, and the IWBC equatorward. The vertical pattern agrees well with Rocha et al. (2014) and Biló et al. (2014a) findings nearly this latitude (26°S). The BC2 seems irrelevant to the section velocity structure. The BT and BC1 combined (left panel at Figure 4.13) shapes both BC and IWBC jets, and 82% of the EV is reached if the BC2 mode (right panel Figure 4.13).

Figure 4.12: Modal decomposition for the North Section. The plots in each panel shows modal Amplitude values along the section. The lower plots correspond to the product of the modal amplitude to the corresponding vertical dynamical mode. The dashed black line point out the zero velocity contour for the total velocity.

Figure 4.13: Velocity modal recomposition for the North Section. BT+BC1, upper left, BT+BC1+BC2, upper right, and BC1+BC2 lower panel. The dashed black line point out the zero velocity contour for the total velocity.

Finally, at the South section (Figure 4.12), the BT component rules with 67% significance versus less than 3% for the BC1 and BC2. In this section the BC flows southward from surface to the ocean floor. The BT amplitude, in figure 4.14, shows its highest over of the Tramandaí Rise. The BC1 and BC2, in Figure 4.14, shapes the BC jet and enable the DWBC formation. The figure 4.15 presents the shaping of the BC closest to the coast and the DWBC more offshore, although both being along the bathymetry. This result strongly agrees with Rocha et al. (2014) description of a highly barotropic, (70% of EV) thickened BC.

Figure 4.14: Modal decomposition for the South Section. The plots in each panel shows modal Amplitude values along the section. The lower plots correspond to the product of the modal amplitude to the corresponding vertical dynamical mode. The dashed black line point out the zero velocity contour for the total velocity.

Figure 4.15: Velocity modal recomposition for the South Section. BT+BC1, upper left, BT+BC1+BC2, upper right, and BC1+BC2 lower panel. The dashed black line point out the zero velocity contour for the total velocity.

Chapter 5

Summary and Conclusions

From the analyzes we developed in this study we could verify ours proposed hypotheses, which were:

- The sSEC jet splits into two branches, where the southern portion carries two times the volume transport of the northern one, in terms of annual and monthly means.
- The sSEC jet is mostly barotropic and the BC has a increasingly barotropic influence as it goes south, from 22.5°S down to 33°S .

In average, for the intermediate layers, the sSEC splits almost equally into the SB branches, and it varies through the year. During May and June (end of austral Autumn and beginning of the winter) the northern branch volume transport is higher, representing around 68% of the sum. During the October and November months, the split reaches its most discrepancy, when the southern branch is responsible for more than 74% (more than $2/3$) of the sum. In the other months, despite the transport of the southern branch is higher, it corresponds to 50-60% of the sum. Thus, we reject the first hypothesis which states that the SB split remains constant with the $2/3$ flowing poleward.

In addition, we surprisingly observed that despite the variations on the split proportion of the SB volume transport, its location is virtually constant through the monthly averages. As presented by the previous studies, the SB vary its location with the depth, occurring more to the south on deeper layers. We observed that it occurs mainly from 500 to 1200 m, being virtually constant below that depths.

We could qualitatively evaluate how coincident are the streamlines at the SB level (900 m) to the bathymetric contours. The sSEC jet is channeled by the Rio Grande Rise elevation and the SPP around 29°S . This contouring pattern in the plateau is even more evident on the BT mode recomposition, revealing the topographic-driven nature of the SB.

We could demonstrate that the sSEC is essentially Barotropic, and its first and second baroclinic modes have less than 3.5% of explained variance. The BT recomposition shows a Gaussian-like amplitude, which gives the jet aspect of the impinging sSEC.

As the occurrence of the SB showed almost null variations in its location, we argue that this fact is related to the influence of the bathymetry. Thus, the São Paulo Plateau and the Rio

Grande Rise enclose the sSEC to bifurcates next to the continental margin at a specific latitude. We consider that this behavior is due to the high barotropicity of the sSEC.

The BT component predominance changes with the latitude, being stronger mostly on the southern branch. The Barotropic regime reveals 67.12% of explained variance for the S section. This value decreases drastically to 21.11% for the N transect. On the other hand, the baroclinicity increases equatorward, changing from 2.84 (N section) up to 34% (S section) of explained variance. Hence, we confirm the second hypothesis which affirm the increase on the barotropicity southward.

Some secular, but also meaningfully, aspect that is interesting to point out here is the presence of propagating mesoscale waves coming from the east as described by Majumder et al. (2018). The influence in the SB caused by the presence of this mesoscale activity coming from east was not investigated yet. Nevertheless, we believe that it could significantly influence the SB intraseasonal variability. This topic should be further investigated and is a potential theme for future works.

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