Natália Tasso Signorelli

Indirect investigations of the Atlantic Meridional Overturning changes in the South Atlantic Ocean in numerical models for the 20th century

Master thesis submitted to *Instituto Oceanográfico*
of *Universidade de São Paulo*, in partial fulfillment
of the requirements for the degree of Master of Sci-
ence in Oceanography, with emphasis in Physical
Oceanography.

Advisor:

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Instituto Oceanográfico

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“O que sabemos é uma gota; o que ignoramos é um oceano.”

Isaac Newton (1642–1727)

À minha mãe e meu pai, pelo amor incondicional, pelo exemplo de vida e família.

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List of abbreviations

AABW	Antarctic Bottom Water
AAIW	Antarctic Intermediate Water
AMO	Atlantic Meridional Oscillation
AMOC	Atlantic Meridional Overturning Circulation
BC	Brazil Current
CCSM	Community Climate System Model
CORE	Coordinated Ocean-ice Reference Experiments
cSEC	central band of the South Equatorial Current
DWBC	Deep Western Boundary Current
ENSO	El Niño Southern Oscillation
GFDL	Geophysical Fluid Dynamics Laboratory
IPCC	Intergovernmental Panel on Climate Change
IPO	Interdecadal Pacific Oscillation
ITCZ	Intertropical Convergence Zone
IWBC	Intermediate Western Boundary Current
ISACW	light South Atlantic Central Water
NADW	North Atlantic Deep Water
NAO	North Atlantic Oscillation
NBC	North Brazil Current
NBUC	North Brazil Undercurrent
NEB	North East Brazilian
NH	Northern Hemisphere

NOAA	National Oceanic and Atmospheric Administration
OCCAM	Ocean Circulation and Climate Advanced Modelling project
OGCM	Ocean General Circulation Models
OMP	Optimum Multiparameter
POP	Ocean Parallel Program
RMS	Root-mean-square
SA	South Atlantic
SAC	South Atlantic Current
SACW	South Atlantic Central Water
SACZ	South Atlantic Convergence Zone
SAMW	Subantarctic Mode Water
SASM	South America summer monsoon
SASTMW	South Atlantic Subtropical Mode Water
SEC	South Equatorial Current
SECC	South Equatorial Countercurrent
SH	South Hemisphere
SODA	Simple Ocean Data Assimilation
SST	Sea surface temperature
SvIT	Sverdrup interior transport
SWT	Source Water Type
TW	Tropical Water
UCDW	Upper Circumpolar Deep Water
WBC	Western Boundary Currents

Resumo

O Oceano Atlântico tem papel importante na variabilidade da AMOC, uma vez que inclui dois dos seus principais fluxos de retorno: a NBUC e a IWBC, com origem na bifurcação da SEC e que transportam, principalmente, a SACW e a AAIW. Uma das hipóteses deste trabalho é que, analisando a variabilidade desta bifurcação é possível obter um indicador de mudanças na AMOC. Quatro resultados de modelos oceânicos, com diferentes resoluções espaciais e forçantes foram utilizados. Os resultados mostram que mudanças na bifurcação estão ligadas às mudanças nas correntes e que são, ambas, forçadas pelo vento. Correlações entre a bifurcação da SEC na superfície e a AMOC também foram observadas e os resultados sugerem que isso acontece por meio da NBUC. Outra hipótese é que, num cenário de aquecimento global, uma intensificação do ciclo hidrológico poderia gerar modificações nas massas d'água ligadas à AMOC e contribuir para a sua variabilidade. O núcleo da SACW tem ficado mais raso e essas alterações parecem estar ligadas ao aumento da salinidade projetada nas superfícies neutras. Já o núcleo da AAIW tem diminuído em virtude do aumento da salinidade em 11°S, enquanto o oposto é observado em 27°S.

Palavras-chave: Oceano Atlântico, Célula de Revolvimento Meridional, bifurcação da Corrente Sul Equatorial, Água Central do Atlântico Sul, Água Intermediária Antártica.

Abstract

The South Atlantic has a relevant role on the AMOC variability as it includes two main conduits of its upper-ocean return flow: the NBUC and the IWBC that carry, mainly, the SACW and the AAIW and are originated from the bifurcation of the SEC. One of the hypotheses of this work is that analyzing the bifurcation variability it is possible to get an index of the AMOC changes. Another hypothesis is that in a global warming scenario, changes in the hydrological cycle would drive modifications in the water masses that are part of the AMOC, and thus, contribute to its variability. Four global model results were used, with different forcing and spatial resolution. Results show that changes in the bifurcation are linked to modifications in the currents both caused by variations in the wind stress curl. Good correlations were found between the SEC bifurcation at the surface and the AMOC. The NBUC seems to be the link between them. Shallowing of the SACW core is related to an increase of the salinity on neutral surfaces. The AAIW is occupying less space in the water column due to an increasing of the salinity in the neutral surfaces at 11°S, while the opposite happens at 27°S.

Keywords: South Atlantic, Meridional Overturning Circulation, South Equatorial Current bifurcation, South Atlantic Central Water, Antarctic Intermediate Water

1 Introduction

Changes in the Atlantic Meridional Overturning Circulation (AMOC) have been observed and its weakening is projected for the next century, as suggested by the climate model scenarios compiled for the Fourth Intergovernmental Panel on Climate Change (IPCC) assessment reports (IPCC, 2007; MEEHL et al., 2007).

Few studies have been applied in the South Atlantic, despite its role in AMOC variability. The South Atlantic (SA) connects the North Atlantic to the Indian and Pacific Oceans, exporting about 15 Sv North Atlantic Deep Water. It is through the SA where this southward outflow is compensated by northward inflows of upper and intermediate waters (GARZOLI; MATANO, 2011).

Close to Brazilian shelf, the flow of the North Brazil Undercurrent is responsible for most of the upper-ocean return flow of the MOC (TALLEY, 2003; GANACHAUD, 2003; LUMPKIN; SPEER, 2003). The Intermediate Western Boundary Current is also an important component. Both of these western boundary currents originate from the bifurcation of the South Equatorial Current (SEC) in surface and intermediate depths, respectively (RODRIGUES; ROTHSTEIN; WINBUSH, 2007). Studying the variability of these systems allows one to obtain indices of AMOC variability and changes, as showed by Chang et al. (2008), Rabe, Schott & Köhl (2008), Zhang et al. (2011).

Not only changes in the western boundary currents in the South Atlantic influence the AMOC variability. The water masses carried by these flows, such as the South Atlantic Central Water and the Antarctica Intermediate Water (AAIW) (WIENDERS; ARHAN; MERCIER, 2000), are modified within the South Atlantic basin with potential impact on the AMOC (WEIJER et al., 2002).

Climate change projections show an intensified hydrological cycle (IPCC, 2007; McCARTHY et al., 2012), and one of the impacts expected is the decrease of AAIW salinity, (McCARTNEY, 1977; TALLEY, 1996; SANTOSO; ENGLAND, 2004; GARABATO et al., 2009).

Shifts in the salinity of the AAIW can drive changes in the stability of the AMOC: Manabe & Stouffer (1988) noted that when the NADW is weak it can be associated with a shallow AAIW reverse cell inside the Atlantic, where cool fresh intermediate water replenishes the departure of warm saline thermocline water from the Atlantic. Saenko, Weaver & Gregory (2003), Gregory, Saenko & Weaver (2003), Sijp & England (2006) discuss the active role of this reverse cell in the suppression of NADW formation, and thus, interrupting the flow of AMOC.

However, little is known about how the AAIW varies on intradecadal and interannual timescales in the South Atlantic and there appears to be a disagreement between the studies related to its salinity trends (e.g. Arbic & Owens (2001), Curry, Dickson & Yashayaev (2003), Aoki, Bindoff & Church (2005), McCarthy, McDonagh & King (2011), Schmidtke & Johnson (2012)).

This work analyzes the outputs of four different ocean models, attempting to understand the link between the South Atlantic and the AMOC variations looking at both the SEC bifurcation (and the associated western boundary fluxes) and the relevant water masses.

This Chapter is primarily concerned with describing the Atlantic Meridional Overturning Circulation along with what is known about its variability and its relationship with the South Atlantic Ocean.

1.1 Atlantic Meridional Overturning Circulation

The Atlantic Meridional Overturning Circulation (here after, AMOC) is the large-scale ocean circulation of mass resulting from different forcings. It summarises the system of currents that together comprise a net northward upper ocean transport and southward deep ocean transport (Figure 1). As buoyant upper ocean water flows northward, heat and freshwater are gradually lost to the atmosphere until, on reaching high latitudes, it sinks into the deep ocean and flows southward to complete an overturning cell (GANACHAUD; WUNSCH, 2000; WUNSCH, 2002) (Figure 2).

The AMOC dominates the Atlantic ocean northward transport of heat (VELLINGA; WOOD, 2002; WOOD; VELLINGA; THORPE, 2003) and it is important for local and global climate (KNIGHT et al., 2005; SROKOSZ et al., 2012). For example, at 35°N the ocean is estimated to make up more than 20% of the combined atmosphere-ocean northward heat transport (TRENBERTH; CARON, 2001).

The AMOC is defined as the double integral of the meridional velocity at each latitude, or

$$\psi_{MOC}(y,z) = \int_{x_e,y}^{x_w,y} \int_{-H}^z v(x,y,z) \cdot dz dx, \quad (1.1)$$

where v is the meridional velocity at cartesian location (x,y,z) . Horizontal and vertical integration limits are respectively from the eastern $(x_e(y,z))$ and western $(x_w(y,z))$ boundaries, and from the bottom $-H$ to the depth z . This is known as depth space overturning.

Figure 2c shows an example of mean AMOC (in Sv; $1\text{ Sv} = 10^6 \text{ m}^3 \cdot \text{s}^{-1}$), obtained by Sévellec & Fedorov (2011), in their latitude-depth model of the ocean meridional circulation. Zonally-integrated meridional mass transport, which follows streamlines, is northwards in the upper (low density) ocean, southwards in the deep (dense) ocean and near-vertical at high latitudes. In other words, as described by Sévellec & Fedorov (2011), the circulation is dominated by a clockwise cell that originates in the Northern Hemisphere, penetrates to about 3000m, and is more intense near the surface and along the northern boundary of the basin. In the abyssal ocean (below 3000m) a second, counter-clockwise cell is evident. At the location of the circumpolar channel occurs another intense clockwise cell—the Deacon cell (e.g. Speer, Rintoul & Sloyan (2000)). In low latitudes, there is a signature of the wind-driven shallow subtropical cells in the upper ocean.

Climate models predict a slowing down of the AMOC over the 21st century due to an increase in high latitude buoyancy (IPCC, 2007; MEEHL et al., 2007). The added input of heat and fresh water

Figure 1: Schematic of the AMOC in the South Atlantic Ocean with white arrows indicating surface flows, orange intermediate flows, green and cyan deep flows. Adapted from Lumpkin & Speer (2007), Speich, Blanke & Cai (2007), Garzoli & Matano (2011).

(due to the melting of land and sea ice and the enhanced hydrological cycle) predicted for a warmer climate is expected to inhibit convective overturning and slow the flux of upper ocean water into the deep ocean. Hansen et al. (2001) reported a 20% reduction in the overflow of deep waters through the Greenland-Scotland Ridge (that feeds the densest portion of the MOC cell). Bryden, Longworth & Cunningham (2005) suggest, based on five hydrographic snapshots, that the strength of MOC has decreased by more than 30% over the last five decades. Sévellec & Fedorov (2011) found that in their

Figure 2: Thermal (a) and salinity (b) fields for the on-state of the AMOC (c) produced by Sévellec & Fedorov (2011) using a latitude-depth model of ocean meridional circulation.

idealised model, the AMOC reduces 0.7 Sv for a $10\text{ cm}\cdot\text{y}^{-1}$ increase in freshwater fluxes. Also, the paleoclimate records suggest that past slowdowns of the MOC triggered ice ages with dramatic decreases of the temperature in the western Europe and beyond (SPEICH; GARZOLI; PIOLA, 2010; GARZOLI; MATANO, 2011).

Those evidences motivated a large number of studies, focusing mainly in the North Atlantic (BIAS-TOCH et al., 2008). Although the contribution of the South Atlantic (here after SA) to the AMOC was implicit in the early Meteor observations (WÜST, 1935), there is still uncertainties relative to the ab-

solute magnitude of the SA inter-ocean fluxes (RUIJTER et al., 1999; GARZOLI; BARINGER, 2007). According to Garzoli & Matano (2011), the SA is not just a passive conduit of the remotely formed water masses: its circulation influences significantly their structure through air-sea interactions, mixing, subduction and advection processes.

How the strength and variability of the AMOC are set by the sum of the various currents and processes is not yet fully understood (LOZIER, 2010; KUHNBRODT et al., 2007). There are indications that the AMOC strength is determined by the strength of convective mixing (BÖNING et al., 2006; LATIF et al., 2006), diapycnal mixing (MUNK; WUNSCH, 1998) and Southern Ocean Ekman induced upwelling (TOGGWEILER; SAMUELS, 1995). Causes of AMOC variability remain poorly understood partly because of the complex spatial distribution and variable nature of the currents and processes involved, and also due to the great gaps in our understanding of the SA variability (GARZOLI; MATANO, 2011).

1.1.1 AMOC variability

Despite the shortcomings in the knowledge of AMOC strength and variability, it is known that it fluctuates in several timescales. Recent studies from the RAPID hydrographic observation program (e.g. Cunningham et al. (2007), Kanzow et al. (2007)) have revealed that the AMOC at 26.5°N undergoes large variations on weekly to intra-annual timescales. Ocean models suggest that large AMOC variability exists at flow-frequencies up to multi-decadal in period (BAILEY; RHINES; HÄKKINEN, 2005; BINGHAM et al., 2007; HODSON; SUTTON, 2012) (Figure 3b).

Figure 3: AMOC variability. Time-averaged overturning stream function (a) and northward transport anomaly between 100m and 1000m depth (b) produced by Bingham et al. (2007) a run of the OCCAM described by Coward & Cuevas (2005).

According to Biastoch et al. (2008), short timescale variability is largely a response to Ekman trans-

port and eddy variability. Ekman transport fluctuations modify the ageostrophic meridional transport, which is compensated barotropically at depth (JAYNE; MAROTZKE, 2001). Eddies modify the eastern to western boundary density gradients and therefore the geostrophic currents (THOMMAS, 2012).

Long timescale variability of the AMOC is mostly the result of relatively slow oceanic adjustment mechanisms (HODSON; SUTTON, 2012) that respond to atmospheric forcing. Climate model simulations suggest that these AMOC fluctuations are linked to decadal climate variability, such as NAO, AMO, ENSO, which have important socio-economic impacts including rainfall over Africa, India, North and South America and Atlantic hurricane activity (KNIGHT; FOLLAND; SCAIFE, 2006; ZHANG; DELWORTH, 2006; SMITH et al., 2010). A brief description of the relevant modes of the variability follows:

1.1.1.1 Atlantic Multidecadal Oscillation

The Atlantic Multidecadal Oscillation (AMO) is a near-global scale mode of observed multidecadal climate variability with alternating warm and cool phases over large parts of the Northern Hemisphere (KNIGHT; FOLLAND; SCAIFE, 2006). It is an ongoing series of long-duration changes in the sea surface temperature of the North Atlantic Ocean, with a period of around 65 years (DELWORTH; MANN, 2000; CHIESSI et al., 2009) (Figure 4).

Figure 4: AMO Index from the National Oceanic and Atmospheric Administration (NOAA), between 1948 and 2007.

Many examples of regional multidecadal climate variability in the 20th century have been related to AMO, such as variations in Atlantic sea surface temperatures (SSTs), particularly North East Brazilian (NEB) (FOLLAND et al., 2001) and African Sahel (FOLLAND; PARKER; PALMER, 1986; ROWELL et al., 1995; ROWELL, 2003), rainfall and Atlantic hurricanes (GOLDENBERG et al., 2001).

Also, AMO index is used as a proxy for the strength of the AMOC (ANDRONOVA; SCHLESINGER, 2000; KNIGHT et al., 2005; STOUFFER et al., 2006; CHIESSI et al., 2009; MEDHAUG; FUREVIK, 2011).

While anthropogenic factors appear to have become dominant in the late 20th, the ostensibly natural temperature swings of the AMO have alternately disguised and accentuated the secular trend (EN-

FIELD; MESTAS-NUFIEZ; TRIMBLE, 2001).

1.1.1.2 North Atlantic Oscillation

The North Atlantic Oscillation is a large-scale out of phase relationship between the North Atlantic regions of subtropical high surface pressure (centered near the Azores) and subpolar-low surface pressure (extending south and east of Greenland) (LAMB; PEPPLER, 1987). It is important for setting the strength and position of the westerly winds (HURRELL, 1995).

Figure 5: NAO Index from the National Oceanic and Atmospheric Administration (NOAA), between 1948 and 2007.

Eden & Willebrand (2001) claim that the strength of the MOC is largely determined by the strength of the NAO. They relate the changing strength of the MOC to a deepening and then shoaling of the mixed layer in the Labrador Sea, resulting from the increasing and then decreasing strength of the NAO. Böning et al. (2006), Huang et al. (2012) also suggest that the decadal variability of the subpolar gyre and AMOC near 48°N is largely driven by changes in both heat flux and wind stress associated with this index, mainly variations by about 5 years.

Besides, surface temperatures over the Northern Hemisphere (NH) are likely to be warmer now than at any other time over the past millennium. The rate of warming has been specially high in the past 40 years or so. A substantial fraction of this most recent warming is accounted for a remarkable upward trend in the NAO (KERR, 2000; HURRELL; KUSHNIR; VISBECK, 2011) (Figure 5) and can be associated with intensified westerly wind and surface heat flux anomalies (HOERLING; HURRELL; XU, 2001).

1.2 Area of study

1.2.1 The South Atlantic system of currents

As described above, the South Atlantic (SA) is important for climate system (LEVITUS, 1984) as it transports large amounts of heat due to the presence of the larger meridional branch of thermohaline

circulation (GORDON; GREENGROVE, 1986). Its large-scale circulation is given by the South Atlantic Subtropical Gyre. At the surface, it consists of Benguela, South Atlantic, South Equatorial and Brazil Currents (STRAMMA; ENGLAND, 1999a). Figure 6 shows schematic pathways of the main SA current system.

As described by Stramma & Schott (1999), Stramma et al. (2005), Rodrigues, Rothstein & Winbush (2007), near 14°S, the SEC reaches Brazil shelf region and bifurcates into the NBUC to the north and the BC, to the south (Figure 6).

Figure 6: Superficial South Atlantic Circulation diagram, adapted from Stramma & England (1999a). MC: Malvinas Current; NBC: North Brazil Current; SESC: South branch of South Equatorial Current; AC: Agulhas Current.

The northward limb feeds into the NBUC (STRAMMA; SCHOTT, 1999; SCHOTT et al., 2005) and—as a western boundary current—carries warm waters of South Atlantic origin across the Equator and into the NH. It also supplies the eastward flow of the South Equatorial Countercurrent (SECC) which partially recirculates into the central band of the SEC (cSEC). In addition, in the 50–300m depth range, the NBUC seems to play an important role in the Atlantic Equatorial Gyre (SCHOTT et al., 2005).

At intermediate depths, the Subtropical Gyre differs substantially, as shown in Figure 7. According to Boebel et al. (1999), fresh waters are injected at the intermediate South Atlantic circulation by vigorous eddy motion at the Brazil-Malvinas region (BOEBEL et al., 1999), and subsequently flow eastward carried by the South Atlantic Current (SAC) (PETERSON; STRAMMA, 1990).

At the eastern boundary, there is some input of Pacific intermediate waters (STRAMMA; ENGLAND, 1999a; SCHMID; SIEDLER; ZENK, 2000). Part of this flows to the northwest, next to 30°S

Figure 7: Intermediate depths South Atlantic circulation diagram, adapted from Boebel et al. (1999) and Stramma & England (1999a). MC: Malvinas Current; IWBC: Intermediate Western Boundary Current.

(BOEBEL et al., 1999). At the Santos Bifurcation, $\sim 28^\circ\text{S}$, the South Equatorial Current bifurcates into two fluxes: Intermediate Western Boundary Current (IWBC, which flows to north), (SILVEIRA; CALADO; CASTRO, 2004) and another one to south (BOEBEL et al., 1999).

The IWBC, as part of the AMOC, is one of the ways which the Antarctic Intermediate Water (AAIW, better described in Section 1.2.2) enters into the tropical waters regime (SCHMITZ, 1995; BOEBEL et al., 1999; SILVEIRA; CALADO; CASTRO, 2004).

Under the AAIW, about 20Sv of NADW the Deep Western Boundary Current (DWBC) transports North Atlantic Deep Water (NADW) across the Equator to the Southern Hemisphere (SH) (GOURIOU et al., 2001; SCHOTT et al., 2002; DENGLE et al., 2004). The DWBC flows southward close to the east coast of the South America (STRAMMA; ENGLAND, 1999a). Its flow across the Equator is of fundamental interest because it is a major component of the export of cold water from the North Atlantic to the rest of the planet (McCARTNEY, 1993) (Figure 8).

Bellow the NADW, one finds the Antarctic Bottom Water (AABW). Between 35°S and 40°S it bifurcates, with one branch flowing through the Vema Channel into the Brazil Basin and the other flowing east then south (COLES et al., 1996; STRAMMA; ENGLAND, 1999a) (Figure 8).

Figure 8: Deep and bottom South Atlantic Ocean circulation diagram, adapted from Stramma & England (1999a). White lines indicate the NADW preferential flux at 2000m. In yellow, AABW preferential flow, below NADW. DWBC: Deep Western Boundary Current.

1.2.1.1 South Equatorial Current bifurcation

The SEC forms the northern part of the SA Subtropical Gyre, carrying subtropical water from the Benguela Current region toward the Brazil shelf around 14°S, where it bifurcates. A small part of the water flows northward to form the BC, whereas the bulk of the flow contributes for the NBUC and the SECC (STRAMMA; SCHOTT, 1999; WIENDERS; ARHAN; MERCIER, 2000; STRAMMA et al., 2005; RODRIGUES; ROTHSTEIN; WINBUSH, 2007).

The bifurcation of the SEC close to Brazilian coast can have important implications on climate variability. According to Talley (2003), Ganachaud (2003), Lumpkin & Speer (2003), the SEC/NBUC system is one of the main routes for upper-ocean return flow of the AMOC. Besides, according to Malanotte-Rizzoli et al. (2000), Zhang, McPhaden & Johns (2003) it is also related to the subtropical-tropical mass exchange, the Subtropical Cells (STC). Observations (WILSON; JOHNS; MOLINARI, 1994; SCHOTT; STRAMMA; FISCHER, 1995; ZHANG; MCPHADEN; JOHNS, 2003; GOES et al., 2005) and modelling studies (HARPER, 2000; MALANOTTE-RIZZOLI et al., 2000; INUI et al., 2002; LAZAR et al., 2002; HAZELEGER; VRIES; FRIOCOURT, 2003; LAZAR; SOLOMON, 2006) show that in the Atlantic Ocean, most of the water encountered in the equatorial thermocline comes from the South Atlantic through the low-latitude WBC (NBUC/NBC). Moreover, the variability in the bifurcation latitude and in the transport of the currents can impact other remote regions downstream of the NBUC and BC, for instance the NBC rings (JOHNS; ZANTOPP; GONI, 2003) and the

Brazil-Malvinas Confluence (VIVIER; PROVOST, 1999) regions, respectively.

The Sverdrup theory predicts that the latitude of the SEC bifurcation at the coast of South America should be 14°S , which coincides with the northern edge of the South Atlantic Subtropical Gyre. In fact, there is a southward deepening of the SEC bifurcation along the shoreline (Figure 9). Previous observational studies suggest that the SEC bifurcation occurs at 14° – 16°S in the surface layer (100m), at 20° – 24°S in the thermocline (400–500m), and at 26° – 28°S in the intermediate layer (600–1200m) (STRAMMA; ENGLAND, 1999a; BOEBEL et al., 1999; WIENDERS; ARHAN; MERCIER, 2000; RODRIGUES; ROTHSTEIN; WINBUSH, 2007). From numerical model results, the bifurcation point in the near-surface layer at the western boundary of the South Atlantic occurs at 18°S in the simulations by Harper (2000) and at 17°S in those by Malanotte-Rizzoli et al. (2000). Silva et al. (2009), using a regional model found that the bifurcation occurs between 13°S – 24°S in the first 500m.

Figure 9: Annual mean geostrophic meridional velocity observed by Rodrigues, Rothstein & Winbush (2007). The velocities ($\text{m}\cdot\text{s}^{-1}$) were averaged within a 2° longitude band off the South American coast. Positive (negative) values indicate northward (southward) flow associated with the NBC (BC), and the contour of zero velocity represents the bifurcation of the SEC.

1.2.1.1.1 South Equatorial Current bifurcation variability

Besides its poleward depth increase along the Brazilian coastline, there is a change of the SEC bifurcation position over time. Earlier studies were focused only in seasonal time scales. Rodrigues, Rothstein & Winbush (2007), Silva et al. (2009) describe that, in the near-surface layer, the bifurcation has its northernmost (southernmost) position during the austral winter (summer). At intermediate depths, though, a semi-annual variability is observed (Figure 10).

Figure 10: Time-depth plot of the SEC bifurcation latitude ($^{\circ}\text{S}$), from Rodrigues, Rothstein & Winbush (2007) determined by the location of zero meridional velocity averaged over a 2° longitude band off the South American coast for each month of the year (observations). Contours at every 1° .

According to Rodrigues, Rothstein & Winbush (2007), changes in the bifurcation position are linked to the variability of the wind, specially due to the seasonal meridional movement of the Atlantic ITCZ complex. As a consequence, the bifurcation region is exposed to negative wind stress curl and associated weaker Ekman pumping from February to August and to positive wind stress curl and stronger Ekman suction from September to January. Also, local positive (negative) wind stress curl produces an anomalous anticyclonic (cyclonic) circulation, which southward (northward) component near the western boundary causes the SEC bifurcation to occur at lower (higher) latitude during the austral spring/summer (winter) months.

1.2.1.1.2 South Equatorial Current bifurcation and boundary currents

Seasonal changes in the SEC bifurcation leads to shifts in the boundary currents transports (RODRIGUES; ROTHSTEIN; WINBUSH, 2007). As the SEC bifurcates in its southernmost (northernmost) position in winter (summer) time, the NBUC transport increases (decreases, Figure 11) and the BC transport decreases (increases) (SCHOTT et al., 2005; RODRIGUES; ROTHSTEIN; WINBUSH, 2007; SILVA et al., 2009).

This seasonal cycle of the NBUC transport is very sensitive to latitude: at 10°S the maximum transport is achieved during the austral fall/winter months (April–July), whereas at 6°S the maximum transport happens during summer. The difference between the NBUC transports along 6°S and 10°S reaches its minimum of 3Sv in June–July when the SEC weakens and the SEC bifurcation is in its southern-

Figure 11: Seasonal variability of the latitude position, averaged in the top 200m (Figure 9) versus the NBUC transport at 10°S. Adapted from Rodrigues, Rothstein & Winbush (2007).

most position. Conversely, this transport difference reaches its maximum of 8Sv in October–November when the cSEC intensifies and the SEC bifurcation is in its northernmost position (RODRIGUES; ROTHSTEIN; WINBUSH, 2007).

1.2.1.2 The North Brazil Undercurrent

The North Brazil Undercurrent is a robust boundary current flowing along the Brazilian continental slope (SILVEIRA; MIRANDA; BROWN, 1994). This subsurface flow, proposed by Schott & Böning (1991), was first described (as NBUC) by Stramma, Fischer & Reppin (1995) using velocity data. It is a strong western boundary current, with a nearshore core position, approximately 50km from Brazilian coast, reaching down about 900m in depth, with maximum speed of about $0.65\text{m}\cdot\text{s}^{-1}$ at 180 to 250m on average (SCHOTT et al., 2005; VELEDA et al., 2012).

As mentioned before, the NBUC originates south of 10°S from the SEC bifurcation (e.g. Stramma & England (1999a), Boebel et al. (1999), Harper (2000), Wienders, Arhan & Mercier (2000), Schott et al. (2005), Rodrigues, Rothstein & Winbush (2007), Silva et al. (2009), Velede et al. (2012)) and the bifurcation position has a direct effect in this current seasonal variability. Previous studies, using mooring data at 11°S show that the NBUC core has a dominant fluctuation with a periodicity of two months (SCHOTT et al., 2005; SCHUCKMANN, 2006).

Longer timescale variability of the NBUC was already described in the literature. Zhang et al. (2011), using historical data, observed a multidecadal change in this current transport, indicating a weakening in the 1970s and 1980s compared to the high transports in the early 1960s and 1990s. Despite of the limited data coverage, these results indicates a link between the NBUC variability and the Labrador Sea Water thickness. These authors also show that the time series of the NBUC is significantly correlated with the AMO index, at zero lag, suggesting a direct connection with the AMOC.

1.2.1.3 The Intermediate Western Boundary Current

The Intermediate Western Boundary current originates at $\sim 28^\circ\text{S}$ where the SEC bifurcates at intermediate layers (BOEBEL et al., 1999; SILVEIRA; CALADO; CASTRO, 2004). Part flows southward and contributes to the BC flux, part flows northward, contributing to the AMOC return flow (SCHMITZ, 1995; McCARTHY et al., 2012).

This boundary current is well defined along relatively smooth isobaths from 27°S to the Victoria-Trindade Rigde at $20^\circ 30'\text{S}$ (LEGEAIS; OLLITRAULT; ARHAN, 2013), usually with velocities greater than $0.3\text{m}\cdot\text{s}^{-1}$ (SILVEIRA; CALADO; CASTRO, 2004). Sharp variations of the continental slope orientation cause a partial water scape from the boundary at 20°S , resulting in a weaker IWBC on average, and strong offshore mesoscale activity. The current is well defined between 15°S and 5°S (LEGEAIS; OLLITRAULT; ARHAN, 2013).

Besides its important role, there is still much to be uncovered about the IBWC.

1.2.2 The South Atlantic water masses structure

The water masses structure of the SAO, as shown in Figure 12, is composed by Tropical Water TW, the South Atlantic Central Water (SACW), Antarctic Intermediate Water, North Atlantic Deep Water (NADW) and the Antarctic Bottom Water (AABW).

Figure 12: T-S diagram, representing the most important South Atlantic water masses. Original data from Simple Ocean Data Assimilation System, version 2.2.4 (SODA, Carton & Giese (2008)) at 25°S and 25°W .

The SACW is a water mass with substantially uniform properties (STRAMMA; ENGLAND, 1999a), with temperatures between 5°C and 20°C , and salinity varying from 34.3 to 36 (TOMCZAK; GODFREY, 1994) (Figure 13).

Figure 13: WOCE potential temperature (a) and salinity (b) sections at 25°W (section A16). Adapted from Schlitzer (2000).

The $27.1\text{kg}\cdot\text{m}^{-3}$ and (in the tropics) and $27.05\text{kg}\cdot\text{m}^{-3}$ (in the subtropics) isopycnals mark the transition between the SACW and the AAIW (STRAMMA; ENGLAND, 1999a). According to Silveira, Calado & Castro (2004), close to the Brazilian coast, this interface occurs between 500–750m.

The AAIW is found mainly between 500–1500m and marked by the vertical salinity minimum (Figures 12 and 13) and densities close to $27.3\text{kg}\cdot\text{m}^{-3}$ (PICKARD; EMERY; TALLEY, 2007).

First described by Wüst (1935), the AAIW flows northward (MÜLLER et al., 1998; CIRANO et al., 2006), mainly via IWBC (as described in Section 1.2.1) and can be observed all the way to 20°N (TALLEY; SPRINTALL, 2005). Along the way, its salinity slowly increases (Figure 14) and its concentration of oxygen decreases (STRAMMA; ENGLAND, 1999a; SCHMID; SIEDLER; ZENK, 2000).

Below the AAIW, the Upper Circumpolar Deep Water (UCDW) takes place. In the tropics, according to Stramma & England (1999b), at about 1000m depth, it is the deepest water mass of the upper

Figure 14: Maps of the salinity minimum produced by the intermediate waters, mainly AAIW. Depth of the minimum (a), salinity at the depth of the minimum (b). Adapted from Tomczak & Godfrey (2003).

ocean with a net northward transport. According to Santoso, England & Hirst (2005), it contributes further to a lighter AAIW.

The NADW is identified by the subsurface salinity maximum (Figures 12 and 13) at depths of $\sim 2000\text{m}$. Its thickness decreases along the way and its flux is strongly influenced by topography (STRAMMA; ENGLAND, 1999a). Thus the NADW flow occurs mainly by the western boundary of the basin until $\sim 45^\circ\text{S}$ (PETERSON; WHITWORTH-III, 1989), where the Antarctic Circumpolar Current (ACC) makes it flows eastward (STRAMMA; ENGLAND, 1999a). The NADW then mixes with the Circumpolar Deep Water (CDW), changing its temperature and salinity characteristics (PROVOST et al., 1995).

Finally, below 3500m there is the AABW (HOGG; OWENS, 1999) (Figure 13). With temperatures between -0.4°C and 0.2°C and uniform salinity (between 34.66 and 34.68) (COLES et al., 1996), this water mass is formed in the Southern Ocean, specially at the Weddell Sea (CARMACK; FOSTER, 1975). Its northward flux, through the Atlantic Ocean western boundary (Figure 8), is also highly influenced by the bathymetry (STRAMMA; ENGLAND, 1999a).

These water masses (AABW and NADW) are very important for the ocean ventilation (MEREDITH et al., 2000), as they are part of thermohaline circulation (CARMACK; FOSTER, 1975; FAHRBACH et al., 1995; KERR, 2006).

1.2.2.1 South Atlantic Central Water

The South Atlantic Central Water is found south of about 15°N, and shows rather uniform properties throughout its range (STRAMMA; ENGLAND, 1999a). According to McCartney (1982) and Donners, Drijfhout & Hazeleger (2005), the SACW is a combination of light South Atlantic Central Water (ISACW), South Atlantic Subtropical Mode Water (SASTMW), and Subantarctic Mode Water (SAMW). Their density layers are shown in Table 1.

Table 1: Density layers of water masses that contribute to SACW formation, from Donners, Drijfhout & Hazeleger (2005).

	Density layer ($\text{kg}\cdot\text{m}^{-3}$)
ISACW	25.5–26.2
SASTMW	26.2–26.6
SAMW	26.6–27.0

The SAMW is formed at the Brazil–Malvinas Confluence (GORDON, 1981). It exhibits a basin-wide influence at mid-pycnocline, detectable at low latitudes. In the South Atlantic, the SAMW is weaker in strength compared to other basins and its influence is laterally restricted to only the western sector of the subtropical gyre (McCARTNEY, 1982). It is responsible for a high-salinity variety of SACW (STRAMMA; ENGLAND, 1999a).

The SASTMW consists of a light variety of subtropical mode waters (PROVOST et al., 1999), formed mainly in the Brazil Current overshoot, and 13°C mode water, presumably formed by mixing in Agulhas rings (TOMCZAK; GODFREY, 1994). Then it is transported by the SAC (STRAMMA; PETERSON, 1990) to the eastern boundary where it is mixed to the Indian Central Waters brought by the Agulhas Current (STRAMMA; ENGLAND, 1999a), which forms part of the ISACW (TOMCZAK; GODFREY, 1994). So, the Benguela Current and the SAC carry it to the Brazilian shelf, via SEC, where it is finally transported by the BC. Also, the SACW is carried towards the equator with the NBUC and the NBC (STRAMMA; SCHOTT, 1999).

Previous studies related to the SACW were concerned mainly with its formation processes and properties, and information about its variability and changes is scarce. Durack & Wijffels (2010), using observations, showed that, the salinity on the $26.75\text{kg}\cdot\text{m}^{-3}$ isopycnal, decreased 0.2 between 1950 and 2008 close to the Brazilian coast. Results of Arbic & Owens (2001) show a warming of the water column in the depth of the SACW.

1.2.2.2 Antarctic Intermediate Water

The Antarctic Intermediate Water is a very prominent water mass that lies above the deep water and spreads below the subtropical thermocline in the SH, marked by the vertical salinity minimum (e.g. Talley (1996)). According to McCartney (1977) it is formed primarily in the southeast Pacific Ocean

just equatorward of the ACC. Pickard, Emery & Talley (2007) also suggested it is formed north of the Subantarctic Front, near the Drake Passage. Its properties are consequence of three different physical processes: convection, subduction (TOMCZAK, 1999) and subsurface mixing (SØRENSEN; RIBBE; SHAFFER, 2001).

Due to its importance for the global climate, there have been many investigations, both regional and global, into intermediate-depth potential temperature and salinity changes on isobars and isopycnals in the southern oceans (BINDOFF; CHURCH, 1992; JOHNSON; ORSI, 1997; WONG; BINDOFF; CHURCH, 2001; AOKI; BINDOFF; CHURCH, 2005; BÖNING et al., 2008; GOES et al., 2008; DURACK; WIJFFELS, 2010; HELM; BINDOFF; CHURCH, 2010; SCHMIDTKO; JOHNSON, 2012). These studies uniformly find a warming trend.

Freshening of AAIW has been observed in the Pacific (WONG; BINDOFF; CHURCH,) and the Indian Oceans (BINDOFF; MCDOUGALL, 2000; BRYDEN; MCDONAGH; KING, 2003; MCDONAGH; KING, 2005). However, in the South Atlantic the situation is not as clear. Curry, Dickson & Yashayaev (2003) found that AAIW salinity had decreased on isopycnals from the 1950s–1960s to the 1980s–1990s. Arbic & Owens (2001), McCarthy, McDonagh & King (2011) noted an increase in AAIW salinity at 24°S from 1958 to 1983. These differences have been linked to changes in the hydrological cycle and in the circulation.

As mentioned by McCarthy et al. (2012), little is known about how AAIW varies on intradecadal and interannual timescales, and these changes in AAIW salinity are still under discussion. Also, according to the authors, higher frequency variability could well be the cause for disagreement between studies.

1.2.3 The South Atlantic and the Meridional Overturning Circulation

The SAO is an important part of AMOC. Its mean meridional circulation structure involves a deep southward flow of cold and salty North Atlantic Deep Water (NADW) along the eastern coast of South America. This water mass is cooled primarily by evaporation before it sinks to the deep ocean (GORDON, 1986). This southward flux is compensated by the net northward flow through the SA and across the equator of surface, intermediate and bottom water layers (BROECKER, 1991; SCHMITZ, 1995; SPEICH et al., 2002; GARZOLI; MATANO, 2011).

The SAO is the only basin extending to high latitudes in which the heat transport is equatorward (GARZOLI; MATANO, 2011). Although this distinct heat flux was recognized by the middle of the last century (MODEL, 1950), the sources for the upper waters are still under discussion. A portion is produced locally (STRAMMA; ENGLAND, 1999a), but most of them are thought to originate in the Pacific and Indian Oceans (GORDON; GREENGROVE, 1986; RINTOUL, 1991; GORDON et al., 1992; STRAMMA; ENGLAND, 1999a; SCHMID; SIEDLER; ZENK, 2000; GARZOLI et al., 2013). In this sense, the SA provides the gateway by which the AMOC connects with the rest of the globe.

The South Atlantic Subtropical Gyre also plays an important role in the AMOC return flow, through

Figure 15: Schematic of the AMOC in the South Atlantic Ocean with white arrows indicating surface flows, linked to NBUC and orange ones the intermediate flows, linked to IWBC. Adapted from Lumpkin & Speer (2007), Speich, Blanke & Cai (2007), Garzoli & Matano (2011).

the western boundary currents, such as North Brazil Undercurrent (NBUC) and the Intermediate Western Boundary Current (IWBC) (SCHMITZ, 1995; BOEBEL et al., 1999; TALLEY, 2003; GANA-CHAUD, 2003; LUMPKIN; SPEER, 2003; SILVEIRA; CALADO; CASTRO, 2004) (Figure 15). Vellinga & Wu (2004) showed, using a coupled General Circulation Model that the NBC is stronger during decades of stronger AMOC composites than during weaker composites. Rabe, Schott & Köhl (2008), comparing the AMOC at 10°S with the NBUC in the 50 year ECCO data assimilation system, observed that in their model, at time scales longer than 4 years, the NBUC variability explains 63% of AMOC variance (Figure 16). Chang et al. (2008), in an idealised North Atlantic “water hosing” coupled model experiment, showed that a dramatic change of NBUC transport is concurrent with a significant weakening of the AMOC, both following a significantly reduced deep water formation at high latitudes. Zhang et al. (2011), in a modelling study, reveal that, in periods longer than 10 years, the correlation between the NBUC and AMOC transport variability is 80% at zero lag.

Paleoclimate studies also show a strong relationship between the SAO and the AMOC. Model simulation of deglacial events suggests that during a meltwater event a weakened thermohaline overturning causes a reduced northward advection of warm, saline surface water, which consequently leads to a warming of the upper layers in the tropical ocean (MANABE; STOUFFER, 1997). Correspondingly,

Figure 16: MOC (solid black), NBUC (gray), and Sverdrup interior (SvTI, dashed) transport anomalies at 10°S. Time series with variabilities longer than 4 years. Adapted from Rabe, Schott & Köhl (2008).

during an intensification of the MOC, net heat loss and cooling of the Southern Hemisphere is observed, which is related to a stronger compensatory cross-equatorial northward heat transport (CROWLEY, 1992). Direct oceanographic studies and also coupled GCM results (ARZ; PÄTZOLD; WEFER, 1999) demonstrate the significance of the western South Atlantic boundary currents, such as North Brazil Current (NBC), NBUC and Brazil Current (BC) within this connection.

Also in a paleoclimate study, Chiessi et al. (2009) proposes that this relationship between WBC and AMOC could lead changes in the South American summer monsoon (SASM) and South Atlantic Convergence Zone (SACZ). With a weakening of the AMOC and the NBUC during negative AMO phases, and the consequent increase of the South Atlantic sea surface temperatures (ARZ; PÄTZOLD; WEFER, 1999), there is an enhancement in the SACZ activity and rainfall over the southeastern South America.

Moreover, another mechanism that might link AMO and SASM/SACZ involves the position of the Intertropical Convergence Zone (ITCZ) in the equatorial Atlantic. Modelling results of Zhang & Delworth (2005) show that during periods of weak AMOC, the ITCZ is shifted to the south causing a positive anomaly in moisture transport into the Amazon basin. According to the authors, the equatorial anomalies would enhance convection in the Amazon, causing an intensification of the northwesterly South American low-level jet that transports the Amazon moisture and forms the SACZ (MARENGO et al., 2004) which would increase the precipitation in the southeastern South America.

Changes in the salinity or temperature of the water masses in the SA also have implications in the AMOC variability. Manabe & Stouffer (1988) observed that the stability of the AMOC is linked to a shallower or deeper AAIW reverse cell. Vries & Weber (2005) show, with modelling studies, that it is mainly a result of a salt feedback in the ocean, which determines whether the thermohaline circulation

is monostable or bistable (DRIJFHOUT; WEBER; SWALUW, 2011; GARZOLI et al., 2013).

1.3 Ocean General Circulation Models

An Ocean General Circulation Model (OGCM) is a mathematical representation of the large-scale circulation of a planetary ocean, that responds according to the physical laws of conservation of mass, momentum, energy, and other properties and in which the fluxes from the atmosphere are imposed.

They enable us to study the ocean globally, providing uniform data resolution in space and time, and for this reason they are very useful when observations are scarce, often not synoptic and collected using a range of methodologies. Moreover, climate runs of OGCMs are excellent tools to study the ocean response to the increasing greenhouse gas concentrations, such as those described in the IPCC Fourth Assessment Report (IPCC, 2007). Running an OGCM under prescribed forcing scenarios can indicate which processes and regions are likely to be more affected, providing a starting place for the design of observational programs or mitigation policies.

In dealing with an under-sampled natural fluid system, it is extremely helpful to invoke the simulation capability of supercomputers to improve understanding of basic processes and their interconnectivity, as well as to help interpret sparse observations (SEMTNER, 1995).

One of the first significant modelling effort extended over the decade of the 1960s at the Geophysical Fluid Dynamics Laboratory (GFDL) (BRYAN, 1969). Bryan built an ocean model that was to be used with atmospheric models to study climate. The first application with specified global geometry was done in the early 1970s (COX, 1975). Cox designed a 2° latitude-longitude grid with up to 12 vertical levels at each point. The model was run for hundreds of hours on the fastest available computer but had only reached a few simulated years after many months. By the mid-1980s, progress had been made in simulating more aspects of real-world ocean circulation, much of it resulting from increases in computing power (SEMTNER, 1995). Nowadays, an increasing model fidelity to real world processes is being achieved with enhanced temporal and spatial resolutions. Even having observed such progress, high resolution, eddy resolving OGCMs are still limited to computer power.

Such models depend on parameterisations of many unresolved ocean processes (GRIFFIES, 2004). Despite this, as shown by Delworth & Mann (2000), state of the art climate models are able to reproduce magnitudes and patterns of decadal variability seen in observations. However, as pointed out by Thommas (2012), it is important to be aware of some of their shortcomings:

- Eddies affect the AMOC by modifying the east-west density gradient in the ocean (BIASTOCH et al., 2008), the pathways of currents (GARY et al., 2011) and recirculations Lozier (2010). In this sense, according to (MARSH et al., 2009), models with resolved eddies have been shown to better represent the time-mean and variable structure of the AMOC;

- According to Biastoch et al. (2008), high latitude processes associated with deep water formation require submesoscale resolution. In consequence, low-resolution models can have a too-shallow overturning (SAUNDERS et al., 2008);
- Deep ocean circulation is too weak in almost all ocean models because of the long model spin-up time required for the deep ocean to reach equilibrium (WUNSCH, 2011).

1.4 Motivation

The South Atlantic Ocean plays a unique role in the global energy balance, transporting, as part of the AMOC, heat from the poles to the equator (e.g., Talley (2003)) as upper layer water spreads northward to compensate for the southward export of colder NADW (BROECKER, 1991; SCHMITZ, 1995; SPEICH et al., 2002; GARZOLI; MATANO, 2011; GARZOLI et al., 2013). This northward flux is a complex mixture of waters originating from the Indian and Pacific oceans (GORDON; GREEN-GROVE, 1986; RINTOUL, 1991; GORDON et al., 1992; STRAMMA; ENGLAND, 1999a; SCHMID; SIEDLER; ZENK, 2000), blended together in the South Atlantic and overlain by large-scale gyre circulations.

The mechanisms and pathways of mass and heat transport in the global ocean are due in large part to the wind-driven and buoyancy-driven components of the MOC. Associated with the strong thermohaline (buoyancy driven) circulation in the Atlantic Ocean, heat and salt are exchanged between the hemispheres (GARZOLI et al., 2013). Vries & Weber (2005) showed from modelling studies that the instability of the AMOC results from salt feedback in the ocean. According to the authors, when we intensify the hydrological cycle, the AMOC transports fresher waters. Also, the AMOC weakens, as more salt flows out of the Atlantic via the Brazil Current and less salt flows equatorward inhibiting the NADW production. Besides, “excess” salt flowing out via the Brazil Current recirculates awhile in the Subtropical Gyre thereby increasing the local salinity, causing the feedback.

Wind-driven variations of the AMOC are associated, in the South Atlantic, for example, with changes in the transport of the NBUC/NBC. In a modelling study, Vellinga & Wu (2004) observed that the NBC is stronger during decades of stronger AMOC composites than during weaker composites. Chang et al. (2008), in an idealised North Atlantic “water hosing” coupled model experiment, showed that a dramatic change of NBC transport is concurrent with a significant weakening of the AMOC, both following a significantly reduced deep water formation at high latitudes. Also using a coupled model, Zhang et al. (2011), showed that the correlation between the NBC and the AMOC transport variability (with periods bigger than 10 years) reaches its maximum value of 0.8 at zero lag.

These and other studies suggest that both changes in the Western Boundary Currents (WBC) and in the water masses that they carry affect the AMOC variability. In a future climate scenario in which a weakening of the AMOC is predicted for the 21st century, it is necessary to better understand the associated physical process that impacts on the AMOC variability.

1.5 *Scientific hypothesis*

The variability and the changes in the bifurcation of the South Equatorial Current (SEC), in surface or intermediate depths, are still not well described in literature, specially in longer timescales.

Since it originates the NBUC in surface and the IWBC at intermediate depths, that are part of the upper branch of the thermohaline circulation in the South Atlantic, it is indispensable to understand properly the mechanisms that drive its (seasonal, interannual and multidecadal) variability.

Therefore, the hypothesis is that, by observing and analyzing the SEC bifurcation, it is possible to get an index of AMOC variability and changes. Since the bifurcation is under different regimes in the surface and intermediate depths with distinct impacts in its variability, at least in seasonal timescales (RODRIGUES; ROTHSTEIN; WINBUSH, 2007), both the NBUC and the IWBC systems should contribute to understand the AMOC variances, particularly in timescales longer than a year.

Another hypothesis is that, in a future climate scenario of global warming, changes in the hydrological cycle would drive modifications in the salinity of the water masses that are part of the AMOC, and thus, contribute to its variability.

1.6 *Objectives*

Based on the hypotheses described above, the main goal of this work is to access the variability of the South Equatorial Bifurcation, in interannual and longer timescales, and verify if it can be used as evidence or index of Atlantic Meridional Overturning Cell changes and variability.

Also, this work aims to verify, in interannual and longer timescales, changes in the water masses related to the upper branch of the AMOC: SACW and AAIW.

More specifically, this project aims to:

- analyze models results, compared to the reanalysis, checking:
 - the position of the bifurcation, and their changes over the time;
 - changes in the main currents (transports);
 - correlate them with wind stress and other climate indices, such as North Atlantic Oscillation.
 - the core of these water masses, isolating it using the OMP analysis;
 - changes in the salinity and temperature projected onto the main neutral surfaces;
 - correlate them with changes in the core of the water masses.

1.7 *Master thesis outline*

In this master thesis, many models simulations are used to investigate changes in the AMOC variability. Chapter 2 describes the model outputs and the analysis used in this investigation. Chapter 3

brings the results, split in two main sections: in Section 3.1, changes in the WBC are investigated by looking to the interannual variability of the South Equatorial Current Bifurcation (SEC); Section 3.2 addresses changes in the properties of the water masses carried by these currents. Chapter 4 concludes the thesis.

2 Data and methods

To address the variability of South Equatorial Current bifurcation in the South Atlantic Ocean, the results of four different general ocean models are analyzed. These are described in Section 2.1. Then, Section 2.2 presents the methods.

2.1 Data

2.1.1 Simple Ocean Data Assimilation

The Simple Ocean Data Assimilation, version 2.2.4 (SODA) is a reanalysis of ocean climate variability. It begins with a state forecast produced by an ocean general circulation model based on Parallel Ocean Program numerics (POP) (SMITH; DUKOWICZ; MALONE, 1992) with the average $0.25^\circ \times 0.4^\circ$ horizontal resolution and 40 vertical levels with 10m spacing near the surface. This ocean model is forced by 20Crv2 winds (COMPO et al., 2011) and is continuously corrected by contemporaneous observations, with corrections estimated every 10 days. The observation set includes the historical archive of hydrographic profiles supplemented by ship intake measurements, moored hydrographic observations, and remotely sensed SST and sea level. See Carton et al. (2000a), Carton et al. (2000b), Carton & Giese (2006) and Carton & Giese (2008) for more details.

This dataset contains monthly global information such as temperature, salinity, wind stress, sea surface height. The output is monthly and mapped onto a uniform $0.5^\circ \times 0.5^\circ \times 40$ -level grid. The time series is from 1871 to 2008. In this work only data between 1948 and 2007 are used.

2.1.2 The ocean component of Community Climate System Model

The Community Climate System Model (CCSM), version 4.0, is a global model developed by the National Center for Atmospheric Research (NCAR). As it is a complex coupled model (more details in Vertenstein et al. (2010)), the results analyzed in this work are a result of a simulation in which the ocean component was coupled to a dynamic and hydrodynamic sea-ice model¹ (CLIVAR, 2010), forced by CORE (Section 2.1.4) dataset.

The ocean component, POP, solves the primitive equations in general orthogonal coordinates in the horizontal with hydrostatic and Boussinesq approximations. For further details, see Danabasoglu et al. (2012).

¹ CICE – The Los Alamos sea ice model – <http://oceans11.lanl.gov/trac/CICE>

This run was initialised from resting state, with temperature and salinity data of Levitus set, atmospheric data from CORE-II and river discharges based on estimates of Dai & Trenberth (2002), (CLIVAR, 2010). The spin-up time was 500 years (KERR; WAINER; MATA, 2009). More details are available in Collins et al. (2006).

This dataset has monthly data with 1° of spacial resolution and 60 vertical levels, from 1948 to 2007.

2.1.3 *Nucleus for European Modelling of the Ocean*

The Nucleus for European Modelling of the Ocean (NEMO) is a primitive equation model adapted to regional and global ocean circulation problems (MADEC, 2008). In DRAKKAR runs, the ocean component OPA (*Ocean Parallélisé*) (DELECLUSE; IMBARD; LEVY, 1998) is coupled to a multi-layer sea-ice model LIM (Louvain-la-Neuve Sea Ice Model) (FICHEFET; MAQUEDA, 1999; VANCOPPENOLLE et al., 2009), using the quasi-isotropic global ORCA grid (for a better description, see Bernard et al. (2006)).

2.1.3.1 ORCA05

One of the DRAKKAR² products is the ORCA05 dataset. It is a global ocean model coupled to a sea-ice model with a nominal grid resolution of $0.5^\circ \times 0.5^\circ$ that reasonably resolves western boundary current structures and captures the different processes of deep-water formation due to the thermohaline and wind forcing, but that does not resolve mesoscale processes. In the vertical 46 levels (with 10 levels in the upper 100m and 250m resolution at depth) are used, whereby the bottom cells are allowed to be partially filled (BIASTOCH et al., 2008; BIASTOCH; BÖNING; LUTJEHARMS, 2008).

This run was initialised using temperature and salinity climatological fields, and it was integrated from 1948 to 2007. The forcings, as described in Brodeau et al. (2010), are a combination of ERA-40 reanalysis, satellite and CORE (LARGE; YEAGER, 2009) data.

2.1.3.2 ORCA12

ORCA12 is a high-resolution simulation produced by DRAKKAR. It is similar to ORCA05, but it has $1^\circ/12$ of horizontal resolution (between 50°S and 50°N) and 46 vertical levels. The first baroclinic Rossby deformation radius is larger than 20km in this latitude band (CHELTON et al., 1998), hence this simulation can be considered as fully eddy-resolving (DESHAYES et al., 2012).

The simulation starts in 1989 and ends in 2007. The output data were monthly averaged. The simulation starts from another simulation (C-HR), and is forced by ERA-Interim (DESHAYES et al., 2012) with corrections of the large-scale shortwave flux bias toward GEWEX observations (DEE et al., 2011) and with corrections of precipitations toward GPCP data (BOLVIN; GU, 2009). This first simulation

² Drakkar: Multi-scale Ocean modelling project – <http://www.ifremer.fr/lpo/La-recherche/Projets-encours/DRAKKAR-Ocean>

(C-HR) starts from rest and climatological temperature and salinity and is forced by CORE-II data (Section 2.1.4). More details are available in Deshayes et al. (2012).

2.1.4 Coordinated Ocean-ice Reference Experiments

The Coordinated Ocean-ice Reference Experiments (CORE), version 2, developed by the World Climate Research Program (WCRP) as part of the Climate Variability and Predictability (CLIVAR) project, is a dataset created to be used in ice models as forcing.

This set contains global sea-air fluxes related to the variability of interannual ocean models, such as momentum, heat and fresh water fluxes, from 1948 to 2007. More information can be found in Large & Yeager (2009).

2.1.5 Climate timeseries

AMO³ and NAO⁴ timeseries were obtained from the National Oceanic and Atmospheric Administration website (NOAA).

The AMO timeseries were calculated from the Kaplan SST dataset, that interpolates back in time using EOF projection, optimal interpolation, Kalman filter, forecast, KF analysis, and an optimal smoother (KAPLAN et al., 1998; ENFIELD; MESTAS-NUFIEZ; TRIMBLE, 2001).

The NAO timeseries used here are historical data, since 1950. Indices are standardised by the 1981–2010 climatology. The procedure used to calculate NAO indices is based on the Rotated Principal Component Analysis (RPCA) used by Barnston & Livezey (1987). The RPCA technique is applied to monthly standardised 500mb height anomalies in the analysis region 20°N–90°N between January 1950 and December 2000.

2.2 Methods

Current speed and wind stress data are analyzed to access the variability of the South Equatorial Current (SEC) Bifurcation in the South Atlantic Ocean.

Three main sections were chosen: 11°S, 27°S and 25°W. As described in Section 1.2, close to 11°S, the SEC bifurcates at the surface resulting in the North Brazil Undercurrent (NBUC, flowing northward) and the Brazil Current (BC, flowing southward). At the Brazilian continental slope, in intermediate depths, near 27°S, the SEC bifurcates, originating the Intermediate Western Boundary Current (IBWC, flowing northward) while the western boundary currents flows farther south at ~40–30°S (BOEBEL et al., 1997). The section at 25°W was used to monitor the South Equatorial Current (SEC).

³ <http://www.esrl.noaa.gov/psd/data/timeseries/AMO/>

⁴ <http://www.cpc.ncep.noaa.gov/data/teledoc/nao.shtml>

2.2.1 Position of the bifurcation

To trace the bifurcation at depths we followed the methodology proposed by Rodrigues, Rothstein & Winbush (2007). In their paper, the authors averaged the meridional velocities within a 2° longitude band off the Brazilian coast.

In this work the position of the SEC bifurcation was determined as the zero contour line (Figure 9) in a section constructed from the average of 4° longitude band off the Brazilian coast (Figure 17).

Figure 17: The white shaded area represents the 4° longitude band off the Brazilian coast area where the meridional velocities were averaged to determine the SEC bifurcation position, following Rodrigues, Rothstein & Winbush (2007).

2.2.2 Volume transports

Volume transports were calculated at 11°S, 27°S and 25°W, following Equation 2.1, where dx and dz are the horizontal and the vertical dimensions of the model cell.

$$\tau_{Sv}(x,t) = \iint_{-H}^0 v(x,t) \cdot dz \, dx \quad (2.1)$$

At 11°S, the length of the section is a function of the climatological (average for the whole period) width of NBUC, and only positive (northward) velocities were used in this analysis. So, for the datasets SODA, ORCA05 and ORCA12, the zonal transports were calculated between 37.5°W and 34°W. For CCSM, the section started at 37.5°W and extend until 32°W.

At 27°S, the parameter was the climatological width of Brazil Current. Only negative velocities (southward) were used to calculate the transport of the BC. At intermediate depths, positive (north-

ward) velocities were used to calculate the IWBC volume transport. The sections limits were 48°W and 45°W to SODA, ORCA05 and ORCA12 datasets. For CCSM, they were 48°W and 43.5°W.

At 25°W, the volume transport of the SEC were calculated in two 5° sections centered in 11°S and 27°S. In these sections, only negative (westward) velocities were used.

2.2.2.1 Sverdrup transport

The Sverdrup's Theory of the oceanic circulation (SVERDRUP, 1947) describes the transports in the ocean interior by relating the curl of the wind stress to mass transport within the upper ocean. He assumed that the flow is stationary, that lateral friction and molecular viscosity are small, and that turbulence near the sea surface can be described using an eddy viscosity. He also assumed that the flow is baroclinic and that the wind-driven circulation vanishes at some depth of no motion. In the subtropics Sverdrup theory is believed to hold to a good order of approximation in the interior (east of the western boundary current), where the ocean is in linear vorticity balance and there is a presumed level of no motion.

The following description was adapted from Stewart (2008).

The horizontal components of the momentum equation are:

$$\frac{\partial p}{\partial x} = f\rho v + \frac{\partial}{\partial z} \left(A_z \frac{\partial u}{\partial z} \right) \quad (2.2)$$

$$\frac{\partial p}{\partial y} = -f\rho u + \frac{\partial}{\partial z} \left(A_z \frac{\partial v}{\partial z} \right) \quad (2.3)$$

where:

- p is the pressure (Pa);
- u, v are the horizontal velocities ($\text{m}\cdot\text{s}^{-1}$);
- A_z is the coefficient of eddy viscosity ($\text{kg}\cdot(\text{m}\cdot\text{s})^{-1}$).

Sverdrup integrated these equations from the surface to a depth $-D$ equal to or greater than the depth at which the horizontal pressure gradient becomes zero. He defined:

$$\frac{\partial P}{\partial x} = \int_{-D}^0 \frac{\partial p}{\partial x} \cdot dz \quad M_x = \int_{-D}^0 \rho u \cdot dz \quad (2.4)$$

$$\frac{\partial P}{\partial y} = \int_{-D}^0 \frac{\partial p}{\partial y} \cdot dz \quad M_y = \int_{-D}^0 \rho v \cdot dz \quad (2.5)$$

where M_x, M_y are the mass transports in the wind-driven layer extending down to an assumed depth of no motion.

The horizontal boundary condition at the sea surface is the wind stress, and the boundary at depth $-D$ is zero stress because the currents go to zero:

$$\left(A_z \frac{\partial u}{\partial z}\right)_0 = \tau_x \qquad \left(A_z \frac{\partial u}{\partial z}\right)_{-D} = 0 \quad (2.6)$$

$$\left(A_z \frac{\partial v}{\partial z}\right)_0 = \tau_y \qquad \left(A_z \frac{\partial v}{\partial z}\right)_{-D} = 0 \quad (2.7)$$

where τ_x and τ_y are the components of the wind stress.

Using these definitions and boundary conditions, Equations 2.2 and 2.3 become:

$$\frac{\partial P}{\partial z} = f \cdot M_y + \tau_x \quad (2.8)$$

$$\frac{\partial P}{\partial z} = -f \cdot M_x + \tau_y \quad (2.9)$$

In a similar way, Sverdrup integrated the continuity equation over the same vertical depth, assuming the vertical velocity at the surface and at depth $-D$ are zero, to obtain:

$$\frac{\partial M_x}{\partial x} + \frac{\partial M_y}{\partial y} = 0 \quad (2.10)$$

Differentiating Equation 2.8 with respect to y and Equation 2.9 with respect to x , subtracting, and using Equation 2.10 gives:

$$\begin{aligned} \beta M_y &= \frac{\partial \tau_x}{\partial y} - \frac{\partial \tau_y}{\partial x} \\ M_y &= \frac{\nabla \times \tau}{\beta} \end{aligned} \quad (2.11)$$

where:

- β is the rate of change of Coriolis parameter with latitude ($\text{m} \cdot \text{s}^{-1}$);
- $\nabla \times \tau$ is the vertical component of the curl of the wind stress ($\text{N} \cdot \text{m}^{-3}$).

The Equation 2.11 means that the northward mass transport ($\text{kg} \cdot (\text{m} \cdot \text{s})^{-1}$) of wind driven currents is equal to the curl of the wind stress.

However, in this work we calculated the volume transports ($\text{m}^2 \cdot \text{s}^{-1}$):

$$My = \frac{\nabla \times \tau}{\rho \beta} \quad (2.12)$$

So, the Sverdrup balance (Equation 2.12) refers to the balance between vorticity input to the ocean by the wind and advection of planetary vorticity in the ocean interior. The meridional Sverdrup transports are cumulative from the western boundary, as in Figure 18.

Figure 18: Sverdrup transport (Sv), calculated using Equation 2.12, where blue is clockwise and yellow-red is counterclockwise circulation. Data are from the NCEP reanalysis 1968–1996 (KALNAY et al., 1996). Adapted from Talley et al. (2011).

2.2.3 South Atlantic Meridional Overturning Cell

The South Atlantic part of the Meridional Overturning Cell transport, defined here as the maximum zonally integrated transport from the bottom, was calculated as described in Chapter 1 (Equation 1.1), between 0° and 30°S and 50°W to 10°W.

Only the vertical maximums of the AMOC at 11°S and 27°S were used.

2.2.3.1 Hamming filter

AMOC time series were filtered using a 4 and 10-year Hamming filter. This kind of window, proposed by Richard Hamming is optimised to minimise the maximum (nearest) side lobe.

The coefficients of a Hamming window are computed from the following equation (OPPENHEIM; SCHAFER, 1989):

$$w(n) = 0.54 - 0.46 \cdot \cos\left(2\pi \frac{n}{N}\right), \quad 0 \leq n \leq N \quad (2.13)$$

where N is the length of the window minus 1.

The filtered time series is a result of a convolution between the original series and the filter (w).

2.2.4 Correlations

Correlation quantifies the strength of a linear relationship between two variables. Two variables that are uncorrelated are not necessarily independent, however, because they might have a nonlinear relationship. Here, correlations between two detrended time series (x, y) were calculated using Pearson's product-moment coefficient (RODGERS; NICEWANDER, 1988):

$$R = \frac{\sigma(x,y)}{\sigma_x \sigma_y} \quad (2.14)$$

where:

- $\sigma(x,y)$ is the covariance between x and y ;
- σ_x and σ_y are the standard deviation of x and y .

The covariance ($\sigma_{(x,y)}$) is defined as:

$$\sigma_{(x,y)} = \sqrt{\frac{\sum_{i=1}^n (x_i - \bar{x})^2 - (y_i - \bar{y})^2}{n - 1}} \quad (2.15)$$

And the standard deviation (σ_x):

$$\sigma_x = \sqrt{\frac{\sum_{i=1}^n (x_i - \bar{x})^2}{n - 1}} \quad (2.16)$$

To define whether the correlations with zero lag were significant, a t-test was applied, in which the probability of getting a correlation as large as the observed value by a random chance, when the true correlation is zero, is calculated.

This probability (*p-value*) is computed by transforming the correlation to create a t-statistic having $N - 2$ degrees of freedom, where N is the number of data points in the series. The confidence bounds are based on an asymptotic normal distribution (Equation 2.17), with an approximate variance equal to $\frac{1}{(N-3)}$. These bounds are accurate for large samples when the series have a multivariate normal distribution.

$$p\text{-value} = 0.5 \cdot \ln \left(\frac{1 + R}{1 - R} \right) \quad (2.17)$$

Using a confidence limit of 95%, a correlation is significant when the *p-value* is smaller than 0.05.

2.2.5 Taylor's Diagram

Taylor's Diagrams (TAYLOR, 2001) provide a visual framework for comparing model results to a reference model or, most commonly, to observations, summarising how closely a pattern (or a set

of patterns) matches the reference. The similarity between two patterns is quantified in terms of their correlation, their centered root-mean-square (RMS) difference and the amplitude of their variations (represented by their standard deviations).

It is built in polar coordinates: its angles are the arccosine of the correlation and its radius are the standard deviation. Because the different variables may have widely varying numerical values the model results are normalised by the reference variables. The ratio of the normalised variances indicates the relative amplitude of the model and observed variations. The RMS determines the distance between each data plotted, excluding the average variances.

The reason that each point in the two-dimensional space of the Taylor diagram can represent three different statistics simultaneously (i.e., the centered RMS difference, the correlation, and the standard deviation) is that these statistics are related by the following formula (TAYLOR, 2001):

$$E'^2 = \sigma_f^2 + \sigma_r^2 - 2\sigma_f\sigma_r R \quad (2.18)$$

where

- R is the correlation coefficient between the test and reference fields;
- E' is the centered RMS difference between the fields;
- σ_f and σ_r are the variances of the test and reference fields, respectively.

These diagrams are especially useful in evaluating multiple aspects of complex models or in gauging the relative skill of many different models (e.g. IPCC (2001)).

2.2.6 Optimum Multiparameter

The Optimum Multiparameter (OMP) is a method described by Tomczak (1981) and Tomczak & Large (1989), based on the assumption that the processes involving water masses are linear and all the parameters are affected equally. In this sense, the physical and chemical properties of a water sample is a consequence of mixing of several different Source Water Type (SWT) with identical exchange coefficients for all parameters (KLEIN; TOMCZAK, 1994; KERR, 2006). It assumes that the water mass properties in the formation regions are invariable in time and analyzes time variations in the residual error fields (POOLE; TOMCZAK, 1999).

This method solves a linear system establishing the percentage of each water type (no mixing) in the sample (Equations 2.19 and 2.20). For example, measured temperature at location (k) can be represented by:

$$T_m(k) = x_1(k)T_1 + x_2(k)T_2 + \dots + x_{n_s}(k)T_{n_s} + \epsilon_T(k) \quad (2.19)$$

where:

- x_i is the i th SWT contribution of fraction;
- T_i its temperature ($i = 1, \dots, n_s$);

- $\epsilon_T(k)$ the disturbing noise.

If enough properties are measured, the unknown fractions ($x_1, \dots, x_{n_s}$) at the point k can be estimated by simultaneously optimising all mixing equations (like Equation 2.19) and the balance equation:

$$x_1(k) + \dots + x_{n_s}(k) = 1 \quad (2.20)$$

The procedure can be repeated for all grid points, resulting in optimal values for the SWT at each sampling location. Its core is defined as the area where the percentage of the SWT is greater than 80%.

In this work, OMP was used to produce a correct description of the SAWC and AAIW distribution at 11°S and 27°S, between 37.5–30°W and 48.5–40°W, respectively (Figure 19), once they are the main water masses carried by the WBC related to the return flow of the AMOC.

Figure 19: Position of the sections used to calculate the water masses distribution using OMP, between 37.5–30°W at 11°S and between 48.5–40°W, at 27°S.

As discussed in Section 1.2.2, the SACW is a combination of three water masses: ISACW, SASTMW and SAMW (their density layers are shown in Table 1). However, according to Klein & Tomczak (1994) and Poole & Tomczak (1999), in the South Atlantic its salinity-temperature relationship can be approximated by a straight line (Figure 21), and thus, two water types are sufficient to represent the SACW (the end-points of the linear section). The AAIW, is more uniform in its formation region and can be represented by a single water type (TOMCZAK; LIEFRINK, 2005).

OMP analysis produces better results when applied to water masses with reasonably conserved properties. Within the mixed layer, the hydrographic properties of a water parcel are strongly influ-

enced by the atmosphere through air-sea interaction and by sunlight via near-surface biological processes. Once removed from the mixed layer, the temperature and salinity of a water parcel are conserved, which means that these properties remain constant unless mixing occurs with another water parcel (POOLE; TOMCZAK, 1999). Therefore, to avoid the mixed layer (Figure 20), only salinity, temperature and potential vorticity fields below 150m (at 11°S) and 200m (at 27°S) were used.

Figure 20: Averaged maximum mixed layer depth, using the 5 deepest mixed layers in 1×1 bins from the Argo profiling float data set (2000–2009) and fitting the mixed layer structure as in Holte & Talley (2009). Adapted from Talley et al. (2011).

Thus, two source waters represent the SACW, ISACW and SAMW, based on the results Donners, Drijfhout & Hazeleger (2005) (Table 1). Salinity and temperature for these water masses were calculated using the longitude-time averaged temperature-salinity diagram, as shown in Figure 21a.

AAIW source water type salinity and temperature were calculated using the mixing triangle (MAYEV, 1975). The tangent lines were calculated at the points where the longitude-averaged temperature-salinity diagram crosses the isopycnal related to the interface of the water masses (Figure 21)b.

But, below AAIW there is the UCDW, and to correctly isolate the intermediate water mass is necessary to account its mixture with the deep water.

The optimal solution for this problem would be isolate the three water masses (i.e., four SWT) in the same water analysis. However, as highlighted by Brauwere et al. (2007), the number of unknown fractions (n_s) may not exceed the number of measured variables (n_v) + 1. Otherwise, there is an infinite number of solutions which are all equally good. In this work, besides temperature and salinity data, only the potential vorticity is available (Section 2.2.6.1), which makes impossible to separate SACW, AAIW and UCDW at once.

To make the analysis possible, the water column was divided into parts, as follows. First, the SACW (i.e. a mixture of ISACW and SAMW) was isolated from AAIW. Only data below the maximum mixing layer and between the densities of $25.5 \text{ kg} \cdot \text{m}^{-3}$ and $27.5 \text{ kg} \cdot \text{m}^{-3}$ were used. The goal was to obtain a good description of the SACW core. Second, data at depths shallower than 2000m and with densities

Figure 21: Time averaged temperature-salinity diagram at 25°S and 25°W. Original data from the Simple Ocean Data Assimilation System, version 2.2.4 (SODA, Carton & Giese (2008)). SACW STW thermohaline indices (a). Gray contours represent the common end-points regarding Keffer (1985) and Donners, Drijfhout & Hazeleger (2005) density layer definitions. AAIW STW thermohaline indices (b). Gray contours indicate the density limits between SACW, AAIW and NADW. Blue dashed lines indicate the tangents used to generate the mixing triangle. For both plots, black points and black dotted lines indicate the salinity and temperature data used as water types parameters.

greater than $26.7\text{kg}\cdot\text{m}^{-3}$ were used to separate AAIW. Only the lower part of the SACW was used since it was not the main objective of the analysis, and so, the upper limit of AAIW layer in this case is SAMW. According to McCartney (1982), mode waters are homogeneous and thus, one STW is enough to represent it. The isopycnal $26.6\text{kg}\cdot\text{m}^{-3}$, the upper limit of this water mass layer, was chosen to obtain the index. The lower limit is the UCDW. To characterise this water mass, the isopycnal $27.75\cdot\text{kg}^{-3}$, the lower limit of its layer was chosen.

A complete list of the parameters, with their weights, selected empirically, is presented in Tables 28 and 29, in Section 3.2.2.

2.2.6.1 Potential vorticity

The third parameter used in OMP analysis is the potential vorticity once it is conserved along a fluid trajectory. According to Pedlosky (1987), it is defined as:

$$\Pi = \frac{\zeta + f}{H} \quad (2.21)$$

where:

- Π is the potential vorticity ($\text{m}\cdot\text{s}^{-1}$);
- ζ is the relative vorticity (s^{-1});

- f is the planetary vorticity, related to the Coriolis parameter (s^{-1});
- H is the depth (m).

For baroclinic flow in a continuously stratified fluid, the potential vorticity can be written as:

$$\Pi = \frac{\zeta + f}{\rho} \frac{\partial p}{\partial z} \quad (2.22)$$

where:

- ρ is the density ($\text{kg}\cdot\text{m}^{-3}$);
- $\frac{\partial p}{\partial z}$ is pressure variation with depth ($\text{N}\cdot\text{m}^{-3}$).

Assuming the horizontal gradients of density are small compared with the vertical ones, which is a good assumption in the thermocline, and also that in most of the interior of the ocean, $f \gg \zeta$, Equation 2.22 is written, according to Pedlosky (1996), as:

$$\Pi = \frac{f}{\rho} \frac{\partial p}{\partial z} \quad (2.23)$$

This allows the potential vorticity of various layers of the ocean to be determined directly from hydrographic data without knowledge of the velocity field.

Using Equation 2.23 and the longitude-time averaged salinity and temperature data, the potential vorticity profile is calculated, and the value for each source water found (Figure 22).

2.2.7 Neutral density

A parcel of water moves locally along a path of constant density keeping below less dense water and above more dense water. More precisely, it moves along a path of constant potential density σ_z referenced to the local depth z . Such a path is called a neutral path (EDEN; WILLEBRAND, 1999). A neutral surface element is the surface tangent to the neutral paths through a point in the water. No work is required to move a parcel on this surface because there is no buoyancy force acting on the parcel as it moves, if we ignore friction (McDOUGALL, 1987; STEWART, 2008).

Jackett & McDougall (1997) developed a practical neutral density variable γ_n and surface that stays within a few tens meters of an ideal surface anywhere in the world. The neutral density values were then used to label the data in the Levitus atlas. This prelabeled data set is used to calculate γ_n at new locations where the potential temperature (θ_0) and the salinity (S) are measured as a function of depth by interpolation to the four closest points in the Levitus atlas. Recently, Serazin (2011) has made some improvements, applying new equations of the thermodynamic properties of seawater (TEOS-10⁵), and each γ_n surface is calculated using specific equations for each ocean basin. Through this practice, neutral density γ_n is a function of salinity, temperature potential, pressure, longitude, and latitude.

⁵ International Thermodynamic Equation Of Seawater, available in www.TEOS-10.org.

Figure 22: Density (a) and vorticity (b) profiles in the first 2000m at 25°W and 25°S, constructed using the time-averaged temperature and salinity data from SODA 2.2.4 (CARTON; GIESE, 2006; CARTON; GIESE, 2008).

According to McDougall (1987), The neutral surface defined above differs from the potential density surfaces because the compressibility of the seawater is a function of salinity and potential temperature, or, equivalently, because $\frac{\alpha}{\beta}$ is dependent on pressure, where α is the thermal expansion coefficient and β is the haline contraction coefficient (Figure 23). It also differs only slightly from an ideal neutral surface. If a parcel moves around a gyre on the neutral surface and returns to its starting location, its depth at the end will differ by around 10m from the depth at the start. If potential density surfaces are used, the difference can be hundreds of meters, a far larger error (STEWART, 2008).

In this work, we calculated the salinity and the temperature in the neutral surfaces that corresponds to SACW and AAIW in each model, in the sections shown in Figure 19, using the same indices calculated to the OMP analysis (Section 2.2.6), i.e., temperature and salinity that correspond to the $25.5 \text{ kg} \cdot \text{m}^{-3}$ and $27 \text{ kg} \cdot \text{m}^{-3}$ isopycnals for SACW and the results of the mixing triangle for AAIW.

Figure 23: Potential (a) and neutral (b) densities calculated near South America coast (section A17). Adapted from Koltermann, Gouretski & Jancke (2011).

3 Results and discussion

This chapter presents and discusses the results, splitted in two main sections: Section 3.1 studies the SEC bifurcation variability at interannual time scales, both at the surface and intermediate depths and its relationship with the WBC and AMOC changes. Following, the results related to changes in the water masses are presented and discussed in Section 3.2.

3.1 *South Equatorial Current Bifurcation and its link with the AMOC*

To study the variability of the SEC bifurcation and its relationship to the AMOC, we based our analysis in the methodology proposed by Rodrigues, Rothstein & Winbush (2007), starting with a section of meridional velocities within a 4° longitude band off the South America coast (Figure 17). Its climatological values are shown in Figure 24. It should be noted that SODA, CCSM and ORCA05 results correspond to an average of the velocities from 1948 to 2007, while ORCA12 results cover the period from 1989 to 2007.

The first important thing to notice here is that the intensity of the northward flow is stronger in the reanalysis (Figure 24a) than in the other three model outputs, specially at intermediate depths (under 600m) where the flux of the IWBC is dominant. It is also clear that in ORCA05 model has the strongest and deepest Brazil Current (BC), and so, the bifurcation does not extend beyond 24°S (Figure 24c) where the Vitória Trindade Ridge is located (SILVA et al., 2009). According to Danabasoglu et al. (2013), compared to other models forced by CORE-II data, ORCA05 has a weak AMOC which can be linked to the weak northward flow observed in this work. This suggests a longer influence of the AMOC in the bifurcation position at intermediate depths.

Except for ORCA05 results, the models reproduce well the bifurcation position observed in previous studies (STRAMMA; ENGLAND, 1999a; BOEBEL et al., 1999; WIENDERS; ARHAN; MERCIER, 2000). The authors found a poleward depth increase of the southern branch of SEC (sSEC) bifurcation along the Brazilian coastline. Their observations suggest that the SEC bifurcation occurs at $14^\circ\text{--}16^\circ\text{S}$ in the surface layer (0–100m), at $20^\circ\text{--}24^\circ\text{S}$ in the thermocline (400–500m), and at $26^\circ\text{--}28^\circ\text{S}$ in the intermediate layer (600–1200m). Table 2 shows the mean bifurcation position for each model results compared to Rodrigues, Rothstein & Winbush (2007). The results presented in Table 2 agree with those found by these authors.

Using the wind stress curl applied to the Sverdrup Theory (Equation 2.12), the bifurcation position at the surface (here taken as the zero velocity contour) was calculated (Figure 25). As expected, all the forcing respond well to the literature (e.g. Talley et al. (2011)) with transports maximum (around 25Sv,

Figure 24: Climatological meridional velocity within the 4° longitude band off South America coast for SODA (a), CCSM (b), ORCA05 (c) and ORCA12 (d) datasets. Black contours indicate the position of the bifurcation.

Table 2: Bifurcation position, in °S, means for each dataset in different layers, compared to Rodrigues, Rothstein & Winbush (2007)¹.

Layers (m)	¹ Literature	SODA	CCSM	ORCA05	ORCA12
Position and standard deviation in °S					
0–100	14–16	15.2±3.2	17.7±2.1	14.4±4.0	14.2±2.1
100–500	14–20	22.8±1.6	22.2±1.4	20.8±1.3	22.2±2.8
500–1000	21–28 ²	27.6±1.2	26.6±0.9	22.5±0.2	26.8±0.5

¹Rodrigues, Rothstein & Winbush (2007): the depth range is 500–1200m.

at 30°S) and the bifurcation close to 14°S, agreeing with the Sverdrup's Theory. Results from CCSM and ORCA05, both forced by CORE data, show some significant differences in the bifurcation position. This result is consistent to those found in the top 100m, using average of the meridional velocities within the 4° longitude band off the Brazilian coast (Table 2) and manifest the importance of the wind in the bifurcation position.

Figure 25: Sverdrup transport (Sv) calculated using the wind stress curl (Section 2.2.2.1).

Based in this main results, this Section is structured as follows: Section 3.1.1 shows the validation results, when we reproduce other analysis done by Rodrigues, Rothstein & Winbush (2007) at seasonal timescales; Section 3.1.2 shows the results for SEC bifurcation and its relationship with wind stress curl and the climate indices; Section 3.1.2.3 shows the results of AMOC and its relationship with the currents.

3.1.1 Validation

Initial analysis with the model outputs aimed at reproducing, the results obtained by Rodrigues, Rothstein & Winbush (2007).

Figure 26 shows the climatological summer and winter sections of meridional velocity within the 4° longitude band off South America coast. It is possible to see differences in the position of the bifurcation, predominantly in the surface layer (top 200m). This seasonal cycle dependence of the position of the bifurcation, mainly in the first 100m was described by Schott et al. (2005), Rodrigues, Rothstein & Winbush (2007), Silva et al. (2009). In Figure 26, during the summer (monthly averages of January, February and March) the bifurcation, in the top 100m, is north of 10°S while in the winter time (July, August and September) it is mainly south of this latitude. The changes in the bifurcation position below 500m, are not clear in this figure.

The seasonality of the bifurcation can be clearly seen in Figure 27. CCSM, ORCA05 and ORCA12

Figure 26: Averaged summer and winter meridional velocity within the 4° longitude band off South America coast section for SODA (a,b), CCSM (c,d), ORCA05 (e,f) and ORCA12 (g,h) datasets. Black contours lines indicate the position of the bifurcation. Gray dashed contours are the climatological reference (Figure 24).

results agree with SODA in the top 100m: the bifurcation reaches its northernmost position during the winter and the southernmost during the summer. It also confirms the results of Rodrigues, Rothstein

& Winbush (2007).

Differently from the surface, the variability at intermediate layers has a semi-annual cycle in SODA (Figure 27a). Different regimes in the bifurcation variability at the surface and at intermediate layers were described by Rodrigues, Rothstein & Winbush (2007) and, according to the authors, it is a result of a combination of both direct and remote wind forcing, and could be also, related to AMOC variability. CCSM fails to reproduce this variability, while ORCA05, despite its underestimated southward bifurcation extension (Figures 24 and 26), reproduces it better, although they are both forced by CORE-II (Section 2.1.4). ORCA12 shows a higher frequency variability. All models in this study show big differences when compared to the observations and the SODA reanalysis. This can also be related with dissimilar variability of the SEC, that will be discussed later. Another explanation for the observed differences could be variations of kinetic energy and buoyancy fluxes. Further analysis is necessary to test this hypothesis and to understand the mechanisms that drive the different responses for each model output.

Figure 28 shows the average position of the bifurcation in the top 200m. It is possible to note that CCSM, ORCA05 and ORCA12 have the same variability, but, while the southernmost bifurcation position was expected to happen between June and July (Rodrigues, Rothstein & Winbush (2007)), it happens a month earlier. In SODA reanalysis results the averaged bifurcation position in the top 200m has an influence of the semi-annual variability, and the southernmost position of the bifurcation occurs in September. In a similar way, Figure 29 shows a seasonal cycle for the bifurcation position averaged between 500 and 1000m. As it was shown in Figure 27, ORCA05 underestimates the southward extension of the bifurcation, and, because of that, it has the northernmost position values. SODA, CCSM and ORCA12, in general, agree with respect to the averaged position at intermediate depth, but, they have big differences in the seasonal variability. Rodrigues, Rothstein & Winbush (2007) show that, at intermediate depths, the bifurcation variability has a semi-annual cycle, which SODA reproduces well, showing two maxima (during May and November) and two minima (March and August). CCSM shows an annual cycle, similar to that found at the surface (Figure 28), while ORCA05 and ORCA12 approach the reanalysis.

As described above, it is known that the bifurcation position at seasonal timescales has a strong influence of the wind stress curl. Figure 30 shows, for each dataset, the monthly mean position of the zero wind stress curl contour. It is possible to observe the meridional excursion of the zero wind stress curl line through the bifurcation region during the year, related to the seasonal movement of the Atlantic ITCZ. The model results agree well with the SODA reanalysis and with the results of Rodrigues, Rothstein & Winbush (2007): the southern zero line is almost horizontal during the winter, while its inclination is bigger during summer and spring. As consequence, the bifurcation is exposed to negative (positive) wind stress curl and associated weaker (stronger) Ekman pumping from February to August (September to January).

The time series of the bifurcation position averaged in the top 200m and the wind stress curl at each grid point were correlated at zero lag to better discuss their link at seasonal time scales. The results can be

Figure 27: Seasonal cycle of the position of the bifurcation of SEC, for SODA (a), CCSM (b), ORCA05 (c) and ORCA12 (d) datasets, calculated at top 1000m using the zero contour line of meridional velocity, as proposed by Rodrigues, Rothstein & Winbush (2007) (Section 2.2.1), between 30°S and 5°S. Contours at every 1°. White contours show the depth of the bifurcation at 11°S and 27°S.

seen in Figure 31. The correlations are strong in a large part of the area, but only correlation coefficients statistically significant at the 95% level are considered. Big differences can be seen between the SODA reanalysis results and the models, specially in the southern limit of the area. In the models, it is possible to observe a east-west dipole in correlation, in the same way found by Rodrigues, Rothstein & Winbush (2007), centered around 10°W, with positive correlations close to the Brazilian coast and negative ones in the east part of the basin (less significant in SODA). It is not clear if changes in the wind stress over the bifurcation region are more important than those in the central and eastern side of the basin as discussed by Rodrigues, Rothstein & Winbush (2007).

Based on literature information and in the results of Figure 24, two main sections were chosen to access the variability of the bifurcation position: 11°S (Figure 32) and 27°S (Figure 34). A section in 25°W (Figure 36) is also used to monitor the South Equatorial Current (SEC).

It is easy to note the effect of different model resolutions. In the CCSM model, the NBUC and the

Figure 28: Seasonal variability of the SEC bifurcation position, averaged in the top 200m, for SODA (blue), CCSM (red), ORCA05 (yellow) and ORCA12 (green) datasets, calculated using the zero contour line of meridional velocity, as proposed by Rodrigues, Rothstein & Winbush (2007).

Figure 29: Seasonal variability of the SEC bifurcation position, averaged between 500 and 1000m, for SODA (blue), CCSM (red), ORCA05 (yellow) and ORCA12 (green) datasets, calculated using the zero contour line of meridional velocity, as proposed by Rodrigues, Rothstein & Winbush (2007).

BC are 1.5 times larger (Figures 32b and 34) than in the other datasets. ORCA12 (Figure 32d) has the strongest velocities.

In Figure 32, there is also a recirculation cell, formed by the NBUC and an adjacent southward flux, described in Schott et al. (2005) at 11°S, from observational data. Comparing the results of this work to those found by the authors between 2000 and 2004, ORCA12 is the model simulation that best reproduces the observed velocity section, with velocities $\sim 50 \text{ cm} \cdot \text{s}^{-1}$ at the NBUC core and $12 \text{ cm} \cdot \text{s}^{-1}$ for the offshore recirculation. In ORCA05, the NBUC is shallower than in the other model results, once again, agreeing with the weak MOC found by Danabasoglu et al. (2013).

The volume transport of the NBUC was calculated using only the positive (northward) velocities

Figure 30: Month averages of zero position of wind stress curl, for SODA (a,b), CCSM (c,d), ORCA05 (e,f) and ORCA12 (g,h) datasets. Red contours are the averages for January (solid line), February (dashed) and March (dotted). Blue ones, averages for July (solid line), August (dashed) and September (dotted).

at 11°S (Section 2.2.2). Its seasonality is shown in Figure 33. SODA and ORCA05 have the strongest and the weakest NBUC transports respectively. Averaged values are shown in Table 3.

The seasonal cycle shows that the lowest transport values of NBUC are found during November and December while the maximum transport occurs in June. This variability was described by Schott

Figure 31: Correlation between the seasonal cycles of bifurcation position in the top 200m (calculated using the zero contour line of meridional velocity, as proposed by Rodrigues, Rothstein & Winbush (2007)) and of the wind stress curl for SODA (a), CCSM (b), ORCA05 (c) and ORCA12 (d). Black contours mark the areas where the correlations coefficients are statistically significant at 95%.

et al. (2005) and by Schott et al. (1993), Johns et al. (1998) from observations at 5°S and is related to the position of the bifurcation (RODRIGUES; ROTHSTEIN; WINBUSH, 2007; SILVA et al., 2009). When the bifurcation is in its southernmost position (during winter, Figures 27 and 28), the northward flux of NBUC is maximum, and the opposite occurs during summer time. The seasonal cycle found here does not correlate with the NBC transport seasonal time series calculated by Johns et al. (1998) at ~5°N.

CCSM and ORCA12 climatological NBUC transports at 11°S, showed in Table 3 are closer to the ones described by Silva et al. (2009), using a regional model, between 14°S and 8°S. Schott et al. (2005), from observations, in this latitude found a NBUC transport of $21.7\text{Sv} \pm 4.1\text{Sv}$.

In Figure 34, in spite of the difference in horizontal resolution, CCSM results (Figure 34b) are closer to the SODA reanalysis (Figure 34a) than ORCA05 (Figure 34c). In the ORCA05 results, the BC is much deeper than in the others and the Intermediate Western Boundary Current (IWBC) is extremely weak. This result is consistent with the shallowest and the weakest northward flow observed in Figure 24, and also agrees with Danabasoglu et al. (2013). ORCA12 (Figure 34d), as in the section at 11°S (Figure 32d), has the strongest surface current.

According to Evans & Signorini (1985), Campos, Gonçalves & Ikeda (1995), Silveira et al. (2000),

Figure 32: Time averages of meridional velocities ($\text{m}\cdot\text{s}^{-1}$) at 11°S , for SODA (a), CCSM (b), ORCA05 (c) and ORCA12 (d) datasets. Black contours indicates the $0\text{ m}\cdot\text{s}^{-1}$ line.

Table 3: Meridional volume transport (Sv) of NBU at 11°S , in the top 1000m, calculated in the sections shown in Figure 34.

	SODA	CCSM	ORCA05	ORCA12
		(Sv)		
average	25.9 ± 2.0	22.1 ± 1.8	15.5 ± 1.8	22.2 ± 1.1
summer	25.2 ± 3.7	21.3 ± 2.7	14.8 ± 2.7	22.6 ± 3.6
winter	26.6 ± 3.9	23.1 ± 2.8	16.2 ± 2.6	22.6 ± 3.5

at 28°S , the BC extends until 750m. Close to this latitude, its mean observed velocities reaches $50\text{ cm}\cdot\text{s}^{-1}$ (MÜLLER et al., 1998; SILVEIRA et al., 2000). So, SODA and CCSM results reproduce better the vertical extent of the BC while ORCA12 shows mean velocities closer to the results found in the literature.

The IWBC is a narrow and weak western boundary current, with velocities that are hardly bigger than $30\text{ cm}\cdot\text{s}^{-1}$, and the current flow rapidly accelerates away from the formation region, at the Santos Bifurcation (BOEBEL et al., 1999; SILVEIRA; CALADO; CASTRO, 2004). According to Legeais,

Figure 33: Seasonal cycle of the volume transport (Sv) of NBUC at 11°S, for SODA (a), CCSM (b), ORCA05 (c) and ORCA12 (d) datasets, calculated at the top 1000m as shown on Figure 32. The dashed blue lines indicate the standard deviation for each month.

Ollitrault & Arhan (2013), close to this latitude, at 26°S, the current reaches velocities above $15\text{cm}\cdot\text{s}^{-1}$.

As the transports calculated in Figure 35 depend on the sections showed in Figure 34, ORCA05 has the strongest southward flux (while it had the weakest northward transports at 11°S, as shown in Figure 33). In fact, the average transports for the ORCA05 results at 27°S is twice the value calculated for CCSM and SODA (Table 4). The volume transport of BC is maximum during the summer (Figure 35), while the bifurcation is further north (Figures 27 and 28), and the northward transport of NBUC is minimum (Figure 33). This behaviour was documented by Schott et al. (2005), Rodrigues, Rothstein & Winbush (2007), Silva et al. (2009). Observing Figure 35, it is clear that neither SODA nor ORCA12 show a well marked seasonal cycle.

Although the BC seen from ORCA05 results is the one that most differs from the SODA reanalysis, its transport value in the top 1000m is the closest to those found in the literature (Table 4). According to Müller et al. (1998), at 28°S the BC transport is 16Sv.

In Figure 36, the main difference between the SODA reanalysis (Figure 36a) and the model results is the intensity of the zonal velocities of SEC (stronger in SODA). As described by Stramma (1991), the south sSEC has velocities $\sim 10\text{cm}\cdot\text{s}^{-1}$. According to the same authors, south of 10°S, west of Green-

Figure 34: Time averages of meridional velocities ($\text{m}\cdot\text{s}^{-1}$) at 27°S , for SODA (a), CCSM (b), ORCA05 (c) and ORCA12 (d) datasets. The black contours indicates the $0\text{m}\cdot\text{s}^{-1}$ line.

Table 4: Meridional volume transport (Sv) of BC at 27°S , in the top 1000m, calculated in the sections shown in Figure 34.

	SODA	CCSM	ORCA05	ORCA12
			(Sv)	
average	-6.9 ± 3.2	-6.8 ± 2.3	-15.0 ± 3.2	-12.7 ± 4.5
summer	-7.3 ± 3.0	-7.3 ± 2.1	-15.2 ± 2.9	-11.0 ± 3.1
winter	-6.1 ± 2.9	-6.3 ± 2.1	-14.9 ± 3.1	-13.6 ± 4.9

which Meridian, the SEC is a broad and sluggish flow. One can identify in this section two near-surface cores of sSEC, specially in the SODA reanalysis (Figure 36a): $\sim 18^\circ\text{S}$ and $\sim 13.5^\circ\text{S}$. These cores were observed by Stramma (1991), Stramma & England (1999b), Lumpkin & Garzoli (2005), Silva et al. (2009). According to Wienders, Arhan & Mercier (2000), the poleward boundary of the SEC's southern branch is at 25°S through the whole thickness of the subtropical gyre, but the latitude of the northern boundary varies from 7.5°S at the surface and in the western South Atlantic, to 27°S at 1400m depth and

Figure 35: Seasonal cycle of the volume transport (Sv) of BC at 27°S, for SODA (a), CCSM (b), ORCA05 (c) and ORCA12 (d) datasets, calculated at the top 1000m as shown on Figure 34. The dashed blue lines indicate the standard deviation for each month.

in the eastern basin. A third core is observed in the north limit of the section, and corresponds to the south part of the central branch of the SEC (cSEC) (LUMPKIN; GARZOLI, 2005).

Between the sSEC and cSEC, a near surface eastward core is identified in almost all the datasets (Figure 36a,b,c) and also by Silva et al. (2009). It is the upper core of the South Equatorial Counter-current (SECC) (BONHOURE et al., 2004). Under the SEC, we find an eastward flux, continuous in CCSM and ORCA05 results (Figure 36b,c) while it appears as several branches in the SODA reanalysis and ORCA12 results (Figure 36a,d). The same branches were observed by Silva et al. (2009), in their 1°/12 regional model.

As described in Section 2.2.2, the transports of the SEC were calculated for different regions, centered in the latitude of the two main sections: between 13°S and 9°S and between 29°S and 25°S.

Figure 37 and Table 5 show the seasonal cycle of the SEC transport, between 13°S and 9°S. Observing the time series, it is possible to verify that the ORCA05 results have the lowest volume transport and the reanalysis the largest. ORCA12 results have the smallest amplitude variability. In this area, the transport calculated is a mix of the cSEC (between SEC (3°S–5°S) and the South Equatorial Counter Current (6°S–9°S), according to Molinari, Voituriez & Duncan (1981), Peterson & Stramma (1990),

Figure 36: Time averages of zonal velocities ($\text{m}\cdot\text{s}^{-1}$) at 25°W , for SODA (a), CCSM (b), ORCA05 (c) and ORCA12 (d) datasets. The black contours indicates the $0\text{m}\cdot\text{s}^{-1}$ line.

Stramma (1991)) and sSEC (south of the SECC). The transport of cSEC is about 17Sv in the first 200m with typical variation between 7 to 26Sv (MOLINARI, 1982), while the transport of sSEC in the top 500m is about 20Sv (BONHOURE et al., 2004). Also, in all the datasets, the SEC is weaker between May and June, while its maximum occurs between August and October in the SODA reanalysis, and between October and November in the models. The results of Molinari (1983), Stramma (1991), the SEC is stronger during the austral winter (July and August), which makes the SODA results closer to the observations.

Figure 38 shows the SEC transport, between 25°S and 29°S . CCSM and ORCA05 results show that the strongest values of SEC transport occur during winter. SODA seems to have a semi-annual cycle, but the high standard deviation in the last 3 months masks it.

Table 6 shows the averages for the SEC transport. SODA and ORCA12 show similar results, but they are much smaller than those found in the literature. According to Bonhoure et al. (2004), the sSEC transport measured in the upper and intermediate layers is 49Sv. As the limit of the sSEC is usually between 10°S and 25°S , such differences can be related to the area where the transports were calculated. Boebel et al. (1999) found that near 30°S , the sSEC (called by the authors “Return Current”) transports

Figure 37: Seasonal cycle of the volume transport (Sv) of SEC at 25°W, for SODA (a), CCSM (b), ORCA05 (c) and ORCA12 results, calculated at the top 1000m as shown on Figure 36, between 13°S and 9°S. The dashed blue lines indicate the standard deviation for each month.

Table 5: Zonal volume transport (Sv) of SEC at 25°W, in the top 1000m, between 13°S and 9°S calculated in the sections shown in Figure 36.

	SODA	CCSM	ORCA05	ORCA12
			(Sv)	
average	-8.8 ± 3.3	-7.1 ± 3.4	-3.5 ± 1.6	-7.1 ± 1.8
summer	-8.3 ± 3.0	-7.8 ± 2.8	-3.8 ± 1.3	-7.5 ± 1.7
winter	-9.8 ± 3.5	-6.5 ± 3.0	-3.2 ± 1.5	-6.8 ± 1.6

12 ± 3 Sv, which is closer to the results obtained in this work.

As both the NBUC and the BC originates from the bifurcation of the SEC close to 11°S, the seasonal time series of the transports of these currents at 11°S (Figure 33) and 27°S (Figure 35) respectively were also correlated with the seasonal time series of the SEC at 25°W, between 13°S and 9°S (Figure 37). The results are shown in Tables 7 and 8. The results from all the models present high, positive correlations between the transports of the SEC and the NBUC (Table 7): which means that, when the SEC flux is minimum (or, more positive values), the NBUC transport is maximum (also more positive values)

Figure 38: Seasonal cycle of the volume transport (Sv) of SEC at 25°W, for SODA (a), CCSM (b), ORCA05 (c) and ORCA12 (d) datasets, calculated at the top 1000m as shown on Figure 36, between 29°S and 25°S. The dashed blue lines indicate the standard deviation for each month.

Table 6: Zonal volume transport of SEC at 25°W, in the top 1000m, between 29°S and 25°S calculated in the sections shown in Figure 36.

	SODA	CCSM	ORCA05	ORCA12
	(Sv)			
average	-10.2 ± 3.1	-7.6 ± 1.9	-4.1 ± 1.6	-9.8 ± 2.5
summer	-10.6 ± 3.0	-7.6 ± 1.8	-4.1 ± 1.1	-10.2 ± 2.2
winter	-10.0 ± 2.8	-7.7 ± 1.9	-4.1 ± 1.2	-9.2 ± 2.2

while the BC transport decreases. Only the low resolution models show correlations between the SEC and BC (Table 7). This is because the seasonal cycle of the transport of the BC in SODA and ORCA12 at 27°S is not well marked as in the other two models, which can indicate an influence of other physical processes, as mesoscale activity (ORCA12).

According to Rodrigues, Rothstein & Winbush (2007), since the bifurcation latitude variability is a manifestation of changes in the flow of the currents involved, the local component of the forcing, responsible for most of the variability in the sSEC bifurcation (Figure 30), explains the decrease in the

Table 7: Correlation between monthly averages of the transport of NBUC at 11°S (Figure 33) and seasonal transports of SEC at 25°W, between 13°S and 9°S (Figure 37).

	SODA	CCSM	ORCA05	ORCA12
correlation	0.40	0.99	0.99	0.79
p-value	0.20	<0.01	<0.01	<0.01

Table 8: Correlation between monthly averages of the transport of BC at 27°S (Figure 35) and seasonal transports of SEC at 25°W, between 13°S and 9°S (Figure 37).

	SODA	CCSM	ORCA05	ORCA12
correlation	-0.15	0.99	0.96	0.12
p-value	0.64	0.01	<0.01	0.70

transport of both SEC and BC during winter and the subsequent increase in summer. This can be seen in Figure 39, in which the transports were correlated to the wind stress curl between 30°S and 10°N and 50°W to 10°E.

The negative and significant correlations between seasonal transports at 11°S and wind stress curl time series are larger than 50% in all the datasets in the northwest South Atlantic (Figure 39a,c,e,g), reaching 15°W. These high negative correlations in the west part of South Atlantic extends from 18–15°S to 3°S. It has also a positive correlation with the southeast part of the area. These regions of high correlations are consistent to those found in Figure 30 (correlation between wind stress curl and the bifurcation position in the surface), suggesting a strong relationship between wind stress curl, bifurcation position and NBUC transports. In fact, as shown in Table 9, and as predicted in the literature (RODRIGUES; ROTHSTEIN; WINBUSH, 2007; SILVA et al., 2009), the correlations between the bifurcation position and the northward transports of the NBUC are greater than 80% in all models. Also, the bifurcation position and the transports at 25°W are well correlated. All these results clearly suggest the dominance of the Sverdrup balance as the main dynamical processes controlling the phenomena in the surface.

The correlations between the seasonal transport at 27°S and the wind stress show the same pattern described above (Figure 39b,d,f) except for ORCA12 results (Figure 39h) where no significant correlations with the local wind stress curl were found. Looking at Figure 35d, that shows the transport of the BC, it can be seen that the seasonal cycle is influenced by a high frequency variability, that can be a result of the mesoscale activity considering that ORCA12 is a high resolution model and the BC was calculated in 27°S, far away from the bifurcation region.

The situation, however, is not as clear at intermediate depths. There is no significant correlations between the SEC transports at 25°W, between 27°S and 23°S and the bifurcation position (Table 11), even for SODA, that responds well to the observations (RODRIGUES; ROTHSTEIN; WINBUSH,

Figure 39: Correlations at zero lag of seasonal time series of wind stress curl with the transport of NBUC at 11°S (left panels) and the transport of BC at 27°S (right panels), for SODA (a,b), CCSM (c,d), ORCA05 (e,f) and ORCA12 (g,h) datasets. Black contours mark the areas where the results are significant (p-value is lower than 0.05).

2007). Huge differences between CCSM and ORCA05 suggests that the wind stress is not the mechanism controlling the bifurcation position at intermediate depths.

Table 9: Correlation between monthly averages of the bifurcation position in the top 200m (Figure 28 and seasonal transports of NBUC at 11°S (Figure 33).

	SODA	CCSM	ORCA05	ORCA12
correlation	-0.92	-0.87	-0.94	-0.86
p-value	<0.01	<0.01	<0.01	<0.01

Table 10: Correlation between monthly averages of the bifurcation position in the top 200m (Figure 28 and seasonal transports of SEC at 25°W, between 13°S and 9°S (Figure 37).

	SODA	CCSM	ORCA05	ORCA12
correlation	0.33	-0.86	-0.96	-0.92
p-value	0.29	<0.01	<0.01	<0.01

Table 11: Correlation between monthly averages of the bifurcation position in the bottom 500m (Figure 29) and seasonal transports of SEC at 25°W, between 27°S and 23°S (Figure 38).

	SODA	CCSM	ORCA05	ORCA12
correlation	-0.11	-0.40	-0.33	0.43
p-value	0.73	0.20	0.30	0.16

3.1.2 Interannual variability

This section shows the results related to the interannual variability of the SEC bifurcation. To avoid the influence of the seasonal cycle, all the results are annual.

3.1.2.1 South Equatorial Currents bifurcation

The section of annual averages of meridional velocities within a 4° longitude band off South America (exemplified in Figure 24) was used to determine the SEC bifurcation position (Figure 40). The deeper the bifurcation, the greater the amplitude of the variability in each latitude. This is a new result, so, there is no comparison with the literature.

Following the same methodology, the bifurcation position was averaged in the top 200m and between 500m and 1000m. Figure 41 shows the annual means of the averaged top 200m bifurcation position. As already described in Figure 28, ORCA05 and ORCA12 display the northernmost bifurcation latitude and the CCSM run the southernmost. Forced by the same dataset, CORE-II (Section 2.1.4), despite of the differences in mean latitude position, ORCA05 and CCSM show similar time series (Table 12), which did not happen in seasonal time scales (Figure 27). Comparing all the time series, we found good correlations between CCSM and ORCA05 (almost 80%), and between SODA and ORCA12 (more than 67%), indicating that these models have similar variability. Unexpectedly, no cor-

Figure 40: Position of the bifurcation of SEC time series (annual averages), for SODA (a), CCSM (b), ORCA05 (c) and ORCA12 datasets, calculated at top 1000m, using the zero contour line of meridional velocity, as proposed by Rodrigues, Rothstein & Winbush (2007) (Section 2.2.1), between 30°S and 5°S. Contours are every 1°.

relations between CORE-II forced models and the SODA reanalysis were found. Except for ORCA05, the datasets showed a positive trend (Table 14), indicating an equatorward movement of the bifurcation position, which corroborates with recent studies showing a scenario of expansion (40%) of the subtropical gyre (GARZOLI; MATANO, 2011).

As already discussed in the beginning of this chapter, ORCA05 underestimates the southern limit of the bifurcation position, and its mean latitude, between 500m and 1000m, does not reach 25°S. According to Müller et al. (1998), Stramma & England (1999b), Boebel et al. (1999) south of SEC bifurcation, the IWBC goes south, enjoying BC, and both currents can be seen as one, with an increased vertical extension. It also explains the inexistent IWBC in ORCA05 results at 27°S, and because of that, the IWBC transports will not be analyzed in this model output.

Between 500m and 1000m, only the models forced by CORE-II data showed a similar variability (Table 13). ORCA12 and SODA show good correlations, but they were not statistically significant at 95%. Also, differently from the top 200m, at intermediate depths, the bifurcation is going south (Ta-

Figure 41: SEC bifurcation position time series, averaged in the top 200m, for SODA (blue), CCSM (red), ORCA05 (yellow) and ORCA12 (green) datasets, calculated using the zero contour line of meridional velocity, as proposed by Rodrigues, Rothstein & Winbush (2007).

Table 12: Correlation between annual averages of the bifurcation position averaged in the top 200m (Figure 41).

	SODA		CCSM		ORCA05		ORCA12 ¹	
	R ²	p-value	R ²	p-value	R ²	p-value	R ²	p-value
SODA			7.6%	0.56	6.9%	0.59	-67.3%	<0.01
CCSM	7.6%	0.56			78.7%	<0.01	10.8%	0.66
ORCA05	6.9%	0.59	78.7%	<0.01			5.0%	0.84
ORCA12 ¹	-67.3%	<0.01	10.8%	0.66	5.0%	0.84		

¹Correlations done using the period between 1989 and 2007.

ble 14).

Table 13: Correlation between annual averages of the bifurcation position averaged between 500m and 1000m (Figure 42).

	SODA		CCSM		ORCA05		ORCA12 ¹	
	R ²	p-value	R ²	p-value	R ²	p-value	R ²	p-value
SODA			20.4%	0.11	-2.3%	0.85	26.6%	0.27
CCSM	20.4%	0.11			50%	<0.01	8.6%	0.73
ORCA05	-2.3%	0.85	50%	<0.01			30.3%	0.21
ORCA12 ¹	26.6%	0.27	8.6%	0.73	30.3%	0.21		

¹Correlations done using the period between 1989 and 2007.

The seasonality of the bifurcation position is dominated by the wind stress curl (RODRIGUES; ROTHSTEIN; WINBUSH, 2007). This hypothesis was also tested at interannual timescales, since

Figure 42: Variability of the SEC bifurcation position, averaged between 500m and 1000m, for SODA (blue), CCSM (red), ORCA05 (yellow) and ORCA12 (green) datasets, calculated using the zero contour line of meridional velocity, as proposed by Rodrigues, Rothstein & Winbush (2007).

Table 14: Trends of the SEC bifurcation position (Figures 41 and 42), in $^{\circ}\cdot y^{-1}$.

	SODA	CCSM ($^{\circ}\cdot y^{-1}$)	ORCA05	ORCA12
top 200m	0.02 ± 0.008	0.02 ± 0.005	-0.02 ± 0.006	0.06 ± 0.02
between 500m and 1000m	-0.001 ± 0.006	-0.04 ± 0.005	-0.01 ± 0.005	-0.02 ± 0.03

significant correlations were found mainly between the models forced by CORE, suggesting a strong link with the wind stress curl (Table 12 and Figure 42).

Considering the top 200m (Figure 43), significant and negative correlations, between 10°S and 15°S (the same region observed by those authors) and close to the equator were found in CCSM, ORCA05 and ORCA12 results, with no correspondences to the SODA reanalysis. Also, there are good similarities between the two datasets forced by CORE-II. As both datasets have resemblant bifurcation position time series (Figure 41 and Table 12) it suggests an influence of the Sverdrup dynamics having influence in the position of the bifurcation variability at interannual time scales. However, it does not seem to be the case for SODA, as significative correlations with the wind stress are virtually absent (Figure 43a).

At intermediate depths, there is also a good correlation between CCSM and ORCA05 bifurcation time series (Table 13), suggesting again a link with the forcing. Figure 44. In these models, there is a dipole in the correlations, centered around 20°S , with negative correlations in the southernmost limit of the area. However, these correlations have no correspondences in SODA and ORCA12 models.

Figure 43: Correlation between the annual means of the bifurcation position in the top 200m (calculated using the zero contour line of meridional velocity, as proposed by Rodrigues, Rothstein & Winbush (2007)) and of the wind stress curl for SODA (a), CCSM (b), ORCA05 (c) and ORCA12 (d). Black contours mark the areas where the correlations coefficients are statistically significant at 95%.

3.1.2.2 Western Boundary Currents

Figures 45 and 46, show the transport of NBUC and IWBC, linked to the AMOC, calculated using only northward (positive) meridional velocities at 11°S and 27°S, respectively. Due to the bifurcation position, which is located further north of 27°S the IWBC transport in ORCA05 output was not calculated, since, as discussed in Section 3.1.1, the IWBC transport, in this model is southward.

The time series of the NBUC transport are shown in Figure 45. It is easy to observe that all the models have similar variability. Table 15 presents the correlation between the time series. The results, between the reanalysis and all the models are greater than 30%. Between ORCA05 and CCSM, it is almost 90%. These results suggest a link with wind stress curl and hence, with the Sverdrup balance. Also, from the anomalies (Figure 45b) we found a positive trend that is presented in Table 17.

Figure 46 shows the time series of the meridional transport anomalies of the Intermediate Western Boundary Current. Correlating the time series, no significant similarities, considering the confidence interval of 95%, between them were found (Table 16). Both SODA and CCSM datasets show a positive trend, indicating a strengthening of the volume transport in the 60 years of run, but comparing to the results found for the NBUC (Table 17), the trends found at intermediate depths are only ~25% of those

Figure 44: Correlation between the annual means of the bifurcation position between 500m and 1000m (calculated using the zero contour line of meridional velocity, as proposed by Rodrigues, Rothstein & Winbush (2007)) and of the wind stress curl for SODA (a), CCSM (b), ORCA05 (c) and ORCA12 (d). Black contours mark the areas where the correlations coefficients are statistically significant at 95%.

Table 15: Correlation between annual averages of NBUC transport anomaly at the top 1000m (Figure 45b).

	SODA		CCSM		ORCA05		ORCA12 ¹	
	R ²	p-value	R ²	p-value	R ²	p-value	R ²	p-value
SODA			31.9%	0.01	42.0%	<0.01	47.5%	0.04
CCSM	31.9%	0.01			89.8%	<0.01	29.7%	0.21
ORCA05	42.0%	<0.01	89.8%	<0.01			17.7%	0.47
ORCA12 ¹	47.5%	0.04	29.7%	0.21	17.7%	0.47		

¹Correlations done using the period between 1989 and 2007.

found at 11°S.

As mentioned above, good correlations between the NBUC transports at the surface (specially between CCSM and ORCA05 results) suggest a strong influence of the forcing on the variability. To check this hypothesis, a correlation between the wind stress curl and the transports detrended time series were done. The results are shown in Figure 47. Observing the relationship between the wind stress curl and the NBUC transports, the first thing to be noted is the similarity between CCSM and ORCA05 results.

Figure 45: Volume transport (Sv) time series of NBUC at 11°S, for SODA (blue), CCSM (red), ORCA05 (yellow) and ORCA12 (green) datasets, calculated at the top 1000m as shown on Figure 32. The anomalies were calculated subtracting the seasonal cycle (Figure 33) monthly data, before averages.

Table 16: Correlation between annual averages of IWBC transport anomaly in the top 1000m (Figure 46).

	SODA		CCSM		ORCA12 ¹	
	R ²	p-value	R ²	p-value	R ²	p-value
SODA			20.5%	0.12	-14.2%	0.56
CCSM	20.5%	0.12			4.0%	0.86
ORCA12 ¹	-14.2%	0.56	4.0%	0.86		

¹Correlations done using the period between 1989 and 2007.

These models, which NBUC time series are correlated in almost 90%, present huge and negative correlations between 15°S and 0° followed by positive correlations below 15°S (Figure 47c,e). This feature can also be observed in the SODA reanalysis and also in ORCA12 results, but with smaller areas of significant correlations, which could explain the significant correlations (at least 30%, Table 15) between the reanalysis and all the other models.

Figure 46: Volume transport (Sv) time series of IWBC at 27°S, for SODA (blue), CCSM (red) and ORCA12 (green) datasets, calculated at the top 1000m as shown on Figure 34. The anomalies were calculated subtracting the seasonal cycle from monthly data, before averages.

Table 17: Trends of NBUC and IWBC transports, calculated in the top 1000m (Figures 45 and 46), in $\text{Sv}\cdot\text{y}^{-1}$.

	SODA	CCSM	ORCA05 ($\text{Sv}\cdot\text{y}^{-1}$)	ORCA12
NBUC	0.04 ± 0.01	0.04 ± 0.01	0.06 ± 0.01	0.01 ± 0.05
IWBC	0.02 ± 0.008	0.03 ± 0.004		-0.008 ± 0.03

At intermediate depths the scenario is different: no correlations were found between the models' IWBC transport time series. Although a similar pattern can be observed in the correlations with the wind stress curl (but, with positive correlations between 15°S and the equator, and negative correlations below 15°S), the absence of significant correlations between the outputs, indicates that the wind stress is not a common mechanism for IWBC variability. Also, the absence of results from two models with the same forcing difficults a robust conclusion for CCSM variability, despite of the significant areas of good correlations. The IWBC variability in the SODA reanalysis does not seem to be driven by the wind

stress curl, but by other mechanisms such as buoyancy fluxes.

Figure 47: Correlation between the annual means of the transports of NBUC (left) and IWBC (right) and of the wind stress curl for SODA (a,b), CCSM (c,d), ORCA05 (e,f) and ORCA12 (g,h). Black contours mark the areas where the correlations coefficients are statistically significant at 95%.

As NBUC transports seem to be linked to the wind stress curl, we calculate its correlation with the

Sverdrup interior transport using Equation 2.12. The results, comparing their anomalies are shown in Figure 48.

Figure 48: Anomalies of the NBUC (black) and Sverdrup interior transport (dashed) for (a) SODA, (b) CCSM, (c) ORCA05 and (d) ORCA12.

As expected, due to their common forcing, there is a great similarity between CCSM and ORCA05 Sverdrup interior transports. The correlation between each time series in each model are listed in the Table 18. As the NBUC timeseries for both models have a large correlation (almost 90%), and their

correlations with the Sverdrup interior transports are similar, one can affirm that the variability of the NBUC at interannual timescales is explained mostly by the Sverdrup transport in CCSM and ORCA05 (the correlations are bigger than 50%). The contribution of the wind stress in the variability of the NBUC transports is greater than 20% in SODA and ORCA12 results, although they are not significant at 95%.

Table 18: Correlation between annual averages of NBUC transport anomalies in the top 1000m and the Sverdrup interior transport at 11°S (Figure 48).

SODA		CCSM		ORCA05		ORCA12 ¹	
R ²	p-value	R ²	p-value	R ²	p-value	R ²	p-value
-21.6%	0.09	-61.4%	<0.01	-56.6%	0.01	-26.3%	0.28

¹Correlations done using the period between 1989 and 2007.

From Section 3.1.1, it becomes clear that the bifurcation has an impact in the transport of these currents, particularly the NBUC: when the bifurcation is in its northernmost (southernmost) position, the northward transport is minimum (maximum). In this sense, negative correlations are expected also in interannual time scales. Table 19 presents the correlations at zero lag, using detrended time series, between the SEC bifurcation position and the IWBC/NBUC transports. All the models showed negative results, as expected. At intermediate depths, the correlations were bigger than 60% (SODA and CCSM). In the surface layer, only ORCA05 results had statistically significant correlations. ORCA12 results also showed a good correlation, but, its short time series decreases the degrees of freedom and increases the confidence interval.

ORCA05 and CCSM data show great similarity in previous results (bifurcation position and NBUC transports), what suggests a strong link to their atmosphere forcing (CORE-II, Section 2.1.4). The absence of significant correlation between NBUC and the bifurcation position at the top 200m was not expected. In fact, our results show that the transports (Figure 47c,e) and the bifurcation position (Figure 43b,c) are correlated to the wind stress curl in different regions. For instance, significant negative correlations between the NBUC transports and the wind stress can be found in the east and west limits of the basin, while, positive correlations close to Africa are only found in ORCA05 (Figure 43c), not in CCSM (Figure 43b). Besides, the biggest correlations between the bifurcation position and the wind stress curl in CCSM occur close to the equator, while, for the NBUC transports, the most important correlations are close to the Brazilian coast at 11°S. In ORCA05 results, the biggest correlations are in the west limit of the basin, close to 11°S. So, these results suggest that although the wind stress curl is an important process controlling the bifurcation position at the top 200m and the northward transport of the NBUC, only local wind stress (close to 11°S) controls the direct relationship between the bifurcation position and the northward transports.

Still following this premise of a negative relationship between the bifurcation position and the northward transports, opposite trends for these two parameters are expected, both in surface and intermediate

Table 19: Correlation between annual averages of NBUC/IWBC transport with the bifurcation position averaged in the top 200m/between 500m and 1000m, respectively.

	SODA		CCSM		ORCA05		ORCA12 ¹	
	R ²	p-value	R ²	p-value	R ²	p-value	R ²	p-value
NBUC	-7.7%	0.61	-20.0%	0.12	-51.3%	<0.01	-38.2%	0.11
IWBC	-60.1%	<0.01	-86.7%	<0.01			-7.4%	0.76

¹Correlations done using the period between 1989 and 2007.

depths. Indeed, considering only significant results (when the errors are smaller than the linear trends) in the Tables 14 and 17, only ORCA05 presented opposite trends, while SODA and CCSM do not confirm our hypothesis. At intermediate depths, only CCSM show significant results, confirming our hypothesis.

3.1.2.3 South Atlantic Meridional Overturning Circulation

The NBUC and the IWBC are the main routes of the AMOC in the South Atlantic (SCHMITZ, 1995; BOEBEL et al., 1999; TALLEY, 2003; GANACHAUD, 2003; LUMPKIN; SPEER, 2003; SILVEIRA; CALADO; CASTRO, 2004). The AMOC stream function was calculated for all the model results in the top 5,250m between 0°S and 30°S according to Equation 1.1, Section 1.1. Figure 49 presents the climatological values for each dataset. One should remember that, for ORCA12, the time series is short and only values between 1989 and 2007 were averaged.

Our results for SODA reproduces well those presented by Kröger, Müller & Storch (2012) using the same product. Comparing SODA upper branch of the SAMOC with the model results shows that the CCSM seems to better reproduce the depth of the northward transport, although with less intensity. Besides the differences in the vertical extent and intensity, for all the models examined, the SAMOC maximum transport is deeper (~1200m) in the southernmost limit of the section than close to the Equator (~1000m). The deep branch of the SAMOC is wider in the ORCA05 and ORCA12 simulations than in the other models and in the literature (RENNERMALM et al., 2007; McCLEAN et al., 2011).

Besides the difference in the vertical structure, compared to SODA, the models show closer results to the literature than the SODA reanalysis (Table 20). Rabe, Schott & Köhl (2008), using the GECCO ocean model, calculated an AMOC of 13.5Sv at 10°S. ORCA05 has the smallest AMOC values, as already described by Danabasoglu et al. (2013) and this result agrees with lower NBUC transports (Table 3), inexistent IWBC (Figure 34) and deepest and strongest BC at 27°S (Table 4), linked to a farther north bifurcation at intermediate depths (Table 2).

Figures 50 and 51 show the time series for the vertical maximum of SAMOC at 11°S and 27°S. Except for SODA, the models present similar interannual variability in both latitudes (Figures 50b and

Figure 49: Climatological averages of South Atlantic Meridional Overturning Circulation (SAMOC), between 11°S and 30°S, for SODA (a), CCSM (b), ORCA05 (c) and ORCA12 (d) datasets. Black contours indicate the zero (Sv) stream function. Contours each 2Sv.

Table 20: South Atlantic MOC vertical maximum (Sv) averages at 11°S and 27°S.

	SODA	CCSM	ORCA05	ORCA12
		(Sv·y ⁻¹)		
11°S	19.5±1.3	16.2±1.0	12.0±1.4	12.4±0.9
27°S	20.1±0.9	14.8±0.9	10.3±1.4	11.9±0.7

51b). The correlations between the time series show no correlations between the reanalysis and the other models (Tables 21 and 22).

As in the other results, there are big correlations between the models forced by CORE (greater than 90%) that suggest a link with the forcing. MOC is influenced by the density structure of the Atlantic Ocean, small-scale mixing and buoyancy fluxes (IPCC, 2007). Using CORE-II, these models are under the influence of the same sea-air fluxes (such as momentum, heat and fresh water fluxes, see Sec-

tion 2.1.4), with a direct impact in their AMOC (via buoyancy fluxes, for instance). AMOC is also a result of the meridional velocity integral across the basin, as described in the Section 2.2.3 which, in consequence, accounts the Sverdrup meridional transport, forced by the wind stress curl (Section 2.2.2.1). Moreover, as discussed in the Section 3.1.2.2, the NBOC, one of the main conduits of the AMOC showed to be driven by changes in the Sverdrup balance in CCSM and ORCA05 models. In this sense, big correlations between those models were expected.

Figure 50: SAMOC maximum anomaly time series at 11°S, for SODA (blue), CCSM (red), ORCA05 (yellow) and ORCA12 (green) datasets. The anomalies were calculated subtracting the seasonal cycle.

Table 21: Correlation between annual averages of SAMOC at 11°S (Figure 50b).

	SODA		CCSM		ORCA05		ORCA12 ¹	
	R ²	p-value	R ²	p-value	R ²	p-value	R ²	p-value
SODA			7.2%	0.58	-1.4%	0.92	40.5%	0.09
CCSM	7.2%	0.58			90.0%	<0.01	68.7%	<0.01
ORCA05	-1.4%	0.92	90.0%	<0.01			64.2%	<0.01
ORCA12 ¹	40.5%	0.09	68.7%	<0.01	64.2%	<0.01		

¹Correlations done using the period between 1989 and 2007.

Figure 51: SAMOC maximum anomaly time series at 27°S, for SODA (blue), CCSM (red), ORCA05 (yellow) and ORCA12 (green) datasets. The anomalies were calculated subtracting the seasonal cycle.

Table 22: Correlation between annual averages of SAMOC at 27°S (Figure 50b).

	SODA		CCSM		ORCA05		ORCA12 ¹	
	R ²	p-value	R ²	p-value	R ²	p-value	R ²	p-value
SODA			11.4%	0.39	10.4%	0.43	9.2%	0.70
CCSM	11.4%	0.39			94.0%	<0.01	45.0%	0.05
ORCA05	10.4%	0.43	94.0%	<0.01			45.8%	0.05
ORCA12 ¹	9.2%	0.70	45.0%	0.05	45.8%	0.05		

¹Correlations done using the period between 1989 and 2007.

Figure 52 shows the anomalies of the AMOC, NBUC and Sverdrup transports calculated at 11°S, in a similar way as done by Rabe, Schott & Köhl (2008), in order to illustrate the discussion of the above proposed. The time series were filtered using a 4-year Hamming filter as described in the Section 2.2.3.1. Table 23 shows the correlations between these time series.

Larger (and negative) correlations are also found between NBUC and the Sverdrup transports (Table 23), compared to the previous analysis (Table 18). For CORE forced models, this correlation overreaches 70% in periods longer than 4 years. For the reanalysis and the ORCA12 models, the correlations are 38.5%, but, due to its smaller time series, this correlation was not significant in the high resolution

model.

Analysing Figure 52, it is easy to see how AMOC anomalies are dependent on the NBUC variability, in periods greater than 4 years, specially due to the trends found in both time series (not accounted in the correlation). Positive correlations, as expected, are always significant and, at least equal to 22% in all the models (Table 23). ORCA12 showed the biggest correlations (55%), followed by CCSM and ORCA05. Correlations here, are, however, smaller than those found between the northward current and the Sverdrup transport.

Finally, there is a significant out of phase behaviour between the SODA reanalysis and the models when we account the correlation between AMOC and the Sverdrup transports. As expected, models show small, negative and not significant results, while, the reanalysis goes in the opposite way (Table 23).

The results in Table 23 confirm, that, for CCSM and ORCA05 results, the similarity between their AMOC time series is related to the forcing but not due to the wind stress, but because of the same sea-air fluxes.

The correlations between NBUC and AMOC are smaller than those found by Rabe, Schott & Köhl (2008) (64%), but, they confirm the influence of the western boundary current in the AMOC variability. Zhang et al. (2011), using the GFDL coupled model also showed that AMOC and the northward transports of NBUC, in time scales longer than 10 years are well correlated at zero lag (80%). Our results corroborate with the author's, except for the reanalysis. Using a low-pass filtered at 10 years, the correlations between the time series are 5.4% for SODA, 47% for CCSM and 41.4%. The results for ORCA12 were not considered because of its short time series.

Table 23: Correlation between annual averages of SAMOC (Figure 50b), NBUC (Figure 45b), and Sverdrup interior transports (Figure 48) at 11°S.

	SODA		CCSM		ORCA05		ORCA12 ¹	
	R ²	p-value	R ²	p-value	R ²	p-value	R ²	p-value
NBUC & Sverd.	-38.5%	<0.01	-78.7%	<0.01	-74.7%	<0.01	-38.5%	0.10
NBUC & AMOC	21.7%	<0.01	46.6%	<0.01	33.8%	<0.01	55.0%	0.01
Sverd & AMOC	35.0%	<0.01	-10.7%	0.41	-0.05%	0.69	-13.0%	0.59

¹Correlations done using the period between 1989 and 2007.

Unexpected negative results were found for ORCA05 when correlating the IWBC and AMOC filtered time series at 27°S. ORCA12 results, even with its short time series, showed a positive correlation. For the SODA reanalysis the results were not significant. Negative correlations were unexpected since IWBC is one of the main branches of the AMOC, transporting AAIW, and so, when the northward transport of the IWBC increases, they should also increase in the AMOC.

Positive trends can be found in Figures 50 and 51. In the results of models forced by CORE-II (Section 2.1.4), a well marked inversion (from negative to positive anomalies) is well defined in 1985 (at 11°S, Figure 50b) and 1990 (at 27°S, Figure 51b). The calculated trends are shown in the Table 26.

Figure 52: AMOC, Sverdrup interior and NBUC transport anomalies at 11°S for SODA (a), CCSM (b), ORCA05 (c) and ORCA12 (d) datasets.

An increase of AMOC was observed by Biastoch et al. (2008) using different ocean models forced by CORE at 26.5°N between 1960 and 2000, related to a positive trend of NAO.

Figure 53 shows the correlation between the NAO index, from NOAA, and our results for the AMOC and also for NBUC and IWBC. The NAO index has a positive trend ($4.8 \cdot y^{-1}$), much bigger than those observed in our results for the AMOC. Figure 53 shows the correlation coefficients between

Table 24: Correlation between annual averages of SAMOC (Figure 51b) and IWBC transports (Figure 46) at 27°S.

R^2	SODA		CCSM		ORCA12 ¹	
	p-value	R^2	p-value	R^2	p-value	
IWBC & AMOC	15.9%	0.22	-43.5%	<0.01	59.2%	<0.01

¹Correlations done using the period between 1989 and 2007.**Table 25:** Trends of South Atlantic MOC vertical maximum ($\text{Sv}\cdot\text{y}^{-1}$) at 11°S and 27°S, calculated from annual averaged anomalies (Figures 50b and 51b).

	SODA	CCSM	ORCA05	ORCA12
	($\text{Sv}\cdot\text{y}^{-1}$)			
11°S	0.05 ± 0.01	0.03 ± 0.01	0.05 ± 0.01	0.03 ± 0.04
27°S	$1.1\cdot10^{-3}\pm0.01$	0.04 ± 0.01	0.07 ± 0.01	0.04 ± 0.03

the AMOC and the NAO time series. Only ORCA05 and CCSM show significant correlations, confirming the results of Biastoch et al. (2008).

Table 26: Trends of South Atlantic MOC vertical maximum ($\text{Sv}\cdot\text{y}^{-1}$) at 11°S and 27°S, calculated from annual averaged anomalies (Figures 50b and 51b).

	SODA	CCSM	ORCA05	ORCA12
	($\text{Sv}\cdot\text{y}^{-1}$)			
11°S	0.05 ± 0.01	0.03 ± 0.01	0.05 ± 0.01	0.03 ± 0.04
27°S	$1.1\cdot10^{-3}\pm0.01$	0.04 ± 0.01	0.07 ± 0.01	0.04 ± 0.03

Good correlations were found between the bifurcation position at the surface and the AMOC (Table 27). As the bifurcation position is driven mainly by local changes in the position of the wind stress curl and the AMOC variability is not explained by changes in the wind (Table 23), NBUC seems to be the link between these two series.

Table 27: Correlation between annual averages of SAMOC and bifurcation position.

	SODA		CCSM		ORCA05		ORCA12	
	R^2	p-value	R^2	p-value	R^2	p-value	R^2	p-value
AMOC _{11S} & Bif ₂₀₀	0.9%	0.94	-21.4%	0.04	-37.1%	<0.01	-58.4%	<0.01
AMOC _{27S} & Bif ₅₀₀	-37.9%	<0.01	1.2%	0.93	-12.7%	0.33	-42.3%	0.07

Figure 53: Relationship between NAO index (red and blue), AMOC (black) and NBUC (dashed) at 11°S, for SODA (a), CCSM (b) and ORCA05 (c).

In most of our results in this section SODA was not (unexpectedly) able to reproduce the results found in the literature, mainly the observations. Schott & Brandt (2007), Balmaseda et al. (2007), Köhl & Stammer (2008) report the challenge of faithfully simulating the decadal-to-multidecadal variability of the AMOC using ocean data assimilations based on different schemes. According to the authors, there appears to be no consistent low-frequency variability in AMOC transport variation in available data assimilation systems. Also, according to Kröger, Müller & Storch (2012), when a model assimilates data, its salinity and temperature are not conserved which could compromise AMOC results from reanalysis, such as SODA, and hampers the reproduction of the observations.

3.2 Changes in the water masses linked to AMOC

3.2.1 Validation

Salinity and temperature outputs of the ocean models are compared to the SODA reanalysis. We wish to verify if these models represent well the water mass structure of the South Atlantic.

First of all, a time-longitude averaged temperature-salinity diagram was constructed in the sections shown in Figure 19. The results can be seen in Figure 54. Once again, one should remember that, while SODA, CCSM and ORCA05 results cover the period from 1948 until 2007, the ORCA12 time series is shorter, and starts in 1989.

Figure 54: Temperature-salinity diagram using time-longitude averaged data at 11°S (a) and 27°S (b).

Comparing the curves in Figure 54, it is possible to observe that ORCA12 is the model that best reproduces SODA results, specially at intermediate and deep layers. At 11°S (Figure 54a) their curves are almost identical. Also, CORE forced models show saltier surface layers, with salinities greater than 37. ORCA05 also show bigger temperatures in the upper layer. In both sections, CORE forced models show a more saline AAIW and ORCA05 fails in reproduce the subsurface salinity maximum related to NADW (Figure 54a,b). This can be related to its weak AMOC (Figure 49, Section 3.1.2.3) (DANABASOGLU et al., 2013). These authors also show that the model seems to be unstable, and this feature might be related to numerical problems.

As discussed (STRAMMA; ENGLAND, 1999a), colder and fresher water masses in the upper layer, specially Tropical Water and ISACW, are observed at 27°S (Figure 54b), comparing the subtropical with the subtropical region. Differently from expected, in this section, in ORCA05 and CCSM models, SACW is not represented by a straight line. We can also verify an increase of the salinity minimum re-

lated to AAIW from 27°S (Figure 54a) to 11°S (Figure 54b), as predicted in the literature (STRAMMA; ENGLAND, 1999a; SCHMID; SIEDLER; ZENK, 2000; TOMCZAK; GODFREY, 2003).

Figures 55 to 58 show the climatological sections of salinity and temperature at 11°S (Figures 55 and 56) and 27°S (Figures 56 and 58).

Figure 55: Climatological salinity at 11°S, for SODA (a), CCSM (b), ORCA05 (c) and ORCA12 (d). Black contours indicate the limits of SACW (DONNERS; DRIJFHOUT; HAZELEGER, 2005) and AAIW (TALLEY, 1996).

At 11°S (Figure 55), as described above, CORE-II forced models (Figure 55b,c) show saltier surface waters (top 150m), but in all of them, the salinity maximum is found close to the coast. Waters lighter

than $26.2\text{kg}\cdot\text{m}^{-3}$ occupy the first 250m of the water column. SACW salinity, marked here by the isopycnals $25.5\text{kg}\cdot\text{m}^{-3}$ and $27\text{kg}\cdot\text{m}^{-3}$ varies from 34.5 to 36.5 in SODA and ORCA12 (Figures 55a,d). This range is similar to those found by Tomczak & Godfrey (1994), Stramma & England (1999a). CORE-II forced models low limits are, though, saltier than expected.

The salinity minimum related to the AAIW is detected between 500 and 1000m, as predicted (TOMCZAK; GODFREY, 2003). Moreover, as discussed before (e.g. Figure 54a), CCSM and ORCA05 results produce a saltier AAIW, when compared to the results of SODA, ORCA12, and other studies. According to Talley (1996), close to 11°S , the AAIW typical salinity is 34.4 much lower than the values found in CCSM and ORCA05 models.

The $27.0\text{kg}\cdot\text{m}^{-3}$ and $27.4\text{kg}\cdot\text{m}^{-3}$ isopycnals do not follow the isohalines as seen in the surface layer, particularly CCSM and ORCA05, indicating greater influence of the temperature. Also, for these models, the area occupied by the AAIW is much smaller than in the others. In the ORCA05 results the $27.5\text{kg}\cdot\text{m}^{-3}$ isopycnal does not exceed the 1000m, while in SODA and ORCA12 results it reaches 1200m.

Bellow the AAIW, there is the subsurface salinity maximum that represents the NADW core. In the bottom, below 4000m, fresher waters are found, marking the presence of the AABW. As mentioned before (i.e. Figure 54a), ORCA05 results fail in reproduce this subsurface salinity maximum. Looking just to the salinity section, it is impossible to determine where the limit between the NADW and AABW is in this model (Figure 55c).

Examination of the temperature sections at 11°S (Figure 56) shows that the models are similar to each other compared to the results for salinity. As seen in Figure 54a, ORCA05 results have bigger surface temperatures. In all the results from the models, the ISACW (between 25.5 and $26.2\text{kg}\cdot\text{m}^{-3}$) is found in the thermocline, with temperatures varying from 15°C to 22°C . According to Stramma & England (1999a), temperatures between 5°C and 20°C characterize the SACW in the South Atlantic. Compared to these authors, in this section, SACW (between $25.5\text{kg}\cdot\text{m}^{-3}$ and $27\text{kg}\cdot\text{m}^{-3}$) is warmer (10°C to 22°C).

Bigger differences between the model outputs are found in the AAIW layer. Historical data from Emery & Meincke (1986), show that the AAIW varies from 2°C to 6°C , Reid (1994) consider 3°C as the typical temperature of the AAIW in the SA. CORE-II forced models show warmer AAIW, with temperatures between 5°C and 8°C , while SODA and ORCA12 results are closer to the literature. Below the AAIW, temperature differences between the models are not significative (Figure 56).

At 27°S (Figure 57), sea surface salinity is smaller than at 11°S , since it is situated in a region of negative freshwater flux (precipitation predominance) whilst at 11°S the freshwater flux is positive (evaporation predominance) (TALLEY et al., 2011). Besides, at 27°S , the ISACW is shallower and occupies a wider layer than at 11°S . Its limit with the SASTMW is also deeper ($\sim 300\text{m}$). The SACW salinity ranges between 34.7 and 36.5, similar to the values at 11°S and the those found in the literature (TOMCZAK; GODFREY, 1994; STRAMMA; ENGLAND, 1999a). These values are also close to those observed by Campos, Gonçalves & Ikeda (1995).

Figure 56: Climatological temperature at 11°S, for SODA (a), CCSM (b), ORCA05 (c) and ORCA12 (d). Black contours indicate the limits of SACW (DONNERS; DRIJFHOUT; HAZELEGER, 2005) and AAIW (TALLEY, 1996).

Although the salinity minimum has decreased from 11°S to 27°S (e.g. Tomczak & Godfrey (2003)), as in the northernmost section, CORE-II forced models show a saltier AAIW. In these model results, the salinity minimum is 34.4–34.5 (Figure 57b,c), greater than the values described by Tomczak & Godfrey (2003) and those observed in SODA and ORCA12. The layer occupied by this water mass is wider in the SODA and ORCA12 results (between 600m and 1350m, Figure 57a,d) than in CCSM and ORCA05 results (between 600m and 1200m). In all of them, the AAIW layer is deeper.

In almost all model results one observes the presence of the NADW by its subsurface salinity max-

Figure 57: Climatological salinity at 27°S, for SODA (a), CCSM (b), ORCA05 (c) and ORCA12 (d). Black contours indicate the limits of SACW (DONNERS; DRIJFHOUT; HAZELEGER, 2005) and AAIW (TALLEY, 1996).

imum. In ORCA05, however, as shown in Figure 54b and in Figure 55c, the presence of the NADW in this section is not marked by a salinity maximum.

Bellow the NADW, in SODA and ORCA12 one notes a minimal presence of the AABW, represented by the presence of fresher water bellow 3500m, since the section is too shallow to include such density water.

Once again, the temperature sections are more similar to each other than the salinity ones (Figure 58). The temperature in the top 100m also decreased comparing to values found at 11°S. SACW

(between 25.5 and $26.2\text{kg}\cdot\text{m}^{-3}$) temperature ranges between 7°C and 20°C , similar to the results of Campos, Gonçalves & Ikeda (1995).

At intermediate layers, the water column is warmer for the CORE-II forced models (Figure 58b,c). Although in all the models the temperature varies between 3°C and 6.5°C , which coincides with previous studies (e.g. Emery & Meincke (1986), Reid (1994)) in these models, the layer between $27\text{kg}\cdot\text{m}^{-3}$ and $27.5\text{kg}\cdot\text{m}^{-3}$ is dominated by temperatures greater than 5°C .

Figure 58: Climatological temperature at 27°S , for SODA (a), CCSM (b), ORCA05 (c) and ORCA12 (d). Black contours indicate the limits of SACW (DONNERS; DRI-JFHOUT; HAZELEGER, 2005) and AAIW (TALLEY, 1996).

To quantify the differences between all the model results used in this study and discussed in the

climatological sections of salinity and temperature, and also assess some differences in these models variabilities, a Taylor's Diagram was elaborated. The results are shown in Figure 59.

Figure 59: Taylor's Diagram for salinity and temperature data of CCSM, ORCA05 and ORCA12 at 11°S (a) and 27°S (b), using SODA as reference.

In both sections, temperature fields showed good correlations with SODA (greater than 99%, Figure 59a,b), and standard deviations close to the reference. Compared to the SODA reanalysis, CORE-II forced models diverge more from SODA than ORCA12. In fact, at 11°S (Figure 59a), the great similarity between SODA and ORCA12, already seen in their climatological results (specially their salinity-temperature diagram, Figure 54a), is also observed in their variability, since, both salinity and temperature fields have normalised standard deviations close to 1, correlations higher than 98% and RSM smaller than 0.25.

Big differences are seen in the salinity fields. At 11°S, ORCA05 shows the smallest correlation (97%) and the biggest standard deviation (difference of ~7%). This behaviour is probably related to its fresher NADW, which is evident in the Figure 54a. At 27°S, as discussed before, the analysis of the climatological results, particularly Figure 54b, highlights that CCSM has a different salinity structure compared to SODA (lower salinities in SACW layers, saltier AAIW and reduced presence of the NADW). This can reflect in its variability resulting in lower correlation and higher differences between the standard deviations (~22%) as observed in the Figure 54b.

Taylor's Diagrams reveal that, besides their particularities and differences, all model results show correlations greater than 90% and reasonable standard deviations.

3.2.2 Water masses variability

In this work, the interannual variability of the water masses carried by the western boundary currents that are part of the AMOC return flow is analyzed. Figures 60 and 61 show, by overlaying velocity contours on density fields, that the SACW and the AAIW are the main water masses carried by the NBUC and the IWBC.

At 11°S, all the model results analyzed show that the core of the current carries the SACW. In CORE-II forced models, the NBUC carries a mixture of the ISACW and the SASTMW. But, in the SODA reanalysis and in ORCA12, the western boundary current carries the SASTMW (Figure 60).

Figure 60: Velocity field ($\text{m}\cdot\text{s}^{-1}$, black contours) superimposed on the density section ($\text{kg}\cdot\text{m}^{-3}$) at 11°S, for SODA (a), CCSM (b), ORCA05 (c) and ORCA12 (d). Dashed contours indicate negative flow. Contours at every $0.02\text{m}\cdot\text{s}^{-1}$. Light gray contours indicate the limits of SACW (DONNERS; DRIJFHOUT; HAZELEGER, 2005) and AAIW (TALLEY, 1996). Fields are an average (between 1948 and 2007 for SODA, CCSM and ORCA05 and between 1989 and 2007).

At 27°S, however, the IWBC transports only the AAIW (Figure 61), while the SACW is carried by the BC.

Figure 61: Velocity field ($\text{m}\cdot\text{s}^{-1}$, black contours) superimposed on the density section ($\text{kg}\cdot\text{m}^{-3}$) at 27°S, for SODA (a), CCSM (b), ORCA05 (c) and ORCA12 (d). Dashed contours indicate negative flow. Contours at every $0.02\text{m}\cdot\text{s}^{-1}$. Light gray contours indicate the limits of SACW (DONNERS; DRIJFHOUT; HAZELEGER, 2005) and AAIW (TALLEY, 1996). Fields are an average (between 1948 and 2007 for SODA, CCSM and ORCA05 and between 1989 and 2007).

This result confirms previous studies that discuss the importance of the central and intermediate waters as indicators of changes in the AMOC (e.g. Garzoli & Matano (2011)). OMP analysis was performed in order to determine the vertical position of the SACW and the AAIW cores at 11°S and 27°S as described in Section 2.2.6.

From Figure 54, the indices for the water masses types used in the OMP analysis were calculated, as described in Section 2.2.6. The results are shown in Tables 28 and 29. Our indices were compared to those found in the published literature, summarised in Table 30.

The temperature index for the ISACW, in all the model results, but mainly for ORCA05, exceeds the maximum range described by Mémery et al. (2000), using WOCE data close to the Brazilian coast. As described in Section 3.2.2, ORCA05 results show higher temperatures in the upper levels. CORE-II forced models have a saltier AAIW, compared to the other results (Figure 54, Section 3.2.2). The similarity of ORCA12 and SODA results is enhanced by the similarity of their salinity and temperature

Table 28: Source water type parameters for each model at 11°S. Parameters: mass conservation, temperature (θ_0 , in °C), salinity and potential vorticity (Π , in $10^{-8}(\text{m}\cdot\text{s})^{-1}$).

	Parameter	SACW			AAIW	UCDW	Weight
		ISACW	SAMW ¹	SAMW ²			
SODA	Mass conservation.			1			3
	θ_0	21.6	12.8	8.2	3.8	3.7	2
	Salinity	36.5	35.2	34.7	34.3	34.9	.1
	Π	0.03	$2\cdot 10^{-3}$	$4\cdot 10^{-3}$	$2\cdot 10^{-3}$	$8\cdot 10^{-4}$	.1
CCSM	θ_0	22.1	13.0	8.9	5.0	4.0	2
	Salinity	36.7	35.3	34.8	34.5	35.0	.1
	Π	0.03	$2\cdot 10^{-3}$	$5\cdot 10^{-3}$	$2\cdot 10^{-3}$	$9\cdot 10^{-4}$	.1
ORCA05	θ_0	22.9	13.2	8.2	6.5	3.3	2
	Salinity	37.0	35.3	34.9	34.6	34.9	.1
	Π	0.03	$3\cdot 10^{-3}$	$6\cdot 10^{-3}$	$3\cdot 10^{-3}$	$4\cdot 10^{-4}$	.1
ORCA12	θ_0	21.6	12.6	8.1	4.0	4.1	2
	Salinity	36.5	35.2	34.7	34.3	35.0	.1
	Π	0.04	$2\cdot 10^{-3}$	$4\cdot 10^{-3}$	$2\cdot 10^{-3}$	$7\cdot 10^{-4}$	.1

¹Related to the isopycnal $26.6\text{kg}\cdot\text{m}^{-3}$. ²Related to the isopycnal $27\text{kg}\cdot\text{m}^{-3}$.

Table 29: Source water type parameters for each model at 27°S. Parameters: mass conservation, temperature (θ_0 , in °C), salinity and potential vorticity (Π , in $10^{-8}(\text{m}\cdot\text{s})^{-1}$).

	Parameter	SACW			AAIW	UCDW	Weight
		ISACW	SAMW ¹	SAMW ²			
SODA	Mass conservation.			1			3
	θ_0	21.2	13.0	7.8	2.5	3.3	2
	Salinity	36.4	35.3	34.6	34.1	34.9	.1
	Π	0.06	$5\cdot 10^{-3}$	0.01	$6\cdot 10^{-3}$	$2\cdot 10^{-3}$	.1
CCSM	θ_0	20.6	11.5	8.1	3.5	3.6	2
	Salinity	36.2	35.1	34.9	34.4	34.9	.1
	Π	0.05	0.02	$1\cdot 10^{-3}$	$6\cdot 10^{-3}$	$2\cdot 10^{-3}$	.1
ORCA05	θ_0	21.7	15.3	12.9	3.5	3.2	2
	Salinity	36.6	35.5	35.2	34.3	34.9	.1
	Π	0.04	0.02	$1\cdot 10^{-3}$	$7\cdot 10^{-3}$	$1\cdot 10^{-3}$	.1
ORCA12	θ_0	21.2	15.0	12.4	3.0	3.9	2
	Salinity	36.4	35.4	35.1	34.2	34.9	.1
	Π	0.02	0.02	$9\cdot 10^{-4}$	$6\cdot 10^{-3}$	$1\cdot 10^{-3}$	.1

¹Related to the isopycnal $26.6\text{kg}\cdot\text{m}^{-3}$. ²Related to the isopycnal $27\text{kg}\cdot\text{m}^{-3}$.

index, mainly at 11°S, and for the upper layers. As noted in Figure S4, the UCDW and also the NADW

from the results of ORCA12 are warmer than in the results from SODA.

Few studies describes the potential vorticity distribution in the South Atlantic, specially related to the upper water masses used in this work. According to McCartney (1982), a characteristic of the Mode Water is its homogeneity, associated to a minimal vertical gradient of potential density and thus, potential vorticity. Mode waters obtained from the results for all model used here showed small potential vorticity values. Keffer (1985) presents maps of observed potential vorticity in the South Atlantic for the thermocline layers which correspond to the SASTMW, SAMW, and AAIW (summarised in Table 31). Our calculated values agree well with his results.

Table 30: Temperature and salinity characteristics of the source water masses according to the literature.

	Temperature (°C)	Salinity	Density ($\text{kg}\cdot\text{m}^{-3}$)	References
ISACW	16–21	35.5–36.4	25.5–26.2	Donners, Drijfhout & Hazeleger (2005) Mémery et al. (2000)
SAMW	4–15	34.2–35.8	26.5–27.1	McCartney (1977) Hanawa & Talley (2001)
AAIW	2–6	33.8–34.8	27.1–27.4	Talley (1996) Emery (2003)
UCDW	3.4–4.0	34.6–34.9	27.55–27.75	Tsuchiya, Talley & McCartney (1994) Hogg & Thurnherr (2005)

Table 31: Potential vorticity characteristics of the source water masses according to the literature.

	(II, in $10^{-8}(\text{m}\cdot\text{s})^{-1}$)		References
	11°S	27°S	
SASTMW	$1.5\cdot 10^{-3}$ –0.02		Keffer (1985)
SAMW	$5\cdot 10^{-3}$ –0.01	0.01–0.02	
AAIW	$2\cdot 10^{-3}$ – $3\cdot 10^{-3}$	$7\cdot 10^{-3}$ – $5\cdot 10^{-3}$	

From Figure 54, the salinity and the temperature in the isopycnals related to the SACW ($25.5\text{kg}\cdot\text{m}^{-3}$ and $27.0\text{kg}\cdot\text{m}^{-3}$) and the AAIW (calculated using the mixed diagram, Section 2.2.6) were used to calculate the neutral densities, as described in the Section 2.2.7. Table 32 shows the neutral isopycnal calculated for each model and section.

Differences between the neutral densities for the SACW are small when compared to the potential densities because the influence of pressure is smaller. According to McDougall (1987), to calculate γ_n , reference level is slightly adjusted at each point, because of nonlinearity of the equation of state. Our values agree well with those found by Koltermann, Gouretski & Jancke (2011), shown in the Figure 23, Section 2.2.7.

Table 32: Neutral isopycnal (γ_n , $\text{kg}\cdot\text{m}^{-3}$) calculated in each model for each water masses, using the methodology of Jackett & McDougall (1997) and Serazin (2011), described in the Section 2.2.7.

	Section	SACW		AAIW
		$\gamma_{n25.5}$	$\gamma_{n27.0}$	
SODA	11°S	25.52	27.11	27.42
	27°S	25.53	27.11	27.42
CCSM	11°S	25.52	27.10	27.42
	27°S	25.51	27.10	27.54
ORCA05	11°S	25.52	27.09	27.29
	27°S	25.52	27.10	27.46
ORCA12	11°S	25.52	27.11	27.40
	27°S	25.52	27.11	27.44

Section 3.2.2.1 shows the OMP results and changes in the salinity and temperature in the isopycnals related to the SACW, while Section 3.2.2.2 shows the same results for the AAIW.

3.2.2.1 South Atlantic Central Water

Figures 62 to 65 show the result of the OMP analysis for the SACW at 11°S and 27°S where a combination of the ISACW and the SAMW indices (Tables 28 and 29) was used to separate the SACW from the AAIW.

Figure 62 is a climatological distribution of the SACW at 11°S. The first 150m were ignored in this analysis due to its proximity to the mixed layer. In all models its possible to note the presence of the SACW core in the top 500m (percentages greater than 80%) coinciding with the isopycnal of $27.0\text{kg}\cdot\text{m}^{-3}$ (Figures 55 and 56). In the next 100–200m, there is a mixed layer where the partial contribution of the SACW decreases and the presence of the AAIW increases. This result agrees with Stramma & England (1999a) and Silveira, Calado & Castro (2004) that describes this interface between 500–700m, between 27.05 and $27.1\text{kg}\cdot\text{m}^{-3}$. The OMP analysis reveals that the SACW core is deeper close to the Brazilian coast, where the NBUC core is located (Figure 60).

The results at 27°S are similar. The $27\text{kg}\cdot\text{m}^{-3}$ isopycnal, between 600m and 700m, is the bottom limit of the SACW core. Differently from the results of the other models, that have the first 500m completely filled by SACW (100%), SODA results present an average of 90–95% of SACW in the first 500m. Examination of its climatological temperature-salinity diagram (Figure 54b), shows a colder SACW, specially in the layer of the SASTMW (26.2 – $26.6\text{kg}\cdot\text{m}^{-3}$). This can be the cause of the unconformity between SODA reanalysis and the model results, as the index of this water mass is not included in the OMP analysis.

The mass balance constraint reinforces the good performance of the reanalysis. At 11°S, the error associated with the OMP is smaller than 2% in almost all the water column analyzed (Figure 63).

Figure 62: Climatological section of fractional contribution from SACW at 11°S for SODA (a), CCSM (b), ORCA05 (c) and ORCA12 (d). White contour marks the limit of the water mass core (80%). Figures (a), (b) and (c) are an average of the results from 1948 to 2007, while the average done in figure (d) comprises the results from 1989 to 2007.

At 27°S, the error is smaller than 1.5% for all models. According to Budillon et al. (2003), Tomczak (1981), Tomczak & Large (1989), Tomczak & Poole (1999) and Frants et al. (2013), the mass balance constraint is used as a measure of the uncertainty of the result and mass balance residuals of 5% or less are considered acceptable.

During the empirical tests to decide the best combination of weights for the OMP analysis, it was observed that if the importance of the potential vorticity was increased, the worse (better) the results for the SACW (AAIW), i.e., bigger (smaller) mass balance constraints. Bigger weights for temperature also changed the water masses distributions, reducing (increasing) the error in the SACW (AAIW) layer. Changes in the weight of the salinity increases the error in all the water column.

Sensitive tests reveal that, in general, raising (decreasing) the temperature of the lSACW in only 5%, yields the SACW upper limit position deeper (shallower), and decreases (increases) the error at the upper levels. In the same way, elevating (reducing) the temperature of the SAMW in the same 5% moves the core (the 80% limit) of the SACW to upper (deeper) layers with little differences in the final

Figure 63: Mass conservation residual fit (%) for the OMP analysis at 11°S for SODA (a), CCSM (b), ORCA05 (c) and ORCA12 (d).

mass constraint. 5% of change in the temperature of the AAIW does not move significantly the position of the SACW core.

Changes in the AAIW salinity (also 5%), however, have significant impact in the position of the SACW core, since the salinity minimum is the main characteristic of this water mass (e.g Talley (1996), Pickard, Emery & Talley (2007)). The mass constraint, for the AAIW layer, can be two times greater. If the same changes are applied to the salinity of the LSACW there is no impact on the position of the SACW core. However, the same changes applied to the salinity of the SAMW restrict the SACW to the upper 250m or less. The error in this case can exceed 10%.

Potential vorticity changes do not generate large changes in the water mass distribution or in the error of the analysis.

When comparing all the model results, ORCA12 proved to be the most sensitive to the tests with known changes. For this model, and an increase or decrease of 5% in AAIW salinity parameter results in a SACW occupying the top 1000m. In this case, the mass constraint surpasses 15% for the AAIW layer (between $27\text{kg}\cdot\text{m}^{-3}$ and $27.5\text{kg}\cdot\text{cm}^{-3}$).

Figure 66 shows the anomalies of the SACW core (80% contour) position. The anomalies were constructed by subtracting the mean position of each longitude (Figures 62 and 64). Some interannual to interdecadal variability can be observed, and changes affect all the section most of the time. In

Figure 64: Climatological section of fractional contribution from SACW at 27°S for SODA (a), CCSM (b), ORCA05 (c) and ORCA12 (d). White contour marks the limit of the water mass core (80%). Figures (a), (b) and (c) are an average of the results from 1948 to 2007, while the average done in figure (d) comprises the results from 1989 to 2007.

the SODA reanalysis the anomalies are larger (Figure 66a,e). Higher frequencies are found in SODA and ORCA12 results. In both sections, the similarity between the results of CORE-II forced models is evident, specially between 1948 and 1995 (Figure 66b,c,f,g). Excluding the trend, their correlations is greater than 80% in both sections (Tables 33 and 34), confirming the influence of the forcing.

At 11°S (Figures 66a–d and 67) it is possible to observe the deepening of the SACW core in the SODA reanalysis (mainly between 30°S and 34°S, Figure 66a) and in CCSM. Their positive trends are shown in Table 35. At the same section, in the results for ORCA05 and ORCA12, the SACW core is getting shallower, (positive trends, Table 35).

Besides their similar variability, CORE-II forced models show opposite trends. This was not expected, since in all the previous analysis, despite differences in the mean value, their behavior was always very similar (e.g. Section 3.1.2).

Differently from the section at 11°S, all the models show a positive trend of the SACW 80% contour, which means, a shallowing of the water mass core (Figures 66e–h and 67b). These trends are shown in

Figure 65: Mass conservation residual fit (%) for the OMP analysis at 27°S for SODA (a), CCSM (b), ORCA05 (c) and ORCA12 (d).

Table 33: Correlations of the SACW core vertical position from each model, averaged between 37°S and 30°S (Figure 67a).

	SODA		CCSM		ORCA05		ORCA12 ¹	
	R ²	p-value	R ²	p-value	R ²	p-value	R ²	p-value
SODA			2.0%	0.88	12.3%	0.35	−43.4%	0.06
CCSM	2.0%	0.88			83.5%	<0.01	1.1%	0.96
ORCA05	12.3%	0.35	83.5%	<0.01			13.4%	0.58
ORCA12 ¹	−43.4%	0.06	1.1%	0.96	13.4%	0.58		

¹Correlations done using the period between 1989 and 2007.

the Table 35, and they are larger than at 11°S. CCSM results have the smallest trends, linked to the deepening in the end of the 20th century, close to the Brazilian coast. ORCA12 results show larger trends, 3 times more than the ones on the reanalysis.

Excluding the trend, significant and positive correlations were found between the results of the CCSM, ORCA12 models and the SODA reanalysis. ORCA05 and ORCA12 results had also significant correlations (Table 34).

Changes in the position of the water masses core is linked to changes in their salinity and tempera-

Figure 66: Hovmöller of anomalies of SACW core depth, for SODA (a,e), CCSM (b,f), ORCA05 (c,g) and ORCA12 (d,h), at 11°S (top) and 27°S (bottom). Anomalies were constructed by subtracting the mean position of the 80% contour for each longitude (Figures 62 and 64).

ture. So, changes in these properties projected onto the neutral surfaces of the main isopycnal of SACW were analyzed and are presented below.

Figures 68, 70, 72 and 74 show the Hovmöller Diagrams for the salinity and the temperature projected in the $\gamma_{n_{25.5}} \text{ kg} \cdot \text{m}^{-3}$ and $\gamma_{n_{27}} \text{ kg} \cdot \text{m}^{-3}$, the neutral surfaces related to the isopycnal considered the end-points of the SACW (DONNERS; DRIJFHOUT; HAZELEGER, 2005). The anomalies were constructed by subtracting from each longitude, its mean salinity/temperature values.

As in the Hovmöller Diagrams construct for the anomalies of the SACW position, the SODA re-

Figure 67: Time series of the anomalies of SACW depth core, at 11°S (a) and at 27°S (b) for SODA, CCSM, ORCA05 and ORCA12 (d).

Table 34: Correlations of the SACW core vertical position from each model, averaged between 48°S and 40°S (Figure 67b).

	SODA		CCSM		ORCA05		ORCA12 ¹	
	R ²	p-value	R ²	p-value	R ²	p-value	R ²	p-value
SODA			26.7%	0.04	15.9%	0.23	44.8%	0.04
CCSM	26.7%	0.04			88.3%	<0.01	29.0%	0.22
ORCA05	15.9%	0.23	88.3%	<0.01			46.2%	0.05
ORCA12 ¹	44.8%	0.04	29.0%	0.22	46.2%	0.05		

¹Correlations done using the period between 1989 and 2007.

Table 35: Trends of the position of the SACW core (80% contour) ($\text{m}\cdot\text{y}^{-1}$) at 11°S and 27°S, calculated using anomalies (Figure 67).

	SODA	CCSM	ORCA05	ORCA12
	($\text{m}\cdot\text{y}^{-1}$)			
11°S	-0.38 ± 0.06	-0.17 ± 0.04	0.16 ± 0.04	0.29 ± 0.17
27°S	0.71 ± 0.11	0.17 ± 0.09	0.48 ± 0.06	2.30 ± 0.39

analysis show greater variability compared to the other models, in both sections and neutral surfaces (Figures 68a,e, 70a,e, 72a,e and 74a,e).

Figure 68: Hovmoller of anomalies of the salinity projected in the $\gamma_{n_{25.5}} \text{ kg} \cdot \text{m}^{-3}$ neutral surface for SODA (a,e), CCSM (b,f), ORCA05 (c,g) and ORCA12 (d,h), at 11°S (top) and 27°S (bottom).

At 11°S a sequence of positive anomalies is replaced by a period of decrease in the salinity and in the temperature between 1985-2000, followed by a period of positive anomalies at the end of the time series (Figures 68a, and 70a). Instead of this period of negative anomalies, the trend for the salinity projected in the neutral surface $\gamma_{n_{25.5}}$ is positive (Table 38). The same variability in the end of the 20th century and the beginning of the 21st century are also observed in the ORCA12 results (Figures 68d,

and 70d). However, when the trends are not considered, SODA and ORCA12 results show negative (but not significant) correlations (Table 40).

Figure 69: Time series of the salinity anomalies projected in the $\gamma_{n25.5}$ neutral surface, related to SACW, at 11°S and 27°S for SODA, CCSM, ORCA05 and ORCA12.

Once again, CORE-II forced models show similar results. Despite the bigger anomalies in ORCA05, the correlation between their salinity time series projected onto the neutral surface related to the isopycnal $25.5\text{kg}\cdot\text{m}^{-3}$ ($\gamma_{n25.5}$) is superior than 80% (Table 40). Besides the interannual variability (Figure 69a), there is a clear decrease in the salinity (Table 38), with correspondences to the temperature (Figures 70b,c and 71a).

Table 36: Correlation between annual averages the salinity projected at the neutral surface of $\gamma_{n25.5}$ related to the SACW, averaged between 37.5°W and 30°W, at 11°S (Figure 69a).

	SODA		CCSM		ORCA05		ORCA12 ¹	
	R ²	p-value	R ²	p-value	R ²	p-value	R ²	p-value
SODA			3.8%	0.77	11.3%	0.39	-33.6%	0.16
CCSM	3.8%	0.77			83.3%	<0.01	-38.6%	0.10
ORCA05	11.3%	0.39	83.3%	<0.01			-31.7%	0.18
ORCA12 ¹	-33.6%	0.16	-38.6%	0.10	-31.7%	0.18		

¹Correlations done using the period between 1989 and 2007.

Figure 70: Hovmöller of the anomalies of the temperature projected in the $\gamma_{n_{25.5}}$ neutral surface, related to SACW, at 11°S (top) and 27°S (bottom) for SODA (a,e), CCSM (b,f), ORCA05 (c,g) and ORCA12 (d,h).

At 27°S, trends are positive in the results of SODA and ORCA12, while CORE-II forced models show an opposite trend (Table 38).

With the climate changing and the increase of sea surface temperatures, the hydrological cycle is expected to be intensified (IPCC, 2007; McCARTHY et al., 2012). Therefore there is an increase in the salinity in the water exposed to high evaporation regions (CURRY; DICKSON; YASHAYAIEV, 2003). In this scenario, negative trends found in CORE-II forced models were not expected, and do not agree with previous studies. Curry, Dickson & Yashayaev (2003), using observed salinities between late 1950s

Figure 71: Time series of the temperature anomalies projected in the $\gamma_{n25.5}$ neutral surface, related to SACW, at 11°S and 27°S for SODA, CCSM, ORCA05 and ORCA12.

Table 37: Correlation between annual averages the salinity projected at the neutral surface of $\gamma_{n25.5}$ related to the SACW, averaged between 48.5°W and 40°W, at 27°S (Figure 69b).

	SODA		CCSM		ORCA05		ORCA12 ¹	
	R ²	p-value	R ²	p-value	R ²	p-value	R ²	p-value
SODA			4.6%	0.73	-11.9%	0.36	10.3%	0.67
CCSM	4.6%	0.73			62.6%	<0.01	38.2%	0.11
ORCA05	-11.9%	0.36	62.6%	<0.01			60.4%	<0.01
ORCA12 ¹	10.3%	0.67	38.2%	0.11	60.4%	<0.01		

¹Correlations done using the period between 1989 and 2007.

and 1990s, show that, between 25°S and 35°N, there was an increase of the salinity (up to 0.5) in the neutral density of $\gamma_n = 25.5 \text{ kg} \cdot \text{m}^{-3}$. A warming in this neutral surface, in this region, was also observed (up to 2°C). These trends agree with our results for SODA and ORCA12 (Table 38).

In the low latitude Atlantic, the factors building positive salinity anomalies must also involve some combination of dynamic and thermo-dynamic processes: altered circulation, precipitation patterns and intensified trade winds (MOULIN et al., 1997), enhanced evaporation rates due to warming of the surface ocean (McCARTHY et al., 2012).

At 27°S, there is a sudden change in salinity anomalies in the beginning of the 1960's (negative to

positive anomalies) and in the beginning of the 1980s (positive to negative anomalies). This period of strong positive anomalies (of around 20 years) could coincide with the negative phase of AMO (KNIGHT; FOLLAND; SCAIFE, 2006) and agrees with the observations of Arz, Pätzold & Wefer (1999) and Chiessi et al. (2009). According to these authors, during negative phases of AMO, there is a reduction of the AMOC, which leads to an increase of South Atlantic temperature. Our time series do show a shift that could be associated with the AMO but are nonetheless too short for further speculations.

Table 38: Trends of the salinity projected in the neutral surface ($\gamma_{n_{25.5}}$) related to the SACW at 11°S and 27°S (Figure 69).

	SODA	CCSM	ORCA05	ORCA12
	(y ⁻¹)			
11°S	0.002±0.004	-0.017±0.002	-0.018±0.003	0.06±0.0.007
27°S	0.007±0.002	-0.005±0.003	-0.010±0.003	0.05±0.0.005

From the sensitive test, it was observed that changes in the STW salinity and temperature would impact the position of the water mass core. Thus, once again, excluding the trends, the time series of the SACW core (80% contour, Figure 67) and the salinity projected onto $\gamma_{n_{25.5}}$ were correlated. The results are presented in the Table 39.

The results show that, in general, an increase of the salinity of the LSACW leads to a deepening of the SACW core. However, this relationship was only significant in the SODA reanalysis, where the correlations are higher than 30% (Table 39).

Table 39: Correlation between annual averages the salinity projected at the neutral surface of $\gamma_{n_{25.5}}$ (Figure 69) and the SACW core position (Figure 67), at 11°S and 27°S.

	SODA		CCSM		ORCA05		ORCA12 ¹	
	R ²	p-value	R ²	p-value	R ²	p-value	R ²	p-value
11°S	-31.7%	0.01	-3.7%	0.77	-14.1%	0.28	14.1%	0.56
27°S	-32.5%	0.01	-18.2%	0.16	-0.1%	0.93	5.3%	0.82

¹Correlations done using the period between 1989 and 2007.

The same kind of the analysis was done for the SAMW, its lower limit was used to identify the SACW in the OMP analysis (Section 2.2.6). Figures 72 to 75 show the results for the salinity and temperature projected onto the neutral surface related to the isopycnal 27kg·m⁻³ ($\gamma_{n_{27.0}}$).

Comparing all the Hovmöller diagrams it is observed that the results for SODA show higher variability, with more intense anomalies, and an increase in salinity and temperature at $\gamma_{n_{27.0}}$ is evident in both sections (Figures 72a,e and 74a,e). Trends are shown in Tables 40 and 39.

Figure 72: Hovmöller of anomalies of the salinity projected in the $\gamma_{n_{27.0}} \text{ kg} \cdot \text{m}^{-3}$ neutral surface for SODA (a,e), CCSM (b,f), ORCA05 (c,g) and ORCA12 (d,h), at 11°S (top) and 27°S (bottom).

CORE-II forced models show similar results. The correlation between their time series (Figure 73) is greater than 50% (Table 39). In both sections, there are well marked periods of positive and negative anomalies.

For these models, salinity and temperature anomalies seem to be out of phase, with negative anomalies happening between 1970 and 1990 at 11°S and positive anomalies at 27°S for the same period (Figures 73 and 75).

Garabato et al. (2009), using observations in the Drake Passage, between 1969 and 2005, show

similar oscillations in temperature and salinity time series of the SAMW, compared to our results at 27°S. The authors observed a salinity and temperature increase on density surfaces during the 1970s, with a possible peak during the 1980s (since there was no observations during this period). This trend then reverses in the 1990s. In the end of the time series, (2005), they registered the coldest and freshest SAMW. This period of positive anomalies in the salinity and in the temperature coincides with the warm phase of the Interdecadal Pacific Oscillation (IPO), calculated by Parker et al. (2007). As shown by Folland et al. (2002) and Parker et al. (2007), among others, a decadal-scale increase (decrease) in the IPO index is associated with surface warming (cooling) around the SAMW formation area of the southeast Pacific-Drake Passage and with a northeastward (southeastward) migration of the South Pacific Convergence Zone that is consistent with a positive (negative) anomaly in the winter-mean evaporation in the region.

Figure 73: Time series of the salinity anomalies projected in the $\gamma_{n_{27.0}}$ neutral surface, related to SACW, at 11°S and 27°S for SODA, CCSM, ORCA05 and ORCA12.

Differences in the SACW between 11°S and 27°S were also observed by Curry, Dickson & Yashayaev (2003). According to these authors, in the South Atlantic, salinity in this isopycnal has been observed to increase 0.05–0.1 (psu) at 11°S, while it has decreased between 0.03–0.05 at 27°S. Observations were done between 1955 and 1999. Positive trends in the salinity of the SAMW on the $\gamma_{n_{27.0}}$ found in the SODA result at 27°S disagrees with Curry, Dickson & Yashayaev (2003).

Correlations between the salinity projected onto the $\gamma_{n_{27.0}}$ $\text{kg}\cdot\text{m}^{-3}$ neutral surface (Figure 73) and

Figure 74: Hovmöller of the anomalies of the temperature projected in the $\gamma_{n_{27.0}}$ neutral surface, related to SACW, at 11°S (top) and 27°S (bottom) for SODA (a,e), CCSM (b,f), ORCA05 (c,g) and ORCA12 (d,h).

the position of the SACW core (80% contour, Figure 67) are shown below. Results show that the variability of the SACW position is linked to variations in the SAMW ($\gamma_{n_{27.0}}$), particularly at 27°S.

3.2.2.2 Antarctic Intermediate Water

Figures 76 and 78 show the results of the second OMP analysis, separating the core of the AAIW. The indices in this analysis are listed in Tables 28 and 29. At 11°S the similarity between the ORCA12 and SODA results is reflected in their salinity, temperature and potential vorticity of the source water pa-

Figure 75: Time series of the temperature anomalies projected in the $\gamma_{n27.0}$ neutral surface, related to SACW, at 11°S and 27°S for SODA, CCSM, ORCA05 and ORCA12.

Table 40: Correlation between annual averages the salinity projected at the neutral surface of $\gamma_{n27.0}$ related to the SACW, averaged between 37.5°W and 30°W, at 11°S (Figure 73a).

	SODA		CCSM		ORCA05		ORCA12 ¹	
	R ²	p-value	R ²	p-value	R ²	p-value	R ²	p-value
SODA			-18.1%	0.17	-17.8%	0.17	6.5%	0.79
CCSM	18.1%	0.17			54.8%	<0.01	55.8%	0.10
ORCA05	-17.8%	0.17	54.8%	<0.01			27.6%	0.25
ORCA12 ¹	6.5%	0.79	55.8%	0.10	27.6%	0.25		

¹Correlations done using the period between 1989 and 2007.

rameters, even though all the model results are similar. The major differences are found in the indices for the UCDW. The absence of the salinity maximum related to the NADW in the ORCA05 model results does not interfere with the salinity index of the UCDW. However, ORCA05 has the coldest UCDW and half of the potential vorticity compared to the other model results. At 27°S the indices are not so uniform, but are in general in agreement with what is found in the literature (i.e. Tables 30 and 31). It is important to note that the same parameters were used for AAIW in both OMP analysis.

Even though there are deficiencies in representing the SACW (using only one SWT, as described in Section 2.2.6), the methodology applied in this work allowed us to properly isolate the core of the

Table 41: Correlation between annual averages the salinity projected at the neutral surface of $\gamma_{n27.0}$ related to the SACW, averaged between 48.5°W and 40°W, at 27°S (Figure 73b).

	SODA		CCSM		ORCA05		ORCA12 ¹	
	R ²	p-value	R ²	p-value	R ²	p-value	R ²	p-value
SODA			-2.2%	0.86	-24.7%	0.06	-41.8%	0.07
CCSM	-2.2%	0.86			54.1%	<0.01	63.2%	<0.01
ORCA05	-24.7%	0.06	54.1%	<0.01			58.1%	0.01
ORCA12 ¹	-41.8%	0.07	63.2%	0.01	58.1%	0.01		

¹Correlations done using the period between 1989 and 2007.**Table 42:** Trends of the salinity projected in the neutral surface ($\gamma_{n27.0}$) related to the SACW at 11°S and 27°S (Figure 69).

	SODA	CCSM	ORCA05	ORCA12
			(y ⁻¹)	
11°S	0.005±0.002	1·10 ⁻⁴ ±0.002	-0.001±3·10 ⁻⁴	0.003±4·10 ⁻⁴
27°S	0.004±0.002	-1·10 ⁻⁴ ±0.003	-0.003±4·10 ⁻⁴	0.004±0.001

Table 43: Correlation between annual averages the salinity projected at the neutral surface of $\gamma_{n27.0}$ (Figure 69) and the SACW core position (Figure 67), at 11°S and 27°S.

	SODA		CCSM		ORCA05		ORCA12 ¹	
	R ²	p-value	R ²	p-value	R ²	p-value	R ²	p-value
11°S	3.5%	0.79	27.1%	0.03	-38.5%	<0.01	19.3%	0.43
27°S	27.2%	0.03	77.0%	<0.01	60.1%	<0.01	18.4%	0.45

¹Correlations done using the period between 1989 and 2007.

AAIW, in both sections. Obviously, these limitations had consequences. Comparing the first (Figures 62 and 64) and the second (Figures 76 and 78) OMP analysis it is easy to verify the differences in the resulting layers of mixing between SACW and AAIW. In the second OMP due to the index of a lighter SAMW (26.6kg·m⁻³—versus the end-point 27.0kg·m⁻³ in the first analysis), there is a larger mixed layer in the results for all the models, with a larger presence of the AAIW, mainly, between 500m and 700m at 11°S (Figure 76) and between 600m and 800m (Figure 78) at 27°S.

At 11°S (Figure 76) the analysis show a good representation of the expected position of the AAIW core, compared to the climatological sections of salinity, temperature and density (e.g. Figures 55, 56 and 60), in which at 11°S the AAIW lies between 550m and 1200m. The 27kg·m⁻³ and 27.55kg·m⁻³ isopycnals coincide here with the line of 50% of water mass contribution, which agrees with the results of Stramma & England (1999a), Silveira, Calado & Castro (2004), Hogg & Thurnherr (2005) that consider these isopycnals as transition limits. As in the results for the SACW (Figure 62), the water

mass concentrates close to the Brazilian coast (greater percentages and also wider layer), where one can find the core of the IWC (Figure 60). The 80% contour follows the density contours, being deeper close to the South America coast (while the opposite happens with SACW layer).

Figure 76: Climatological section of fractional contribution from AAIW at 11°S for SODA (a), CCSM (b), ORCA05 (c) and ORCA12 (d). White contour marks the limit of the water mass core (80%). Figures (a), (b) and (c) are an average of the results from 1948 to 2007, while the average done in figure (d) comprises the results from 1989 to 2007.

The mass conservation residual fit plots (Figure 77) are consistent with the limitations of the methodology applied. The error for the SACW layer increases toward the lower densities, and are much bigger than those found in the previous OMP analysis for this section (Figure 63). ORCA12 results are an exception, since larger errors are related to the UCDW (Figure 77d). These results show the good performance of the OMP analysis, since the mass balance constraint is within the acceptable limits (BUDILLON et al., 2003; TOMCZAK, 1981; TOMCZAK; LARGE, 1989; TOMCZAK; POOLE, 1999; FRANTS et al., 2013).

At 27°S (Figure 78) the AAIW core occupies the layer between 700m and 1200m, being deeper in the CCSM model results (Figure 78b) and shallower in the results of ORCA05 (Figure 78c). Compared to the other models, the AAIW core is narrower in the SODA reanalysis, specially in the east part of

Figure 77: Mass conservation residual fit (%) for the second OMP analysis at 11°S for SODA (a), CCSM (b), ORCA05 (c) and ORCA12 (d).

the section. The 50% contour agrees well with the isopycnals that indicate the upper and lower limits of the AAIW (STRAMMA; ENGLAND, 1999a; SILVEIRA; CALADO; CASTRO, 2004; HOGG; THURNHERR, 2005), which can be seen in Figure 61.

The mass balance constraint is presented in Figure 79. In this section, the error is maximum in the transition layer, between AAIW and SACW, as consequence of the simplified representation of the second water mass. The errors do not exceed the 5%. The exception is the CCSM model results (maximum mass conservation residual fit: 12%, increasing toward lighter layers, similar to Figure 77). From Figure 54, it is possible to note that the SAMW is less saline while the AAIW of this model is saltier which can affect the AAIW distribution in the OMP analysis.

The sensitivity of the indices was tested again. As in the previous OMP analysis, changing in 5% the temperature of the SACW causes small alterations in the position of the AAIW core: it gets shallower (deeper) with higher (lower) temperatures, and the error increases. The results are also influenced slightly by changes in the AAIW temperature. A warmer SWT in this case results in a wider AAIW core, although within this layer, the percentage of AAIW hardly reaches 90%. Changes in the UCDW temperature have little effects in the AAIW core, but a warmer SWT can reduce (increase) the mixing layer between these water masses.

Meanwhile, salinity plays a fundamental role in the success of the OMP analysis. Lower SAMW

Figure 78: Climatological section of fractional contribution from AAIW at 27°S for SODA (a), CCSM (b), ORCA05 (c) and ORCA12 (d). White contour marks the limit of the water mass core (80%). Figures (a), (b) and (c) are an average of the results from 1948 to 2007, while the average done in figure (d) comprises the results from 1989 to 2007.

salinities (5%) decreases the mixed layer between the SACW and the AAIW, but the mass distribution of the intermediate water mass in its core does not exceeds 70%. Saltier SAMW increases the AAIW core (area where the percentage of this water mass is greater than 80%), but it spreads to the SACW layers. In both cases, the error does not overpass 5%. Changes in the AAIW salinity (in both directions) have important implications for the analysis. The OMP analysis can not properly separate the water mass, percentages do not exceed 40% and the error can be greater than 30%. Changing the salinity parameter in the deep water masses increases de mixing between the AAIW and the UCDW. Decreasing 5% the salinity of the UCDW deepens the AAIW core until 2000m at 11°S, which does not agree with observational studies (e.g. Talley (1996)).

Figure 80 shows the results of the vertical extension of the AAIW core at 11°S and 27°S, calculated as the vertical distance between the 80% contours (Section 2.2.6). In all the model results, the vertical extension of the AAIW core ranges between 200 and 460m. SODA results have wider AAIW cores compared to the other models results, mainly at 11°S (as observed in the Figure 76a) and before 1960

Figure 79: Mass conservation residual fit (%) for the second OMP analysis at 27°S for SODA (a), CCSM (b), ORCA05 (c) and ORCA12 (d).

that coincides with the period of a small number of observations to be assimilated in the reanalysis. It is possible to observe a eastward decrease of the vertical extension of the AAIW core at 11°S (Figure 80a–d) and at 27°S (Figure 80e,h). CCSM and ORCA05 results at 11°S show a wider AAIW core between 1970 and 1990 (Figure 80b,c).

Figure 81 shows the anomalies of the vertical extension of the AAIW core. SODA and ORCA12 results are marked by an interannual variability (Figure 81a,d,e,h) while CORE-II forced models show variabilities with longer periods (Figure 81b,c,f,g). At 11°S, between the late 1960s and the early 1970s (Figure 81b,c) there is a clear shift in the anomalies (negative to positive). Another shift happens between the end of 1980 and the beginning of 1990 (positive to negative anomalies).

Observing the longitude-averaged time series, it is easy to note the similarity between CCSM and ORCA05 results (Figure 82). At 11°S the correlations are larger than 70% and at 27°S, 48% (Tables 44 and 45). At 11°S, ORCA12 results show negative correlations with the SODA reanalysis and also positive correlations with CCSM model results (Table 44).

Observing the trends (Table 46), only ORCA05 results show an increase in the AAIW layer at 11°S. SODA results show the biggest trend, and the AAIW decreases more than 120m during the 60 years analyzed. Larger trends can be found between 1960 and 1965, when positive anomalies are replaced by negative ones (Figure 81a). At 27°S, model results show a different scenario, with an increase in the

Figure 80: Hovmöller of the vertical extension of AAIW core for SODA (a,e), CCSM (b,f), ORCA05 (c,g) and ORCA12 (d,h), at 11°S (top) and 27°S (bottom). The vertical extension was constructed the depth position of the top and the bottom 80% contour.

AAIW layer.

The sensitive tests showed that the results of the OMP analysis are related mainly to salinity changes. Figure 83 shows the results for the salinity anomalies calculated in the $\gamma_{n_{AAIW}}$ neutral surface (Table 32). Results show that SODA presents a more frequent variability with larger amplitudes in both sections (Figure 83a,d). As in other results (e.g. Figure 68), CORE-II forced models show variabilities with longer periods (~ 20 years), with salinity minimum between 1970 and 1990 at 11°S (Figure 83b,c). Their similarity is expressed in the large correlation of their longitude-averaged time series, shown in

Figure 81: Hovmoller of the anomalies of the vertical extension of AAIW core for SODA (a,e), CCSM (b,f), ORCA05 (c,g) and ORCA12 (d,h), at 11°S (top) and 27°S (bottom). The anomalies were constructed by subtracting the mean vertical extension from each year of result.

Figure 84 (almost 78%, Table 47), excluding the trend, suggesting a connection with the forcing. At 27°S their correlations were smaller ($\sim 40\%$, Table 48). It is possible to note negative correlations between CCSM and SODA results (Tables 47 and 48).

As expected, changes in the temperature projected in the $\gamma_{n_{AAIW}}$ neutral surfaces follow the modifications in the salinity. Observing the results shown in Figures 83 to 85, summarized in Table 49, model results show an increase in the AAIW salinity and temperature at 11°S (except for ORCA05 results) while the opposite is observed at 27°S (except for CCSM). Schmidtko & Johnson (2012) also

Figure 82: Time series of the anomalies of the vertical extension of the AAIW core, at 11°S (a) and at 27°S (b) for SODA, CCSM, ORCA05 and ORCA12 (d).

Table 44: Correlations between annual averages of the vertical extension of the SACW core from each model, averaged between 37°S and 30°S (Figure 82a).

	SODA		CCSM		ORCA05		ORCA12 ¹	
	R ²	p-value	R ²	p-value	R ²	p-value	R ²	p-value
SODA			-18.9%	0.15	9.1%	0.49	-43.4%	0.06
CCSM	-18.9%	0.15			74.2%	<0.01	58.9%	0.01
ORCA05	9.1%	0.49	74.2%	<0.01			35.8%	0.13
ORCA12 ¹	-39.9%	0.09	58.9%	0.01	35.8%	0.13		

¹Correlations done using the period between 1989 and 2007.

observed this same dipole when analyzed the salinity minimum trend from ARGO data. This inversion of salinity and temperature trends in neutral surfaces linked to the AAIW was observed by Curry, Dickson & Yashayaev (2003), using observational data collected close to the Brazilian coast, between 1955 and 1999.

The AAIW ($\gamma_{n_{AAIW}}$) and the SAMW ($\gamma_{n_{27.0}}$) show similar oscillation in results of the CORE-II forced models at 11°S (Figures 72b,c, 74, 83b,c and 85b,c). As discussed before, Folland et al. (2002) and Parker et al. (2007), this positive salinity/temperature anomalies coincide with the period of the positive phase of IPO, between 1970 and 1990, since it leads a surface warming in the SAMW area. Garabato

Table 45: Correlation between annual averages of the vertical extension of the AAIW core at 11°S, averaged between 48°S and 40°S (Figure 82b).

	SODA		CCSM		ORCA05		ORCA12 ¹	
	R ²	p-value	R ²	p-value	R ²	p-value	R ²	p-value
SODA			13.8%	0.30	7.1%	0.59	34.0%	0.15
CCSM	13.8%	0.30			48.0%	<0.01	34.3%	0.15
ORCA05	7.1%	0.59	48.0%	<0.01			43.1%	0.07
ORCA12 ¹	34.0%	0.15	34.3%	0.15	43.1%	0.07		

¹Correlations done using the period between 1989 and 2007.**Table 46:** Trends of the vertical extension of the AAIW core at 11°S and 27°S, (Figure 82).

	SODA	CCSM	ORCA05	ORCA12
	(m·y ⁻¹)			
11°S	-2.2±0.25	-0.53±0.19	0.18±0.08	-0.77±0.45
27°S	0.7±0.22	-1.4±0.2	0.61±0.16	3.1±0.65

Table 47: Correlation between annual averages the salinity projected at the neutral surface of γ_{nAAIW} , averaged between 37.5°W and 30°W, at 11°S (Figure 84a).

	SODA		CCSM		ORCA05		ORCA12 ¹	
	R ²	p-value	R ²	p-value	R ²	p-value	R ²	p-value
SODA			-30.2%	0.02	-12.8%	0.33	-25.1%	0.30
CCSM	-30.2%	0.02			77.8%	<0.01	25.0%	0.30
ORCA05	-12.8%	0.33	77.8%	<0.01			-20.3%	0.41
ORCA12 ¹	-25.1%	0.30	25.0%	0.30	-20.3%	0.41		

¹Correlations done using the period between 1989 and 2007.**Table 48:** Correlation between annual averages the salinity projected at the neutral surface of γ_{nAAIW} , averaged between 48.5°W and 40°W, at 27°S (Figure 84b).

	SODA		CCSM		ORCA05		ORCA12 ¹	
	R ²	p-value	R ²	p-value	R ²	p-value	R ²	p-value
SODA			-43.8%	<0.01	-10.7%	0.42	20.5%	0.40
CCSM	-43.8%	<0.01			40.6%	<0.01	19.7%	0.42
ORCA05	-10.7%	0.42	40.6%	<0.01			51.9%	0.02
ORCA12 ¹	20.5%	0.40	19.7%	0.42	51.9%	0.02		

et al. (2009), using observational data collected at the Drake Passage only observed the effect of IPO oscillation in SAMW neutral surfaces. However, this oscillation seems to be negative correlated to changes

Figure 83: Hovmöller of anomalies of the salinity projected in the $\gamma_{n_{AAIW}} \text{ kg} \cdot \text{m}^{-3}$ neutral surface for SODA (a,e), CCSM (b,f), ORCA05 (c,g) and ORCA12 (d,h), at 11°S (top) and 27°S (bottom).

in the AAIW, mainly in CORE-II forced model results. Even though part of the AAIW is formed in the southeast Pacific Ocean, as a by-product of the SAMW, formed in a deep convective mixed layer, our results seems to be out of phase with those found by these authors, since during the positive phase of the IPO there is a decrease in the AAIW and the SAMW salinities and temperatures (Figures 72, 72, 83 and 85). These results are not robust since our time series are not long enough to analyze a significant number of whole cycles.

Correlations between the annual averages of the salinity projected at the neutral surface of $\gamma_{n_{AAIW}}$

Figure 84: Time series of the salinity anomalies projected in the $\gamma_{n_{AAIW}}$ neutral surface, at 11°S and 27°S for SODA, CCSM, ORCA05 and ORCA12.

Table 49: Trends of the salinity projected in the neutral surface ($\gamma_{n_{AAIW}}$) at 11°S and 27°S (Figure 84).

	SODA	CCSM	ORCA05	ORCA12
	(y ⁻¹)			
11°S	$7 \cdot 10^{-4} \pm 8 \cdot 10^{-5}$	$1 \cdot 10^{-4} \pm 3 \cdot 10^{-5}$	$-2 \cdot 10^{-4} \pm 9 \cdot 10^{-5}$	$2 \cdot 10^{-4} \pm 1 \cdot 10^{-4}$
27°S	$-2 \cdot 10^{-4} \pm 6 \cdot 10^{-5}$	$3 \cdot 10^{-4} \pm 4 \cdot 10^{-5}$	$-3 \cdot 10^{-5} \pm 8 \cdot 10^{-5}$	$-0.002 \pm 2 \cdot 10^{-4}$

(Figure 84b) and changes in the vertical extension of the AAIW core (Figure 82) at 11°S and 27°S show that the AAIW layer is strongly related to changes in its salinity. High and negative correlations presented in Table 50 show that an increase (decrease) in the salinity in the AAIW layer results in a narrower vertical extension of the water mass.

Figure 85: Hovmöller of anomalies of the temperature projected in the $\gamma_{n_{AAIW}} \text{ kg} \cdot \text{m}^{-3}$ neutral surface for SODA (a,e), CCSM (b,f), ORCA05 (c,g) and ORCA12 (d,h), at 11°S (top) and 27°S (bottom).

Table 50: Correlation between the annual averages of the salinity projected at the neutral surface of $\gamma_{n_{AAIW}}$ (Figure 84b) and changes in the vertical extension of the AAIW core (Figure 82) at 11°S and 27°S.

	SODA		CCSM		ORCA05		ORCA12 ¹	
	R ²	p-value	R ²	p-value	R ²	p-value	R ²	p-value
11°S	-82.6%	<0.01	-83.4%	<0.01	-61.5%	<0.01	-88.5%	<0.01
27°S	-68.9%	<0.01	-64.9%	<0.01	-89.3%	<0.01	-69.5%	<0.01

¹Correlations done using the period between 1989 and 2007.

4 Conclusion

In this work, the role of the South Atlantic in the AMOC variability was studied and changes in its currents and water masses were addressed through the analysis of different model results (different origins and resolutions) and an ocean reanalysis result. As described in the Section 2.1, SODA is an ocean model (POP), forced by the 20Crv2 winds (COMPO et al., 2011), with 0.5° of horizontal resolution, which assimilates observational data (CARTON et al., 2000a; CARTON et al., 2000b; CARTON; GIESE, 2006; CARTON; GIESE, 2008). CCSM here referred to as the results obtained with the Ocean Model Component (POP) of the CCSM (Coupled Climate System Model) which is coupled to a sea-ice model (CICE) forced by CORE-II data, with 1° of horizontal resolution (COLLINS et al., 2006). ORCA05 and ORCA12 are also results from an ocean model (OPA) coupled to a sea-ice model (LIM) (BERNARD et al., 2006; MADEC, 2008; VANCOPPENOLLE et al., 2009). ORCA05 has 0.5° of horizontal resolution and is also forced by CORE-II data while ORCA12 is a fully eddy-resolving simulation forced by the reanalysis ERA-Interim.

The variability of the SEC bifurcation position, with respect to interannual to multidecadal time scales, is a new result with great relative importance in the context of climate change, due to its link with the NBUC/IWBC transports which feed directly into the AMOC. A number of significant results were found while connecting the variability of the SEC bifurcation with AMOC changes.

We were able to reproduce the results of Rodrigues, Rothstein & Winbush (2007): all models analyzed could reproduce the SEC bifurcation position and their seasonal variability in the upper 200m, which is linked to changes in the NBUC/BC system and are driven by changes in the wind stress curl.

However, only the SODA reanalysis could reproduce the seasonal cycle of the SEC bifurcation position observed by Rodrigues, Rothstein & Winbush (2007) in the first 1000m. The cause of the differences is still an open question, probably related to the different numerics and deep layer spin-up time scales.

A significant result of this study is the link of the upper limit of the Subtropical Gyre at intermediate depths to a weaker AMOC, observed in the ORCA05 results which is also discussed by Danabasoglu et al. (2013). This feature affects the results related to the interannual variability of the AMOC and its link to the western boundary currents and the SEC bifurcation position, since the latitude chosen to analyze the variability of the IWBC was not suitable for this model.

For periods longer than a year, CCSM and ORCA05 show very similar bifurcation position time series, even though ORCA05 results fails to reproduce the southernmost extension of the bifurcation position. This strongly suggests that the atmospheric forcing is dominating the processes linked to the variability of the SEC position. In fact, as in the seasonal cycle, the SEC bifurcation is influenced by

changes in the wind stress curl and the results are similar for almost all models. For the SODA reanalysis, however, the correlations with the wind stress curl were minimal and the mechanism that drives the bifurcation variability in this case is not well understood.

Correlations between the CORE-II forced models also suggest a link between the NBUC transports and the forcing. When correlated to the wind stress curl, the results of all models show, similar behaviour. The results comparing the NBUC transport with the Sverdrup interior transport, show that, even in the SODA reanalysis, part of the variability of the NBUC can be explained by the wind stress curl variability, via Sverdrup balance.

The SEC seasonal cycle described by Schott et al. (2005), Rodrigues, Rothstein & Winbush (2007), Silva et al. (2009), shows that the northernmost position of the bifurcation is related to the smallest northward volume transports. Our results show that this behaviour exists at interannual timescales too. Negative correlations were found in all datasets, but they are only significant for ORCA05 results, even though good correlations with the wind were observed in both series, especially in the CORE-II forced models.

Examination of the AMOC series calculated in this work (as the vertical maximum of the zonally integrated meridional velocities), yields similarity between the results for ORCA and CCSM. However, little of the Sverdrup transports variability can explain AMOC variance among the model results, which disagree with Rabe, Schott & Köhl (2008), Zhang et al. (2011) results. SODA reanalysis results, on the other hand, show positive and significant correlations, even though at 11°S the direction of the AMOC transport and the Sverdrup transport are opposites.

Good correlations were found between the SEC bifurcation at the surface and the AMOC. As the bifurcation position is driven mainly by local changes in the position of the wind stress curl and the AMOC variability is not explained by changes in the wind, the NBUC seems to be the link between them.

Observing the trends, all the results from the models (with exception of ORCA05) show an equatorward movement of the bifurcation position, which is in agreement with recent studies showing a scenario of expansion (40%) of the subtropical gyre (GARZOLI; MATANO, 2011).

Still considering the seasonal cycle, the northward displacement of the SEC bifurcation position could lead to a decrease in the NBUC transports. Our results do not confirm this idea: positive trends indicating an increase of the northward transports of the western boundary current were found for all the models results considered. This is related to the positive trends of the AMOC, linked to the increase in the NAO index, as found by Biastoch et al. (2008).

Good correlations, albeit not as high as at the surface, were also observed between the time series of the bifurcation position, averaged between 500m and 1500m, which also suggest a link with the forcing. The correlation maps of the SEC bifurcation position with the wind stress curl show similar values for both CCSM and ORCA05 results, with larger correlations in the latter, that can be related to its weak AMOC.

As there was no IWBC- related northward transport calculated for ORCA05, since the SEC bifur-

cation happens north of 27°S (and thus, the IWBC in this section is southward, as part of the BC), there is no evidence of clear link between the IWBC and the wind in CORE-II forced models. The results in SODA, CCSM and ORCA12 show the correlations displayed in a dipole pattern, with positive (negative) correlations north (south) of around 15°S.

Unlike what occurs at the surface, at intermediate depths there is a negative correlation between the SEC bifurcation variability and the transport of the IWBC. This agrees with the observations of Rodrigues, Rothstein & Winbush (2007) at seasonal time scales, describing that when the bifurcation is in its southernmost (northernmost) position, the IWBC northward transport is maximum (minimum).

The relationship between the IWBC and the AMOC could not be well defined within our results. The results from CCSM show a negative correlation, while ORCA12 showed a strong and positive result. The SODA reanalysis does not yield a significant correlation. There is no robust conclusion for the direct link between the IWBC and the AMOC. A negative correlation between the SEC bifurcation and the AMOC variability, was observed in the SODA and ORCA12 results. Correlations for the CORE-II forced models were not significant.

When we look at the long term trend, the results are more robust. Negative trends for the SEC bifurcation position and positive trends for the IWBC and the AMOC transports are found which is consistent with the seasonal cycle dynamics described earlier. Positive trends in the transports are related to the increase of the NAO index.

At intermediate depths the results are not conclusive. The IWBC current was not well developed or in the case of ORCA05 was absent.

Schott & Brandt (2007), Balmaseda et al. (2007), Köhl & Stammer (2008) report the challenge of faithfully simulating the decadal-to-multidecadal variability of the AMOC using ocean data assimilations based on different schemes. According to the authors, there appears to be no consistent low-frequency variability in the AMOC transport variation in available data assimilation systems which is consistent with the lack of the AMOC variability in the SODA reanalysis. Furthermore, according to Kröger, Müller & Storch (2012), when a model assimilates data, its salinity and temperature are not conserved.

In the second part of this work the interannual variability of the water masses carried by the western boundary currents, which are part of the AMOC return flow is analyzed.

The analysis of the vertical sections and the salinity-temperature diagrams show that the results of CORE-II forced models present saltier water masses in the upper layers (Tropical Water, SACW and AAIW). ORCA05 results show a shallower AAIW. These model results also fails in reproduce the subsurface salinity maximum related to the NADW. Both results agree with the weak AMOC presented by this model.

These variations can also be observed on Taylor's Diagram. The results of the CORE-II forced models show some differences in their variability compared to the reanalysis SODA, mainly in the salinity. Despite the variations, all model results analyzed show correlations greater than 90% and reasonable standard deviations.

OMP analysis proved to be an effective way of separating the water masses. The use of two different water mass types to represent the SACW was indeed the best way to separate this water mass, confirming the results of Klein & Tomczak (1994) and Poole & Tomczak (1999). However, due to limitations in the number of different available data, it was impossible to maintain two different SWT to represent the SACW in order to separate the AAIW, and only one SWT was used. Despite this restriction, the OMP analysis resulted in a well positioned core, when compared to the literature (e.g. Talley (1996), Tomczak & Godfrey (2003), Pickard, Emery & Talley (2007)). In both cases, the good performance of the OMP analysis is enhanced by their good mass constraint results (smaller than 5%, as suggested by Tomczak (1981), Tomczak & Large (1989), Tomczak (1999), Budillon et al. (2003) and Frants et al. (2013)).

For both water masses, the right definition of the salinity index proved to be an important feature for the success of the OMP analysis. In the ocean, the range of temperatures is larger than the range of salinities, which makes little changes in the latter more relevant to the definition of the water mass core.

In general, the time series of the SACW and AAIW cores of the reanalysis SODA present bigger anomalies. In SODA and ORCA12 results, an interannual variability can be observed. For the results of CORE-II forced models variabilities at longer time scales (inter to multidecadal) show to be an important feature. This pattern can be also observed in the salinity and in the temperature projected in the neutral surface layer related to this water masses.

The salinity and the temperature projected in the neutral surface related to the ISACW ($\gamma_{n_{25.5}} \text{ kg} \cdot \text{m}^{-3}$) in CCSM and ORCA05 results, mainly at 27°S, could be linked to the AMO, since the period of strong positive anomalies (between 1960 and 1980) coincides with the negative phase of AMO (KNIGHT; FOLLAND; SCAIFE, 2006). According to Arz, Pätzold & Wefer (1999) and Chiessi et al. (2009), during AMO negative phase there is a reducing of AMOC, which leads an increase of South Atlantic temperature.

In the SAMW layer ($\gamma_{n_{27.0}} \text{ kg} \cdot \text{m}^{-3}$), salinity and temperature could be linked to the IPO. Our results at 27°S agree with the observations of Garabato et al. (2009) done at the Drake Passage. Folland et al. (2002) and Parker et al. (2007) show that the IPO index is associated with surface warming (cooling) around the SAMW formation area of the southeast Pacific-Drake Passage and with a northeastward (southeastward) migration of the South Pacific Convergence Zone that is consistent with a positive (negative) anomaly in the winter-mean evaporation in the region. Results at 11°S, both for the SAMW and for the AAIW ($\gamma_{n_{AAIW}} \text{ kg} \cdot \text{m}^{-3}$) show to be out of phase, and, during the period of positive phase of IPO, there is a decrease in the salinity and in the temperature projected in the surfaces related to this water mass.

Changes in the position of the SACW core seems to be related to changes in the ISACW ($\gamma_{n_{25.5}} \text{ kg} \cdot \text{m}^{-3}$): an increase of the salinity on this neutral surface leads in a deepening of the SACW core. However, this negative correlation was only significative in the SODA results. Positive correlations were found between SACW core position time series and the salinity projected in the neutral surface related to the SAMW: an increase of the salinity in this neutral surface leads in a shallowing of the SACW core. This

relationship was significant in SODA, CCSM and ORCA05 results.

The vertical extension of the AAIW core seems to be strongly related to the changes in the salinity of this water mass: negative correlations, always larger than 60%, show that a decrease in the salinity increases the vertical extension of the AAIW core in both sections.

Trend analysis show an increase of the the LSACW salinity in the results of SODA and ORCA12 which agrees with the observations of Curry, Dickson & Yashayaev (2003) for this region, while CORE-II forced models show opposite results. This same pattern can be noted in the SAMW salinity trends. While Curry, Dickson & Yashayaev (2003) observed positive trends north of 20°S, which agrees with our results for SODA and ORCA12, they also found negative trends south of it, with agrees with CORE-II forced models results.

The relationship between the trend of the SACW core position and the salinity projected in the neutral surfaces is a complicate scenario. At 11°S, SODA and CCSM show negative trends, showing a deepening of the SACW core, while ORCA05 and ORCA12 show positive trends. At 27°S, all the models show a shallowing of the SACW core. From the correlations obtained (excluding the trend), we expected that, in case of positive (negative) correlations between the salinity in a given neutral surface and the vertical position of the SACW core, similar trends, both negative or positive (opposite trends) would be found. However, considering only the significantly statistical results it is not possible to determine which trend is more important to the long term changes in the SACW core position.

Finally, AAIW results confirm the observations of Curry, Dickson & Yashayaev (2003) and Schmidtko & Johnson (2012): positive trends north of 20°S while a decrease in the AAIW salinity is observed south of it in the SA. Trends of the vertical extension of the AAIW core show that this water mass is occupying less space in the water column due to an increasing of the salinity in the neutral surfaces at 11°S, while the opposite happens at 27°S.

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