Matheus Vasconcellos Cortezi

**Observations of the South Atlantic
Subtropical Mode Water using PIES data**

Dissertation presented to Instituto Oceanográfico da Universidade de São Paulo, as part of the requirements for the obtention of the title as Master of Science, in the Oceanography program, concentration area of Physical Oceanography.

Corrected Version

Advisor:

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Universidade de São Paulo
Instituto Oceanográfico

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Judged in / / 2017 by

Prof. Olga Tiemi Sato, PhD.	Concept
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*To my father Alci. Thanks
for always trusting me.*

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“[...] Beyond the darkness into light

Is a warrior breaking free

Did you think that I would not fight?

I had no doubt, I count on me

Beyond my pain beyond my fear

For no one will I bend a knee

My heart is fire, my mind is clear

My spirit is a roaring sea.”

– Zen Pencils

Abstract

Subtropical mode water is a voluminous body of water in the ocean whose main feature is the homogeneity in both vertical structure and horizontal extension. The subtropical mode water (STMW) of the southwest Atlantic is formed between the months of July and October near the Brazil-Malvinas confluence and along the Brazil Current recirculation gyre. The formation region extends on the order of 3000 km zonally, from 20°W to 50°W, and 1000 km meridionally, from 30°S to 40°S, and it is typically about 170 m thick. *In situ* data from pressure-equipped inverted echo sounders (PIES) installed in the western portion of the basin, along 34.5°S, are available from 2009 to the present. These data after processed and calibrated can provide an unprecedented description of the STMW involving processes since its formation at the surface until the final stage of its residence in the interior of the ocean. Temperature and salinity data estimated by the PIES are based on empirical look-up tables that relate the acoustic travel time with the baroclinic structure of the ocean. This technique is known as the Graves Empirical Mode (GEM), and here it is used to detect profiles containing homogeneous segments of temperature and salinity that characterize the mode water. The GEM method was seasonally corrected to reconstruct surface variability necessary for STMW formation. The interannual covariance between STMW layer thickness and the Brazil Current was calculated, but no significant correlation at that timescale was observed. The mode water layer detected was about $220 \text{ m} \pm 55 \text{ m}$ thick on all sites, agreeing with previous studies.

Key words: Subtropical mode water, South Atlantic, PIES, inverted echo sounders, Graves Empirical Mode, GEM, Brazil Current

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List of Abbreviations

- **AAIW**: Antarctic Intermediate Water
- **BC**: Brazil Current
- **CPIES**: Current equipped PIES
- **EDW**: Eighteen Degree Water
- **EKE**: Eddy Kinetic Energy
- **GEM**: Gravest Empirical Mode
- **HYCOM**: Hybrid Coordinate Ocean Model
- **ISAS**: In Situ Analysis System
- **NAC**: North Atlantic Current
- **NADP**: North Atlantic Deep Water
- **NASTMW**: North Atlantic Subtropical Mode Water
- **NOAA**: National Oceanic and Atmospheric Administration
- **NPSTMW**: North Pacific Subtropical Mode Water
- **PIES**: Pressure equipped Inverted Echo Sounders
- **PV**: Potential Vorticity
- **SAM**: Southwest Atlantic Meridional Overturning Circulation
- **SAMW**: Subantarctic Mode Water

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- **SASTMW**: South Atlantic Subtropical Mode Water
 - **SPMW**: North Atlantic Subpolar Mode Water
 - **STMW**: Subtropical Mode Water
 - **UCPW**: Upper Circumpolar Water
 - **WOA**: World Ocean Atlas
 - **WOCE**: World Ocean Circulation Experiment
 - **WOD**: World Ocean Database

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1. Introduction

1.1 Definition of Mode Waters

Mode waters are large homogeneous volumes of water found within a geographical area. This homogeneity is characterized by a *picnostad*, a layer marked with very low vertical density gradients. These are usually caused by deep vertical convection, promoted by the air-sea interaction processes during the months of winter and early spring. Mode waters are distinguished by low potential vorticity (PV) values (McCartney 1982). The process of mode water formation can be related to the formation of a regular mixed layer. In both cases the loss of buoyancy during winter and early spring causes convective mixing processes (Provost et al. 1999; Hanawa and Talley 2001). After that, the restratification of the upper layer creates a seasonal thermocline that isolates mode waters from surface and they are transported by local advection away from their formation site. This process is what will allow those thick mixed layers to be effectively called mode waters (Hanawa and Talley 2001). Once submerged, mode waters are often found between the permanent and seasonal thermocline (Qiu and Chen 2006; Maze et al. 2013), with exception to its formation period, when there is no seasonal thermocline yet (Hanawa and Talley 2001)(Figure 1.1).

Figure 1.1: Example of a typical vertical profile of temperature for mid latitudes. During summer or spring, a seasonal thermocline develops above the mixed layer. The same process occurs in mode water formation, with the exception that mode waters are advected away from formation sites by local circulation. <https://en.wikipedia.org/wiki/Thermocline>

The nomenclature for mode water came with the description of a layer with temperature ranging between 16 °C and 18 °C in the North Pacific, at the northwest portion of its gyre (Masuzawa 1969). Many different studies called "mode water" a number of thick near-surface layers. Worthington (1959) first described the Eighteen Degree Water (EDW), at the Gulf Stream Extension, which was later on compared by Masuzawa (1969) to the Subtropical Mode Water (STMW) in the North Pacific. The term can also be found in studies about the Subantarctic Front (the Subantarctic Mode Water - SAMW); the North Atlantic Subpolar Mode Water (SPMW) in the subpolar gyre (*e. g.* McCartney 1977, 1982).

1.2 Location of Mode Waters

Mode waters can be found on the warm side of fronts or boundary currents in every ocean basin, although subtropical mode waters are bounded to Western Boundary Currents (Hanawa and Talley 2001). Hanawa and Talley (2001) listed the known mode waters throughout the globe. Worthington (1959) first described the EDW in the North Atlantic as an isotherm ($17.9\text{ }^{\circ}\text{C} \pm 0.3\text{ }^{\circ}\text{C}$) and isohaline (36.5 ± 0.10) layer that extended from the Sargasso Sea to as far east as 40°W , typically at 300 m depth with thickness no less than 284 m. A similar feature was also observed by Masuzawa (1969) for the North Pacific, which was named Subtropical Mode Water (STMW). In the North Pacific and in the North Atlantic the mode waters are related to warm Western Boundary Currents of the corresponding subtropical gyre, namely the Gulf Stream and the Kuroshio, meaning that they can be referred to as subtropical mode waters. Another type of mode water, subpolar mode waters, can be found in two different locations, one in the North Atlantic, the Subpolar Mode Water (SPMW), and the other one in the Antarctic Circumpolar region, the Subantarctic Mode Water (SAMW). Those mode waters form at higher latitudes than their subtropical counterparts, also due to winter convection. SAMW is laterally restricted to the southwestern sector of the subtropical gyre. The potential density range in which subpolar mode waters are found is similar to the lower pycnocline of the subtropical gyre, in contrast with STMW which are formed in the upper pycnocline water (McCartney 1982).

For convenience subtropical mode waters can be classified in types. Type I corresponds to subtropical mode waters related to western boundary currents and

can be found in every ocean basin (Hanawa and Talley 2001). They form when cold and dry winds cause intense heat loss during winter. The already mentioned EDW, or North Atlantic Subtropical Mode Water, belongs in this group. Type II mode waters are also subtropical and similar to the first type but found on the eastern side of the basin. We can mention, *e. g.*, the Madeira Mode Water, described by Siedler et al. (1987). These waters appear close to the subtropical gyre, on the equatorward side of zonal fronts. In contrast to their western counterpart, type II subtropical mode waters occupy smaller areas on the eastern side of the basin and are much less thick. Type III subtropical mode water is associated with subpolar fronts. These waters are denser than the two other types forementioned and are related to the poleward boundaries of subtropical gyres. For instance, one can cite the SAMW located north of the Subantarctic Front which is the northernmost front of the Antarctic Circumpolar Current (McCartney 1977), and extends itself to all basins with varying temperatures and it is not clear if it is always connected; SAMW is also closely related to the formation of Antarctic Intermediate Water (AAIW) (Piola and Georgi 1982; Sloyan et al. 2010). There is a fourth kind (Type IV mode water) which is sometimes called Subtropical Underwater, found in high evaporation regions where this phenomenon causes subduction of saline water. Figure 1.2 depicts the locations and types of described mode waters according to Hanawa and Talley (2001).

Figure 1.2: Distribution of mode waters in the global ocean. The colors indicate mode water types: red are Type I mode waters linked to Western Boundary Currents, pink are Type II subtropical mode waters, type III subtropical and subpolar mode waters are coloured brown. Approximated values of potential density (σ_θ) are noted. The contours depict the subtropical gyre circulation. Source: Lynne D. Talley (2011)

1.3 The Study of Mode Waters

Every mode water type is characterized by its homogeneity, regardless of formation processes or the presence of boundary currents or fronts. Highly mixed water will present low vertical gradients of temperature, salinity, oxygen, density, etc. Each one of these variables may be proxies for mode water identification, although potential vorticity (or isopycnic potential vorticity, Hanawa and Talley (2001)) is preferably used for mode water detection (McCartney 1977; McCartney 1982; Talley and McCartney 1982; Provost et al. 1999; Billheimer and Talley 2016).

In contrast, some studies have used only the vertical gradient of T (Worthington 1959; Siedler et al. 1987; Kwon and Riser 2004; Qiu and Chen 2006). McCartney (1982) first proposed potential vorticity (q) as an indicator of mode water. Potential vorticity is a conservative property which means that away from regions of high heat exchanges and salinity alterations (rivers, high precipitation or evaporation) mode water will preserve this information for a long time. EDW has been recorded to preserve homogeneity of a single cold core for up to 5 years (Talley and Raymer 1982) and exhibit a turnover time of 4.00 ± 0.77 years (Kwon and Riser 2004).

Generally mode water studies have set the q limit criterium to $q \leq 10^{-10} m^{-1} s^{-1}$ (Talley and Raymer 1982; Provost et al. 1999; Qiu and Chen 2006; Sato and Polito 2014; Billheimer and Talley 2016), and many different methodologies have been used to detect mode water. The goal of this section is to list the many methods that can be used in the study of mode waters.

1.3.1 In situ data

Talley and Raymer (1982) studied the EDW from hydrographic data recorded from 1954 to 1978 on Panulirus station near Bermuda. The authors were able to identify changes in EDW on scales of five to ten years, but could not find a correlation between the shifts of temperature and density on EDW to surface heat flux interaction between the ocean and the atmosphere. However, there was evidence that those differences were due to a different source region for EDW formation and the stratification of the water column when the EDW is formed.

Provost et al. (1999) utilized *in situ* historical hydrographic stations and World Ocean Circulation Experiment (WOCE) XBT data to study mode waters in

the South Atlantic. Three different types of subtropical mode water were identified along the basin and with different properties, but consistently low q values ($\leq 1.2 \times 10^{-12} m^{-1} s^{-1}$), with densities of $\sigma_\theta = 26.2 \text{ kg m}^{-3}$, $\sigma_\theta = 26.5 \text{ kg m}^{-3}$ and $\sigma_\theta = 26.7 \text{ kg m}^{-3}$.

Kwon and Riser (2004) combined Argo, profiling floats and World Ocean Database 2001 (WOD2011) data to evaluate annual variability of NASTMW (determined using only T and $\frac{\partial T}{\partial z}$) throughout 40 years worth of data. The authors found that heat content in NASTMW is more correlated to the volume of mode water than its average temperature. Winter temperatures were colder (although the difference was not statistically significant), however the greater volumes of the NASTMW in this season still accounted for more heat content.

Palter et al. (2005) analysed the influence of NASTMW, or EDW, on the nutrient pool utilizing WOCE and Bermuda Atlantic Time Series (BATS) data and hydrostations. The authors concluded that as NASTMW is advected laterally it diminishes primary production inside the gyre, because during its formation period while the mode water is still at the surface prime production consumes most of the NASTMW nutrients.

Lévy (2005) compared the works of Palter et al. (2005) and Williams (2001), based on WOCE data, to understand how mode water formation and dispersion are related to the primary production deficit in the North Atlantic Subtropical and production surplus in the Subpolar gyre on the same ocean. The authors found that in the Subtropical gyre prime production drops. This is caused by NASTMW in the euphotic zone. NASTMW is very poor in nutrients as they are consumed during the mode water formation (as also seen in Palter et al. (2005)). The Subpolar

gyre in contrast is actually "fertilized" by Sub-Antartic Mode Water. This mode water resurfaces in the northern hemisphere after a long advection from the Southern Ocean.

Maze et al. (2009) enhanced a previous framework by Walin (1982) to study air-sea interaction connected to mode water, namely the NASTMW. They analysed formation and transformation maps (maps showing the formation process and the destruction of this mode water) derived from Argo data combined with Ocean Comprehensive Atlas (OCCA) data developed at Massachusetts Institute of Technology by Estimating the Circulation and Climate of the Ocean (ECCO). The authors found that the primary formation for EDW is limited to between the Gulf Stream (GS) and 30°N to the west of 45°W.

To cite another initiative, the **Clivar Mode Water Dynamic Experiment** (Marshall et al. 2009) carried out an extensive effort to register the biochemistry and physics involved in every aspect of NASTMW, rising from keystone questions as: the unbalanced volume budget of mode waters (formation and destruction rates do not balance with the residual volume as for current calculations) and its somewhat uncertain formation process (although there are some qualitative descriptions for it, there are still questions about how every aspect of it actually takes place, *e. g.* the role of surface heat fluxes or the water column stratification during the formation period). Amongst the sampling methods employed are moorings, profiling floats, surface drifters and cruises, along with modeling techniques.

1.3.2 Numeric modeling

de Miranda et al. (1999) used a high resolution model for the South Atlantic to simulate mode water formation and the influence of ocean-atmosphere heat exchanges and eddies distribution in this process. They found that STMW form in a quasi-zonal line to the north of the Subtropical Front. The main mechanism to STMW renewal was found to be seasonal variability. Eddies were responsible to mix the mode waters into the gyre, in this simulation. The study analysed the last 5 years of a 34-year long experiment.

Maze et al. (2013) used a numerical model to analyse how potential vorticity fluxes related to formation of NASTMW are influenced by loss of buoyancy and by Ekman pumping, ultimately analysing the role of mesoscale eddies on NASTMW formation, which resulted that mesoscale processes (mechanically driven PV changes by Ekman pumping or eddies) not strongly affecting the PV structure or mode water formation as surface heat flux.

1.3.3 Remote sensing

Qiu and Chen (2006) used data from World Ocean Atlas 2001 (WOA01) (Conkright et al. 2002), composed of CTD, expendable bathythermograph (XBT) and Argo data, with altimetry data to acquire eddy information to understand how Eddy Kinetic Energy (EKE) and its decadal variability influences North Pacific Subtropical Mode Water (NPSTMW) formation and found that inside the Kuroshio Extension (KE) recirculation gyre decadal changes in STMW are strongly related to the KE jet dynamic state.

Qiu et al. (2007) deployed profiling floats and combined the data with altimetry measurements to determine σ_θ (potential density (ρ_θ) - 1000; potential density is the equivalent density a water volume would have if it were adiabatically displaced to a reference level, normally the surface) and study the influence of eddies on NPSTMW, finding that cold rings lower the PV of STMW by carrying high PV water from the northern KE region into the recirculation gyre.

1.4 South Atlantic Mode Water

Knowing what defines mode waters, where they can be found and how they can be observed, we can now focus on the South Atlantic and what is known about STMW in this ocean. The SASTMW is part of the South Atlantic Subtropical Gyre. The gyre is bounded west by the BC, and the Brazil-Malvinas Confluence (approximate latitude 38°S), bounded south by the South Atlantic Current and east by the Benguela Current (Figure 1.3).

Figure 1.3: The South Atlantic surface circulation as adapted from (Stramma and England 1999)

Provost et al. (1999) identified mode waters in the South Atlantic using *in situ* data. They used a combination of CTD data from the South Atlantic Ventilation Experiment (SAVE), the Oceanus and Ajax sections, cruises of the confluence program, the Marathon expedition (Gordon 1989), and other CTD cruises. Expendable bathythermograph (XBT) of the WOCE program were also used. Mode waters were identified as layers thicker than 40 m, with vertical temperature gradients lower than $0.015\text{ }^{\circ}\text{Cm}^{-1}$ for stations west of 20°W or lower than $0.010\text{ }^{\circ}\text{C/m}$ for stations east of 20°W , and q values below $1.2 \times 10^{-12}\text{m}^{-1}\text{s}^{-1}$. Their variation was compared with atmospheric forcing variation, in the form of heat flux variation as observed in the NCEP data. Three types of STMW in the South Atlantic were found. They differ in potential density σ_{θ} , and consequently in depth. The first type

is a subsurface water with $\sigma_\theta = 26.2 \text{ kg/m}^3$, the second type with values around 26.5 kg/m^3 and the third around 26.7 kg/m^3 . Mode water variability was correlated to atmospheric events, as harsh winters were followed by larger newly formed STMW volumes, and weak winters preceded smaller volumes. The main region for ventilation was the western boundary of the subtropical gyre (from coast to 20°W). Provost et al. (1999) did not find any correlation between mode water thickness and heat flux variations on a decadal scale, which was proposed to be a result of the eddy nature of STMW, which would difficult detection of decadal variations on it. Concluding, only qualitative relations with atmospheric forcings could be found, but on limited timescales. Decadal variability was not detectable.

Sato and Polito (2014) utilized Argo profilers data to detect the STMW. Using a cluster analysis taking into account potential temperature, salinity, potential density, location and time of the year, three kinds of mode water were identified. The goal of the study was to determine with a larger database (comparing to Provost et al. (1999)) whether air-sea interaction processes, in the form of surface net heat flux, could be connected to variations in volume (thickness) of STMW. They also examined the role of EKE in mode water formation. No clear relation between heat flux and layer thickness could be found quantitatively. EKE does not seem to be affected by mode water, or vice versa, except by the fact that 80% of the eddies found with mode water signatures were anticyclonic, and had also a thicker layer of mode water. This raises two possibilities: eddies are more often formed when STMW is thicker, or cyclonic eddies dissipate or destroy the mode water layer, whereas anticyclonic eddies create it.

1.5 The use of PIES to study mode waters

Inverted echo sounders (IES) are instruments composed of a transducer which can receive and emit sound through the water column, and a precise clock to register the return time of the signal. This model of instrument was idealized and built first by Rossby (1969). Pressure equipped inverted echo sounders (PIES) have been used for many different purposes to study dynamical oceanic processes, such as internal waves (Li et al. 2009), planetary waves, fronts, currents, mesoscale eddies and internal waves (Li et al. 2009).

Qiu et al. (2007) have used current equipped PIES (CPIES) to investigate the impact of cold core eddies in the NPSTMW formation on a monthly scale. It was previously suggested by Qiu and Chen (2006)) that the STMW formation in that region was more sensitive to KE changes rather than atmospheric conditions.

Combined with empirically-derived look-up tables, via the Gravest Empirical Mode (GEM) (Meinen and Watts 2000; Watts et al. 2001; Meinen et al. 2012) method, full-water-column profiles of temperature, salinity and density can be determined. The GEM are lookup tables obtained from *in situ* data (collected between latitudes 32°S and 3°S, longitudes 55°W and 40°W). The available return time τ is integrated from temperature and salinity profiles (section 3.3). In other words, the available CTD and Argo profiles in a specific region are gathered; from each profile τ is calculated, and a look-up table matching τ and T (or S) is calculated. Through the GEM look-up tables, the τ data from a PIES can be converted into vertical profiles of temperature, salinity or density. These PIES derived T and S data can be used to calculate specific volume anomaly, steric height, the mass load-

ing component of sea surface height, and, if multiple PIES are available, barotropic and baroclinic velocities (resulting in absolute velocity), depending of the array grid used. Velocity fields derived from PIES and CPIES have been shown to have a close relation with current meter data (Meinen and Watts 2000; Meinen et al. 2017).

The GEM technique has been used to study currents and fronts (Watts et al. 2001; Meinen and Watts 2000). Meinen and Watts (2000) used PIES data to study the North Atlantic Current (NAC) in Newfoundland Basin, describing its transport, structure and finding evidence that suggests there is a recirculation gyre on the inshore part of the NAC. Watts et al. (2001) observed the Subantarctic Front vertical structure along the WOCE SR3 transect, south of Australia. In the latter, an improvement of the PIES data analysis was suggested with the inclusion of a seasonal correction look-up table, that was developed for the surface layer data (100-200 dbar).

In the present study, we intend to use GEM with the inclusion of a similar seasonal look-up table specific to investigate the South Atlantic subtropical mode water. The PIES data are obtained by the Southwest Atlantic Meridional Overturning Circulation Program (SAM) PIES. Combined with modeled dynamic variables, we test the covariance between mode water thickness and the local dynamics.

2. Objectives

In this study we work with the hypothesis that the interannual changes observed in STMW layer thickness are correlated with local dynamics, mainly dominated by transport of tropical waters carried by the Brazil Current (BC). Ocean-atmosphere interactions have strong influences on STMW, but they are not the only processes to affect the mode water. Ocean dynamics are also a key factor, as local stratification or the eddy kinetic energy state of the study areas (Qiu and Chen 2006). Our main objective is to obtain a better understanding of the interannual variation of subtropical mode water formation through the use of temperature and salinity data derived from the PIES array observations of acoustic round-trip time and evaluate its covariance with the dynamics of the BC. For that, it is necessary to adapt the technique of processing the raw data measured by PIES instruments and adjust them to detect STMW both in formation (on the surface) and fully formed (in subsurface).

2.1 Specific Objectives

The study presents the following specific objectives:

- i) Develop empirical temperature and salinity fields based on *in situ* data which will be the foundation for the translation of the PIES travel time into those variables. Adapt the technique with a seasonal correction. This empirical field is known as Gravest Empirical Mode (GEM) and will be described in detail in section 3.

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- ii) Detect the presence of STMW on the western side of the South Atlantic, through the data gathered by the PIES and analyse the changes in the mode water layer thickness from seasonal to interannual scales.
 - iii) Compare the results for mode water formation with the WOA13 and In Situ Analysis System (ISAS) data.
 - iv) Correlate interannual changes in SATMW layer thickness with the changes in the BC volume transport as modeled using Hybrid Coordinate Ocean Model (HYCOM).

3. Data and Methods

3.1 Summary

When studying ocean dynamics, the collection of reliable data is constrained by weather, funding, and methodology. Many alternatives have been used through history: XBT, CTD sensors, remote sensing, etc. Watts et al. (2001) presented a new method for obtaining water properties, such as T, S and specific volume anomaly δ with better accuracy in the upper 200-300 m using PIES. Following the methods presented by Watts et al. (2001), CTD and Argo profiles collected inside the region comprised between 32°S and 38°S, 55°W and 44°W were used to make "look up tables" that relate τ and T profiles. Travel time is an integral property and uses sound speed and density. The speed of sound in the ocean depends on T and S, and can be simulated using

$$\tau = 2 \int_0^p \frac{dp'}{\rho g_0 c}, \quad (3.1)$$

(Meinen and Watts 2000) where τ is the return time (s), g_0 is the acceleration of gravity ($m s^{-2}$), ρ is water density, p is pressure (dbar) and $c = c(T, S, p)$ is the sound speed in the water column from an empirical equation (Del Grosso 1974; Meinen and Watts 1997).

Using equation 3.1 we calculate τ for each profile available. As described in Meinen and Watts (2000), vertically integrated properties are subject to changes in the water column's structure. PIES records τ , one of such integrated properties,

which can also be calculated for T and S profiles from CTD and Argo data. The GEM method uses the relation between τ and T or S profiles via look-up tables. Those tables can then be used to reconstruct the profiles from T and S. The standard GEM fields reconstruct the T and S profiles without a time dependency (Meinen and Watts 2000), so the seasonal corrections of Watts et al. (2001) must be applied to recover the seasonal variations in the near surface layer once STMW formation processes depend on surface air-sea interaction and are subject to relatively high variability in comparison with subsurface layers. Following Watts et al. (2001) methods, to recover STMW formation information from the data, a seasonal look-up table was fit to correct the water properties in the upper 200 m (although other levels were considered as in section 3.4.3). A seasonal correction look-up table was fit to correct this issue and try and identify mode water as it forms. With T and S information, mode water can be identified and traced along time series of q , its layer width determined and formation periods detected. Later, we calculate correlation between mode water cycles and dynamic variables to understand how local dynamics influences mode water formation and variability. The use and elaboration of the GEM fields to reconstruct temperature profiles is explained ahead.

3.2 Data and Study Area

In this study we evaluate the data collected by the PIES moored at 34.5°S from 51.5°W to 44.5°W. This array is part of the SAM project. The moorings are a few degrees north of the Brazil-Malvinas confluence and the separation of the BC from the continental shelf (Goni and Wainer 2001).

3.2.1 PIES

PIES (Figure 3.1) are equipments that emit a sound pulse and receive the reflected signal, recording the return time τ of the pulse and the ambient pressure with daily resolution. The PIES array was designed to study the Deep Western Boundary Current (DWBC) and deployed in 2009 (Meinen et al. 2012). The coordinates and measurement timespan of each instrument is listed on Table 3.1. Here we explain thoroughly how the PIES and CPIES work. The instrument emits a 12 kHz signal in bursts up to 24 pings from a transducer every hour with 0.06 ms precision for each ping and a standard deviation typically below 2.2 ms for each 24 ping burst, and records the travel time using a high quality crystal clock. The data for each hour corresponds to the median value of these pulses. This process is taken to prevent waves and other surface short time variations to generate interferences in the resultant data. Typically, the return time of the pulse ranges between 1 and 8 seconds depending on the ocean depth, and the significant variations of the signal occurs in a scale of milliseconds. When equipped with a pressure sensor, they are called PIES, and when there is also a current meter, CPIES. CPIES have a doppler current meter as part of the mooring at 50 m above the bottom.

Figure 3.1: Photography of a CPIES. Actual moorings have the current meter (red instrument) 50 m apart from the PIES (bottom).

<http://www.po.gso.uri.edu/dynamics/ies/hoverbox/index.html>

These equipments are often moored at the ocean bottom, and record daily information of τ . From the combination of pressure, converted to depth, and the two-way travel time, average sound speed can be calculated for the water column. The time measured is later converted with the GEM technique into temperature and salinity profiles with daily resolution. The array containing the four stations used in this study is part of the South Atlantic Meridional Overturning Circulation (SAMOC) and their location is shown in Figure 3.2. The PIES data has a daily resolution, and in the present study we used data from March 2009 to December 2015.

Figure 3.2: Location of the PIES moorings from the SAM program used in this study (as seen on Meinen et al. (2012)).

3.2.2 HYCOM

The HYCOM is an evolution of the Miami Isopycnic Coordinate Ocean Model (MICOM). The model was developed in a joint effort of the University of Miami, the Naval Research Laboratory (NRL), and Los Alamos National Laboratory (LANL). Those institutions are part of the HYCOM Consortium for Data Assimilative Ocean Modeling founded by the National Ocean Partnership Program (NOPP) in 1999 to develop and evaluate a hybrid coordinate isopycnic-sigma-pressure ocean model with data assimilation support (Bleck 2002; Chassignet et al. 2003; Halliwell 2004). This model, as the MICOM, uses five prognostic equations: the mass continuity equation, two equations for the horizontal components of velocity and two conservative equations for a pair of thermodynamic variables, such as temperature

Table 3.1: Coordinates for the PIES moorings.

Latitude	PIES A	PIES B	PIES C	PIES D
34.5°S	51.5°W	49.5°W	47.5°W	44.5°W
Start of time series	18/03/2009	07/07/2011	19/03/2009	07/04/2009
End of time series	11/10/2014	10/11/2015	06/10/2014	07/09/2013

and salinity or density and salinity (Bleck 2002). The use of hybrid coordinates allows the HYCOM to benefit of the three most used coordinates: isopycnic coordinates that represent surfaces of constant density to model deep and stratified oceans; z levels that represent set depth levels or surfaces of constant pressure to model near surface (mixed layer) processes, where a refined vertical resolution is necessary; and sigma (σ) coordinates used in intense topographic gradients regions such as the continental shelf break and continental slope (Bleck 2002; Halliwell 2004; Chassignet et al. 2003; Chassignet et al. 2007). The data presented here was a selection of a run with a horizontal resolution of $1/12^\circ$ (8.5 km), and 32 hybrid vertical levels. This resolution can resolve mesoscale processes (eddy resolving). The topography used is NRL DBDD2, from NRL, with 2' spatial resolution. These data were already validated by Castellanos et al. (2016), and as this process is not the center of our study it was not replicated.

3.2.3 ISAS temperature data

The ISAS is produced and distributed by the French Research Institute for Exploitation of the Sea (IFREMER) (Gaillard et al. 2009; Brion and Gaillard 2012). The T data were used in this study as a second source of the depth of the

16 °C isotherm. In addition to WOA13, this data set was used to evaluate the penetration of the annual signal in upper layer of the ocean. The ISAS is a monthly mean climatology of global T and S profiles obtained from interpolation of available in situ data from Argos, XBTs, CTDs and instruments on moored buoys (Gaillard et al. 2009). This data set span the period between 2002 to 2014. From the 12-year long time series we calculated the mean year for T and S.

3.3 GEM

τ information is recorded/measured by PIES, and afterwards converted to vertical profiles of T or S using "look up tables", with τ as the abscissa and T or S as the ordinate. The standard GEM tables use *in situ* data (from both Argo profilers and CTD casts) to convert acoustic travel time (sound round trip between p and 0 dbar) τ to temperature profiles.

The GEM technique (Meinen and Watts 2000) has proper vertical coherence with in situ data, as we will discuss in section 3.5. The result does not show any seasonality (Figure 4.6), so a seasonal correction look-up table is fit to data for that matter. Seasonal corrections (section 3.4) are carried as described in (Watts, Sun, and Rintoul 2001), which attempt to include some temporal dependency on reconstructed T profiles by altering the GEM fields.

The process to produce the GEM fields has 4 steps.

- First, the *in situ* data are processed to exclude spurious data. This was done by the National Oceanic and Atmospheric Administration (NOAA). The data used in this study was obtained via personal co-

munication. The processed data can be assessed in the NOAA site http://www.aoml.noaa.gov/phod/research/moc/samoc/sam/data_access.php;

- Then τ is calculated using equation 3.1 and a cubic smoothing spline is applied to T and S as function of τ for each pressure level (standard depths from 0 to 5100 dbar). The signal return-time is calculated using 1000 dbar as the reference level p for the equation 3.1 integral's limit. Henceforth, τ_{1000} is referred to as just τ . Each profile from CTD and Argo data has its τ calculated (τ_G). Figure 3.3 is an example for $\tau_p = \tau_{200}$. The blue line is the fitted spline on a regular grid of τ from 1.3100 to 1.3460 s.
- A second smoothing is done, now over p levels, with a cubic spline with knots. This results in regularly spaced temperatures, and also corrects not realistic oceanic features that are introduced to the data at the edges of the τ grid due to extrapolation of the first smoothing process (Meinen and Watts 2000). At this stage there is no time-of-year dependency on T and S derived from τ , as all values are combined together despite the time of the measurement.
- T and S tables as function of τ_G are grouped together. More precisely, τ_G values that differ to ± 0.5 ms of the regular grid of τ . Figure 4.1 is an example of the look-up table for T.

At this stage there is no time-of-year dependency on T and S derived from τ , as all values are combined together despite the time of the measurement. This results in typical thermocline representations (Figure 3.4) that average effects of time dependent cycles, such as seasonal cycles. As the directly measure CTD and Argo data used to simulate τ are not evenly distributed through time of the year,

Figure 3.3: Example of how τ_{ds} was calculated. The scatterplot is τ plotted against T_p , which is T_{200} in this case. The blue line is the fitted spline to that pressure level. The value in blue will be used when the profiles are reconstructed using the GEM table.

the resulting GEM estimates are a sort of weighted average, tending to the time with more profiles available. This represents an issue as we intend to evaluate time dependent processes. Therefore a seasonal correction look-up table is fit to this data to correct this problem, and will be discussed in section 3.4. After taking the steps listed above, but not yet considering the seasonal correction, the result is what is called a GEM field. A table that relates a specific τ value to a T (and also S , not shown in figure) profile, as can be seen in Figure 4.1.

Figure 3.4: Example T profiles on site D of the SAM array from 02/03/2010 to 05/02/2011.

There is one profile at every 20 days.

3.4 Seasonal Correction

The first 200 dbar of the *in situ* data are subjected to seasonal variability, thus they are more scattered when compared to the GEM derived T or S profiles

(Watts, Sun, and Rintoul 2001). The formation and dissipation of the seasonal thermocline along with the mixed layer annual variability are responsible for these fluctuations, or variability in the upper portion of the observed profiles. Since this variability is lost during the GEM tables elaboration and application because different profiles can correspond to the same τ and the GEM tables do not consider time of the year correspondency to the profiles. A regional seasonal correction look-up table can be fit to capture this variability and enhance surface data precision and maintain its seasonal cycles.

Mode water formation is closely connected to the seasonal thermocline. The seasonal thermocline forms during late spring and summer, after the STMW surfaces in winter and strong convection lowers the upper layer's stratification. The mode water is then trapped between seasonal and permanent thermoclines until it is advected away from the formation region. In the GEM tables there is no such seasonal variability in the upper layer to form and destruct a seasonal thermocline, hence the importance in this study to examine the inclusion of a seasonal correction look-up table, otherwise mode water formation would not be a reasonable subject for discussion.

The seasonal correction consists of including $T_{(p,t)}$ and $S_{(p,t)}$ cycles to the GEM profiles as a function of time of the year, irrespective of which year, for levels above a reference level in the water column influenced by the annual signals. Two different evaluations were made to test which reference level would best fit the data and they will be discussed later in section 3.4.3. Using the same *in situ* data that generated the GEM tables, we could fit the annual marches of T, S and τ , and then apply the correction to the profiles. This returns the seasonality to the reconstructed

profiles, which the GEM fields do not reproduce. Since this cycle is derived from the same data that resulted in the fields GEM, we expect that this correction will agree with observations, and the tests for this will be discussed later on this section.

3.4.1 Seasonal correction for τ

The correction in the τ data is to remove an annual cycle fitted to the acoustic travel time correspondent to the layer above the reference level τ_p . Annual variability is overwhelmed by higher frequencies in the reconstruction, then removing it from the GEM secures that the correction applied later to $T_{(p,t)}$ and $S_{(p,t)}$ will account for all the annual variability. This procedure will allow us to 'de-season' the travel times. Watts et al. (2001) explained that their reference level of choice corresponded to a pressure below which the annual cycle contribution was less significant. In this study we tested several different levels and used the criteria described in section 3.4.3 to find the best value to fit our data. The seasonal changes were isolated from the signal by taking the residual from the τ_p to τ curve (Figure 3.3).

The residual was after grouped and averaged in monthly bins and repeated three times, making a three year record. The seasonal correction for τ was calculated as a curve was fit to τ_p vs τ_{1000} data. The mean year was repeated three times and a second order lowpass (three months) Butterworth filter was passed forwards and backwards. The middle year curve $\tilde{\tau}_{(t)}$ (Figure 3.5) was retained, to avoid imprecisions on the edges of the total curve. Finally, the de-seasoned τ was calculated as

$$\tau_{ds} = \tau - \tau_p.$$

Figure 3.5: Examples of how τ_p was calculated. The scatter is τ_p vs τ_{1000} and the blue line is the butterworth-filtered $\tilde{\tau}(t)$.

3.4.2 Seasonal correction for T and S

Similarly as done for τ_p , near surface processes dominate the variability of physical parameters at the upper layer, and the seasonal corrections extend to T and S. We then proceeded in isolating all seasonal changes by removing the GEM fields, thus modeling the seasonal residue. At each pressure level between 0 and p (reference level), the residue of T (T') was calculated by subtracting at each individual *in situ* profile its GEM profile, that is, $T'_{(p)} = T_{(p)} - T'_{G(\tau,p)}$, indexed by the time of the year. Just as in section 3.4.1, the T (or S) annual march ($\tilde{T}_{(p,t)}$ and $\tilde{S}_{(p,t)}$) at each pressure level was calculated with the same method used on τ . The exact same process was

applied to salinity ($S'_{(p)} = S_{(p)} - S'_{G(\tau,p)}$). The resulting $\tilde{T}_{(p,t)}$ and $\tilde{S}_{(p,t)}$ are discussed later in chapter 4. As mentioned before on section 3.4.1, several reference levels were tested and two different guidelines led us to the choice of which level to use. That matter shall be discussed on the following subsection.

3.4.3 Reference level for seasonal corrections

Using the same *in situ* data that generated the GEM tables, we could fit the annual marches of T, S and τ , and then apply the correction to the profiles. This returns the seasonality to the τ retrieved T profiles, which the GEM fields do not reproduce. Since this cycle is derived from the same data that resulted in the GEM fields, we expect that this correction will agree with observations, and the tests for this will be discussed later in this section.

Deciding the reference level was the first step to reproduce a seasonal correction. Watts et al. (2001) chose their reference level by testing several values to find out at which depth the contribution of the annual signal was considered insignificant. One way to find out how the annual cycle contributes to the local variability was to observe how the amplitude of an annual cycle changes with depth from a long-term annual mean data set. Using the WOA13 (Locarnini et al. 2012) temperature data for the nearest grid points relative to the PIES stations, we fit a sinusoidal curve with a period of one year to every depth available to observe how the amplitude changes with depth. Reference levels from 100 dbar to 360 dbar were tested. Analogously, the closest point on ISAS data to each PIES mooring was selected, and compared to its corresponding GEM derived T time series. We tested only T, because our goal was to find the best level on which seasonal changes should be referenced. Even

though salinity has a seasonal cycle, its contribution is complementary to T when dealing with density (which was our main goal). Besides, the S GEM fields would be subjected to the exact same method for correction applied to T. Thus, knowing the relevant level for T would be sufficient. We took two different approaches on those tests. First, a mean year was calculated for both ISAS and GEM derived T data (using the 200 dbar reference level), and a preliminary analysis was taken to check whether both were considerably agreeable on the depth of the 16°C isotherm. After that, the data from both sources for the same time had their 16°C isotherm depth subtracted, for every reference level above mentioned, to find which of those would yield less cumulative difference. Lastly, all calculated depths for the 16°C isotherm were plotted and the dispersion of the data was considered. From the results of the methods above mentioned, we decided to interpolate the chosen $\tilde{T}_{(p,t)}$, $\tilde{S}_{(p,t)}$ and $\tilde{\tau}_{(t)}$ corrections to a 365 days model, instead of the monthly binned model yielded from applying Watts et al. (2001) method.

Each and every one of the tests mentioned were key to our understanding of the reference level we should use, and whether the seasonal correction was at all useful to make STMW formation identifiable in the GEM derived profiles from PIES.

3.5 Error

A simple yet robust way to evaluate the error of the GEM estimates and the directly measured T or S profiles is to take their root mean square (rms). We calculated the error of the GEM estimates as follows:

- τ is calculated for the CTD and Argo profiles using equation 3.1;

- Pressure and τ bins are set to calculate the errors, 100 dbar and 3 ms respectively;
- Calculate the rms of directly measured profiles and the GEM estimates.
- Take the signal-to-noise ratio of the rms and the range of the CTD and Argo profiles.

3.6 STMW detection

Using the methods described in section 3.3 we obtained the time series for temperature and salinity for each PIES station. Then the profiles must be examined to establish the presence of STMW. Provost et al. (1999) and Sato and Polito (2014) used T and S criteria as a first step to determine the mode water depth and also used vertical temperature gradients ($\frac{\partial T}{\partial z}$) as well as PV as limits for mode water. We evaluated mean profiles of $\frac{\partial T}{\partial z}$ profiles for each station to determine a viable limit. STMW when submerged are limited above by the seasonal thermocline and below by the permanent thermocline. The limit for $\frac{\partial T}{\partial z}$ must be a value that excludes those features from the profiles, so we can evaluate only SASTMW. Provost et al. (1999) chose $\frac{\partial T}{\partial z} = 0.010^\circ C m^{-1}$ and $\frac{\partial T}{\partial z} = 0.015^\circ C m^{-1}$, and Sato and Polito (2014) $\frac{\partial T}{\partial z} < 0.01^\circ C m^{-1}$. In this study we chose q criteria to be $1.5 \times 10^{-10} m^{-1} s^{-1}$. Two factors contributed to that choice. Other studies (e. g. Worthington (1959), McCartney (1982), Provost et al. (1999), Sato and Polito (2014)) have found mode waters with $q \sim 10^{-10} m^{-1} s^{-1}$, and this order of magnitude was also sufficient for GEM estimated profiles. Considering Sato and Polito (2014) used this value and our study is also in the South Atlantic, we tested the criterium and the STMW layer

thickness was coherent with the literature (e. g. Worthington (1959), McCartney (1982), Provost et al. (1999), Sato and Polito (2014)).

3.7 Brazil Current Transport

Subtropical mode waters have their formation process connected to boundary currents and fronts. The BC is associated with SASTMW, and as we aim at studying the correlation of the BC dynamics and mode water thickness, we used volume transport as a dynamic indicator. We used Hycom's meridional velocities at 34.5°S to estimate the BC volume transport.

In this study we used the meridional velocity to calculate BC integrated volume transport as follows,

$$V = \int_0^z \int_{x_e}^{x_w} v dx dz, \quad (3.2)$$

where v is the meridional component of water velocity, x_e and x_w are the eastern and western limit of the BC, respectively. We selected data from 2009 to 2015, to match the PIES data, and the closest latitude of the model to the array, at 34.5031°S.

To solve equation 3.2 we need to define zonal and vertical limits for the BC. No studies about the variability of the BC close to 34.5°S have been found. We needed to establish the vertical and horizontal limits for the current using the available data. The presence of the BC was examined from the long-term mean of the meridional current velocity in the western side of the section. The bracketing of BC using the location of a specific isopycnal is tested in this study. The mean flow in the region according to the model can be seen on Figure 3.6. (Meinen et al. 2017) assessed the characteristics of the Deep Western Boundary Current (DWBC) at this exact same array. Their upper limit for the DWBC transport integration

was decided as 800 dbar, as integrations starting from this level and starting from 2000 dbar were very high correlated ($r = 0.98$). Being this the same study area, we could infer that the bottom limit for the BC could easily be set as 800 dbar. Figure 3.6 has the -0.6 m/s isotach outlined, and below its deepest limit (1000 dbar), the flow is very homogeneous in terms of velocities. It is reasonable to establish that this could also be a limit for the integration.

Figure 3.6: Average meridional velocity section at $\tilde{34.5^\circ\text{S}}$ from HYCOM. Bold lines are the isotachs of 0 and -0.06 m/s. The BC is the structure between 52°W and 49°W . The white patch on the left side of the figure is the continental shelf. Blue contoured squares shows SAM sites A to C (D not shown).

In order to be safe and follow common methods in the study of the BC, we chose to use an isopycnal indicator. The limit between Upper Circumpolar Water (UCPW) and North Atlantic Deep Water (NADW) was chosen as such limit.

Table 3.2: Comparison between BC transport integrated from 0 dbar to 800 dbar and from 0 dbar to 1200 dbar.

Reference	800 dbar	1200dbar
Mean Transport (Sv)	-2.6380	-3.7769
Standard Deviation (Sv)	2.3242	3.3830

The vertical limit z on equation 3.2 was defined as the mean depth of the $\sigma_1 = 32.2 \text{ kg m}^{-3}$ (potential density ρ_θ referenced at 1000 m) where the closest value was $z = -1200 \text{ m}$ (Figure 3.7). The zonal limits were defined based on the sign of the meridional component of the current's velocity. Using this limits might overestimate the transport a little, but as we are interested in its variability, this should not compromise our analysis. Furthermore, the resulting transport using this level (not shown) was not statistically different from the one calculated using 1200 dbar as shown in Table 3.2.

The western limit of the current was the first negative value between 52°W and 51.5°W . If no value satisfied this condition, 51.5°W was adopted as the western limit, assuming that the current will not detach more than $\sim 50 \text{ km}$ from the continental shelf break. The eastern limit was taken as the first positive value of v between 50.5°W and 48.5°W . Again, if no value satisfied, 48.5°W was adopted, respecting the condition that the BC may not deviate more than $\sim 50 \text{ km}$ from its average location. With the integration limits defined, equation 3.2 was calculated at each time step.

Figure 3.7: The isopycnal that defines the limits of UCPW and NADP used to define the lower limits of the BC (grey line). This value was used to define the vertical limit for the integration on equation 3.2.

4. Results and Discussion

4.1 Gravest empirical Mode

The processing of the data on section 3.3 allowed us to build the look-up tables that related T and S profiles to τ data recorded by the four PIES moorings located at the sites listed on table 3.1. A total of 545 profiles were used to create the GEM fields presented here. The GEM table for temperature can be seen on Figure 4.1.

Figure 4.1: The GEM fields relate typical T or S profiles to τ values interpolated at every p level and then a second interpolation. This example illustrates the "lookup table" for T (contours, colorbar units in $^{\circ}\text{C}$) profiles as function of τ .

From the PIES round-trip travel time we could recover the temperature and salinity time series by matching the τ values measured by each PIES, comparing it with the look-up table and using the correspondent T or S profile. The results for T are depicted on Figure 4.2. We do not discuss the role of salinity alone, but we will address it later along with temperature in the form of PV values.

Figure 4.2: Time series of T profiles for the first 600 dbar of 2009. Blank sections are periods of data unavailability. Contours are T ($^{\circ}$ C)

Meinen et al. (2012) has already evaluated GEM application in the South Atlantic. In this study we used the same data, the same methods for GEM and replicated the error calculations that Meinen et al. (2012) used. The signal-to-noise ration (SNR) is very similar to that found on GEM tables on the North Atlantic (Meinen and Watts 2000; Watts et al. 2001). SNR ratios on the thermocline and halocine depth range are ~ 20 , and drop to 1-3 below 2000 dbar in the deep ocean (Figure 4.3). The rms values on the permanent thermocline/pycnocline express the already discussed smoothing of the upper layer processes, which is more evident on the SNR on the middle pannel of Figure 4.3. Upper layer variability in the ocean

is much more intense than below 2000 dbar due to superficial advection and ocean-atmosphere interaction processes, but still the GEM estimates explain $\sim 80\%$ of the data variability, which is fairly accurate. The estimates tend to be less accurate near the edges and bottom of the GEM tables because these regions are often filled with interpolations and extrapolations due to less data availability. As our STMW is confined to the first 600 dbar, the SNR of ~ 20 is enough to validate that we are using profiles that retain a fair amount of the variability. Thus, as we analyse interannual scales the variability representation is fairly coherent to directly measurements.

Figure 4.3: Upper panel shows two dimensional look-up table of temperature as function of pressure and simulated travel time between and 1000 dbar created using the GEM method. Middle panel shows rms difference between directly measured CTD and Argo profiles and the GEM estimates using the GEM look-up tables. Rms differences were calculated with 3 ms bins for τ and 100 dbar for pressure. Gray dotted lines represent direct measures availability. Lower panel has rms as SNR based on observations at each pressure level. Source: Meinen et al. (2012).

4.2 WOA13

We used WOA13 data to evaluate the penetration of the annual signal into the water column as a function of the time of the year. Figure 4.4 shows a mean climatological year for temperature at the nearest grid points of WOA13 data to PIES-A. Warmer temperatures at the upper layer of the ocean at this region occur between late spring (November) to late fall (May) whereas the colder T are between June and October. Most of the changes in the upper layer can reach as deep as 150 m consistent with the cycle of a mixed layer and seasonal thermocline formation during the year. As a validation method for our GEM derived profiles, the TS diagram of each station was calculated and the result compared with another TS diagram from the South Atlantic region. Considering figure 4.5, the general aspect of the diagram for each station matches the WOA13 diagram. Site A shows a slight deviation for fresher water, probably due to the influence of Río de la Plata. This indicates that the general structure of the South Atlantic is reasonably present on GEM tables.

Figure 4.4: The annual mean temperature from WOA13 at the site of PIES-A. Contours are T ($^{\circ}C$)

Figure 4.6 shows the annual mean year computed from PIES A. It is conspicuous that compared to Figure 4.4, the annual mean of PIES-A temperature lacks the seasonal variability. The absence of seasonal changes was mentioned in section 3.3, because GEM derived profiles do not express seasonal variability. Compared to the climatological mean from WOA13, the temperature obtained from PIES-A (the same occurs to all other stations - not shown) is significantly colder (about $2^{\circ}C$) for the first 500 m. This will affect the criteria to search for the subtropical mode water in this region. Thus a seasonal correction look-up table like the one used by Watts et al. (2001) was applied.

Figure 4.5: TS diagram of each PIES station (black dots) and the corresponding point on WOA13 (red).

4.3 Seasonal Correction

Mode water forms due to surface processes, which are strongly influenced by seasonal variability. The T and S profiles recovered with the GEM technique provide a sort of weighted average profile (section 3.3). This characteristic of the data does not allow a discussion of mode water formation, but only the presence of mode water. Here we present the results of the tests and processes taken to decide how we would use the seasonal correction look-up table. We will be able to assess if the the final product was capable of express a process identifiable as the STMW formation and also fully formed STMW.

Figure 4.6: The annual mean temperature in the upper layer of the ocean retrieved from PIES-A time series between 2009 and 2014. Contours are T ($^{\circ}C$)

4.3.1 Seasonal model tests

In order to decide which level would be the best as our reference level for the depth of annual cycle vertical penetration, we examined the WOA13 data. By fitting a sinusoidal curve to the mean annual temporal section data with annual and semi annual (the later was not considered as a component of seasonal variability) periods, we were able to determine the relative contribution of their variability to the total variance. Evaluating this percentage (Figure 4.7), we decided to chose the level at which the percent variance had a minimum. The relative contribution of the annual and semiannual cycles in the upper layer have different vertical distribution.

The surface changes have a closer relation with the annual signal, from which we can infer that annual changes dominate the mean processes from 0 dbar to 100 dbar. Both cycles have their minimum at ~ 200 dbar, probably reflecting the mean depth of the seasonal thermocline (Figure 4.16). Figure 4.8 illustrates the result for the amplitude and variance. The annual cycle is much less intense below 200 dbar. This also indicates that seasonal corrections below this level are not necessary.

We reconstructed the seasonal signal on the surface as done in Watts et al. (2001). The 200 dbar depth was the chosen level, given that the surface signal reaches its lowest amplitude at that depth. Since the signal recovers some amplitude below 200 dbar, we also observed the percent variance that represented. The increase in the amplitude is probably connected to the seasonal variability of deep circulation, and was not a consequence of local surface processes (Meinen et al. 2012).

Figure 4.7: Percent variance of a fitted sinusoidal curve, using least squares for six months period (left) and one year (right).

After this first attempt to decide the reference level for our seasonal correction look-up table, we evaluated the difference between the depth of the 16°C isotherm found on PIES and the monthly mean climatology data from ISAS. We chose this isotherm because it represents the top of the STMW on Sato and Polito (2014) and is a good tracer of the formation of mode waters. STMW exhibits temperatures ranging between 13.4°C and 18.4°C in this area (Sato and Polito 2014). The 16°C isotherm is within that range and should become shallower or outcrop during the STMW formation. The annual mean temperature was calculated for all PIES stations and compared with the nearest grid point available on the ISAS database, which also had its mean year built from its entire time series. Figures 4.9 and 4.10 illustrate the result. There was outcropping of the 18°C isotherm on Figure

Figure 4.8: Amplitude of a fitted sinusoidal curve, using least squares for six months period (left) and one year (right) on the WOA13 temperature profiles selected at every PIES site.

4.9 and a rapid shoaling of the isotherms from June to November, consistent with STMW formation (Sato and Polito 2014). Figure 4.10 shows that GEM estimates of annual mean temperature changes are less rapid and the 18°C isotherm never outcrops, meaning there is no STMW formation present in the profiles.

Figure 4.9: Average year of T for the ISAS data. There is no data for site A because the closest grid point on ISAS data fell on land. The 16°C isotherm is marked magenta.

The ISAS database consists of interpolated *in situ* data. If the difference from the depth of the isotherms is not expressive (e.g. 50 dbar) we could say that the data are slightly consistent on a first evaluation. By calculating the distance between the isotherms we can verify how it changes with different reference levels. We took the cumulative sum of the resulting curve at each month. The cumulative sum was used because a mean distance would not account for the variability that could be present in each month. The cumulative sum of the differences in depth of the 16°C isotherm was calculated, this time using the actual time series and every reference level tested for the seasonal correction look-up table as described before (section 3.4.3). This cumulative sum was later converted into a normalized anomaly. This was done by subtracting the mean value and taking the ratio with the mean value. Figure 4.11 shows the normalized anomaly in the distance between the two

Figure 4.10: Seasonal cycle of GEM temperature (reference level 200 dbar for seasonal correction look-up table). The 16°C isotherm is marked magenta.

isotherms. The change of reference level only increased the mean distance of the isopycnals to a maximum of 10%. At the 200 dbar level the isopycnals are almost identical (the distance between them is never greater than 1% of their mean depth).

Figure 4.11: Normalized anomaly of 16°C isotherm distance from GEM data to ISAS data. The x axis represents the reference levels tested, from 100 dbar to 360 dbar.

By calculating the GEM fields with all the corrections described in sections 3.4.1, 3.4.2, 3.4.3, using reference pressure levels from 150 dbar to 360 dbar, a time series of the 16°C isotherm was calculated. Figure 4.12 shows the depth of the isotherm for all reference pressure levels tested. The difference between the estimated isotherm depths using all tested pressure levels was in the order of 1 m. The small change in depth can be used as a parameter to set the reference level on 200 dbar as was inferred before, since there's no significant difference from lower levels and even some shallower levels. The results regarding the variance and amplitude of the signal seemed to be far more relevant, and were kept as standard.

Figure 4.12: Depth of 16°C isotherm for all pressure reference levels tested.

4.3.2 Seasonal correction tables

Section 3.4 described the process used to obtain the tables that will act as the seasonal variability in the PIES derived data. Basically, after the T and S profiles for each τ are set, for the first 200 dbar the time of the year is taken into account and the correspondent values are added to the profiles. This process is used to evaluate if it is possible to detect mode water formation on the new time dependent profiles. Figure 4.13 ($\tilde{T}_{(p,t)}$, as referred to in section 3.4.2) is an example of the seasonal correction table used for T. The seasonal cycle is quite evident as the temperatures in Figure 4.14 there is not a raise in temperatures in the first half of each year and colder temperatures on the second half of the year. Figure 4.15 shows this seasonal variability with an amplitude of $\sim 20^{\circ}C$, as it was corrected using the data shown in Figure 4.13. The result resembles the seasonal cycle reproduced by Watts et al. (2001).

Figure 4.13: Seasonal correction field. Contours represent T ($^{\circ}C$)

We verified that the seasonal cycle was effectively restored to our profiles analysing the surface temperature with and without this correction. Figure 4.14 shows the data before and Figure 4.15 shows them after the correction. The result is quite evident, and the cycle was successfully restored.

Figure 4.14: Surface temperature time series at the PIES stations prior to seasonal correction.

Figure 4.15: Surface temperature time series at the PIES stations after seasonal correction.

The seasonal correction tables did succeed in restoring the seasonality of the profiles, but it was not enough to characterize typical STMW formation (Sato and Polito 2014; Provost et al. 1999) as it will be addressed in section 4.4. The choice for the reference level was considered as an explanation for this result, but when taking in account the changes of the reference level and how it altered the position of the 16°C isotherm (a temperature that Sato and Polito (2014) related to the formation process) so little, this leaves the conclusion that the best choice was actually the 200 dbar level based on the evidence present on figures 4.7 and 4.8.

4.4 Mode water formation

Subtropical mode waters have their formation window from the middle of winter until the end of spring. During that time the isotherms have a sudden shoaling due to intense convection. The subsurface layers tend to become very homogeneous and slightly cooler than during the subducting period of mode water, when it is considered fully formed. The goal of the seasonal correction of the GEM tables is to study the formation periods by making them detectable in the dataset. However, with the criteria used for mode water detection (minimum q and $\frac{\partial T}{\partial z}$) the results were not compatible with the formation phenomena. The limit that could be applied to all stations and eliminate seasonal and permanent thermoclines was $0.0225^{\circ}\text{Cm}^{-1}$. The $\frac{\partial T}{\partial z}$ on Figure 4.16 indicates that a value below this would take away some of the data that in fact contains STMW.

Any limit above this value would risk having thermocline waters on our data.

Figure 4.16: Mean $\frac{\partial T}{\partial z}$ profiles estimated using all PIES station data. Depth in meters (m).

Figure 4.17: Time series of PIES retrieved T profiles showing only data that had $\frac{\partial T}{\partial z} < 0.0225 \text{ } ^\circ\text{Cm}^{-1}$.

Figures 4.17, and 4.18 do not present outcropping of the typical isotherms of STMW. The patches of the seasonal thermocline start near December and end about the start of winter in the middle of the following year on the first 200 m. The permanent thermocline is the white patch without contours at about 400 m. The formation and dissipation of seasonal thermocline is visible but the outcropping of isotherms that indicate mode water formation is not present. The only restriction of temperature applied was that T should be less than 20°C and more than 10°C. This excluded values above the seasonal thermocline and below the permanent thermocline. Still, a $\frac{\partial T}{\partial z} = 0.0225 \text{ } ^\circ\text{Cm}^{-1}$ was more restrictive than the q criterium. This value was discussed on section 3.6, and is slightly less restrictive than observed in

the literature (Provost et al. 1999; Sato and Polito 2014). Considering all that, mode water formation was considered not present on the current dataset. This is observable by noting that the isotherms do not outcrop rapidly as winter starts and the q value for STMW during its formation should be even lower than fully formed STMW (Provost et al. 1999; Sato and Polito 2014). The limit for q was set as $1.5 \times 10^{-10} m^{-1} s^{-1}$. This is not a very restrictive value in comparison with other studies (Billheimer and Talley 2016; Sato and Polito 2014; Provost et al. 1999) for formed mode water.

Figure 4.18: Time series of T profiles showing only data that had $\frac{\partial T}{\partial z} < 0.0225 \text{ } ^\circ C m^{-1}$ and $q \leq 1.5 \times 10^{-10} m^{-1} s^{-1}$. The first 200 m have a q higher than expected for mode water. Below that depth the excluding factors were T limits and $\frac{\partial T}{\partial z}$.

We found reasonable to state that mode water formation is not present either in the original profiles using only the GEM-PIES method or after the seasonal

correction, at least for sites A, B and C. One possible explanation for this result is that the method neglects near-surface variations that are necessary to the STMW formation process. Once the profiles are weight average (Watts et al. 2001), this might have biased the results to profiles with characteristics closer to the water column state of periods outside the formation window. A second interpretation is that in fact there is no formation of mode water along the array region. To investigate the second hypothesis we examine the STMW distribution in ISAS data set. Figure 4.19 shows the number of times STMW was found in the surface on September between 2002 and 2014 using ISAS data (Bernardo et al.)(in prep.). As observed in Figure 4.19, the SAM array is located at the edge of the formation area (Figure 3.2). As the time series of T profiles for site D (Figure 4.18) shows rapid outcropping of isotherms that can be interpreted as a not so intense STMW formation because the q value for the first 200 dbar is too high to represent this process. Therefore, as the data from the PIES stations do not show a surface layer with characteristics of the STMW formation process, we chose not to use the seasonal correction on the data and kept only the standard GEM estimates.

Figure 4.19: Surface SASTMW occurrence number horizontal distribution map for September between 2002 and 2014

4.5 STMW Thickness

We have defined our parameters for mode water on section 3.6. After the determination of the vertical limits of STMW, its thickness was obtained. Because of the daily resolution of the PIES data the layer thickness was subject to high frequency variability. As we are interested in intraseasonal to interannual variability, the result was then filtered with a lowpass filter. Different cut-off period were considered. We tested for 2, 4, 6 and 12 months. This way only the signal with higher periods (or lower frequencies) remained. Here we present the results for the 13 months cut-off period, as the correlation with the BC transport for the remaining cut-off periods is later discussed on 4.6.

Each PIES started and ended at different time intervals, and some had long gaps without data. To keep the analysis coherent, only the largest time interval with valid data was considered.

Figure 4.20: Time series of STMW layer thickness anomaly with the seasonal cycle removed and low-pass filtered for 12 months.

Fully formed mode water was detected in all PIES stations, and its layer thickness was calculated. This result was expected as seen on Figure 4.21. Figure 4.21 shows the mean thickness of subsurface (formed) STMW using ISAS data. The area occupied by the SAM array is inside the detection area of STMW (Bernardo et al.)(in prep.). Table 4.1 lists the calculated thickness for the SASTMW layer at each PIES mooring. There is no statistical difference between the values found at each PIES station. As seen in Figure 4.20 the variability of the signal shows that sites A and D kept a consistent pattern, although D has a slight reduction in amplitude after 2012. The changes before 2012 range from about 60 m, and after that period the range drops to about 30 m. Site B has an increased change rate starting on 2014, going from about 20 m to about 40 m after 2014. Site C shows

Table 4.1: SASTMW thickness (m) at each SAM site.

	PIES A	PIES B	PIES C	PIES D
Layer thickness (m)	202 ± 49	276 ± 47	176 ± 78	231 ± 46

a slight thinning (about 40 m) from the end of 2013 until the end of the series in 2015. The time series of STMW layer thickness from all PIES stations seem to show independent variabilities.

Figure 4.21: SASTMW subsurface layer horizontal monthly distribution maps. Colors indicate mean thickness from 2002 to 2014 using ISAS dataset. (Bernardo et al.)(in prep.)

4.6 BC Volume Transport

Applying the method described in section 3.7 we calculated a time series of volume transport for the BC. To keep the analysis consistent with the STMW layer thickness, the volume transport was subject to the same data processing: a sinusoidal curve was fit using the least square method and this was subtracted from the original signal before the lowpass filter was applied. The removal of the sinusoidal fit served as a means to eliminate covariations that did not reflect a direct relation between

the layer thickness and the BC transport only because the two are subject to the same atmospheric and astronomic forces that cause seasonal variation. Our interest lies on interannual variability, so this procedure is required to eliminate complexity on the interpretation of the results.

The average transport for the BC was -3.78 ± 3.38 Sv. Figure 4.22 shows the original time series of transport and its anomaly after removing the seasonal cycle and applying the lowpass filter with a cut-off period of 13 months, so that any signal with smaller period was removed.

Figure 4.22: BC transport (SV) from HYCOM outputs, integrated from the surface to the $\sigma_1 = 32.2 \text{ kg m}^{-3}$. The original transport is shown in blue and its anomaly after the removal of annual cycle and low pass filtered is shown in black.

4.7 BC and STMW Layer Thickness Covariance

STMW layer thickness and its covariance with BC transport was calculated. We discussed before that we focus on the interannual covariance, but the lowpass filter was also calculated with different cut-off periods of 2, 4 and 6 months to observe the effect of the low-pass filter on the data and evaluate if there were relevant covariances of processes with periods inferior to the annual scale.

On average, the BC volume transport is southward, that is, it has a negative mean. Therefore the covariance was calculated with the signal for transport inverted, to simplify interpretation. As the BC flows southward this means any intensification on the current will result in values more negative. With this change of signs, when the current is more intense, the values become larger. Layer thickness is always positive, meaning that when the values diminish the layer is thinner. We are looking for negative correlations to test our hypothesis. If the strongest correlation with minimum lag is negative then this means that, as the BC intensifies the STMW responds by getting thinner.

Figure 4.23: Covariance with lag between BC transport and STMW thickness. Positive lag means BC transport leads. Seasonal cycle was subtracted after fitting with least squares method and a lowpass filter was applied (2 months cut-off period). Blue line is the covariance and the red line represents the statistical significance of each covariance coefficient.

With covariances calculated we had to evaluate its statistical significance. For that matter we used Monte Carlo simulations. These simulations consisted of creating random data that mimic the standard deviation of the PIES STMW layer thickness time series as well as the BC transport. Following that, we removed a fit sinusoidal curve (using the least squares method) corresponding to the annual cycle and applied the same low-pass filter. At this point the random data has gone through the same analyses as the BC transport and STMW layer thickness data. Then, the randomly generated correspondents to BC transport and STMW layer thickness had their covariance calculated. We calculated the covariances of 10000 simulations, and

calculated the percentage of times that the absolute value of the covariance found was equal or greater than the original data: the p-value. The red lines (figures 4.23 to 4.26) correspond to the p-value (1 minus the statistical significance level) calculated using Monte Carlo simulations, and the blue lines correspond to the covariance index. From all simulations the indexes never exceeded 0.5, regardless of the lag or the cut-off period used. In contrast, some of the values have relative high statistical significance. The statistical significance drops to 20% or below with covariance coefficients equal or lower to 0.2. We focus on positive lags when the transport leads and the changes in thickness come after. We present the most significant negative peaks on covariance for each cut-off period used and each PIES array site.

Figure 4.23 is result of covariance calculation with lag using a cut-off period of 2 months. The first relevant positive peak for the A site corresponds to roughly 4 months (114 days) of lag with a negative covariance of -10% between the BC transport (leading) and the STMW thickness, on a 77% significance level. For the B site, a covariance of -16% with a lag of 6 days is most significant with 80% statistical relevance. Site C has -27% covariance at ~ 7 months of lag (216 days) with 99% significance. Site D shows a peak with -31% covariance at 6 days with 99% significance.

Figure 4.24: Similar to Figure 4.23, but with cut-off period of 4 months.

For the 4 months cut-off period (Figure 4.24), site A has -7% covariance with 3.6 months (108 days) of delay with 38% significance. Site B has a peak at 18 days, corresponding to -17% covariance significant at the 66% level. Site C shows a peak at 7.2 months (216 days) a covariance of -22% significant at the 85% level. Site D shows a peak at 18 days a covariance of -45% significant at the 99% level.

Figure 4.25: Similar to Figure 4.23, but with cut-off period of 6 months.

For the 6 months cut-off period (Figure 4.25), site A has no relevant peaks. Site B has relevant covariance at 18 days of -10% significant at the 47% level. Site C shows a peak at 7.2 months (216 days) a covariance of -15% significant at the 45% level. Site D shows a peak at 18 days a covariance of -56% significant at the 99% level.

Figure 4.26: Similar to Figure 4.23, but with cut-off period of 13 months.

For the 13 months cut-off period (Figure 4.25), sites A B and C show no negative covariances. Site D shows a peak at 30 days a covariance of -37% significant at the 73% level.

Testing different cut-off periods yielded a few relevant covariances between th BC tranport and the SASTMW thickness, but the 12 months period restricts the signals to only interannual or longer period changes. We won't discuss correlations with lags greater than one year. No significant covariances were found except on site D, but all of them for signals with cut-off limits below 6 months. Apparently, covariances on an interannual scale between the BC transport and the STMW layer thickness are not present or statistically relevant.

We expected to see some coherence between the correlations found in all four sites, but as it seems, independent of the cut-off period, the sites show very different

behaviours. Not even A and B sites, that are on each border of the BC show a similar pattern, or covariance indexes. The low covariances and low statistical significance levels on an interannual scale are evidence that our hypothesis was negated.

5. Conclusion

Subtropical mode waters are very homogeneous layers with low $\frac{\partial T}{\partial z}$ and q trapped between the seasonal and permanent thermoclines, formed near fronts or boundary currents. The SASTMW forms in the South Atlantic subtropical gyre, influenced by the Subtropical Front and the BC. As SASTMW is identified by its hydrographic characteristics, in this study we used GEM estimatives for T and S profiles derived from τ data from PIES, part of the SAM project. The T and S information were used to identify the SASTMW, and determine its layer thickness. We found that the GEM estimates do not reproduced the formation process, even with a seasonal correction for the first 200 dbar. The SASTMW layer thickness and the BC volume transport had their covariance calculated to determine if the processes are connected what is the time lag of the response of SASTMW layer thickness to BC volume transport variability on an interannual timescale, although lower time scales were also evaluated, but not the main goal of this study.

We assessed many issues regarding the methodology of the seasonal correction in this study, the reference level for the correction, the detection and expression of SASTMW formation. Here we discuss the implications and interpretations of our main results as well as some remarks about the seasonal corrections.

- The WOA13 long term means were used to analyse when compared with the mean variability of the T and S time series made evident the lack of seasonality in the GEM derived profiles. This motivated the use of the seasonal correction in order to analyse SASTMW formation.

- When analysing the seasonal correction the determination of a reference level for pressure was very significant, and a strong justification for the choice was necessary.
- After careful observation of the WOA13 seasonal signal the first choice was 200 dbar, but there was still a reasonable uncertainty about how this would affect the profiles. To answer this question we observed the 16°C isotherm and the resulting data showed that no significant changes affected its position in the water column. In conclusion, the 200 dbar level of reference was satisfactory.
- The detection of SASTMW formation showed that the process was not identifiable in the dataset. As discussed, that can be an artifact of the methodology employed or an actual agreement with Bernardo et al. (in prep.). The decision to not consider the seasonal correction in the final results came after observation of the effects of the identification criteria. Changes in the first 200 dbar would not affect detection of formed SASTMW or its seasonal variability, leaving no reason to keep the correction. There was no strong evidence to keep considering the correction once the temperatures observed at the only station that showed a process similar to SASTMW formation had $\frac{\partial T}{\partial z}$ too high when compared to Sato and Polito (2014).
- The SASTMW thickness temporal variability showed no single pattern on all array sites even though the values are coherent and not statistically different. The mode water layer detected was about 220 m $\pm$ 55 m thick on all sites, agreeing with Sato and Polito (2014, Provost et al. (1999). But on contrast we observed a greater temperature stratification. Billheimer and Talley (2016) were able to detect EDW using $\frac{\partial T}{\partial z} = 0.006 \text{ } ^\circ\text{Cm}^{-1}$, much lower than our

$0.0225\text{ }^{\circ}\text{Cm}^{-1}$. Preliminary results of Bernardo et al. (in prep.) suggest this might be because the SAM array lies on the border of the SASTMW formation area and beyond that other processes such as advection and instability related to the BC can disturb the upper layer homogeneity.

- Lastly, while there is some covariance between the BC variability and SASTMW thickness on sites C and D, it was not very relevant on interannual timescales. The BC led the changes in mode water thickness with a delay of a month on site D, with -37% covariance significant to the 73% level. Site D showed covariances of -31%, -45% and -56% when using cut-off periods of 2, 4 and 6 months respectively. From this we can infer that there might be other processes that influence the behaviour of formed SASTMW on an interannual scale. Qiu and Chen (2006) found that the ocean influenced NASTMW formation via preconditioning. NATSMW had a strong correlation with the KE jet but on a decadal scale. Qiu et al. (2007) states that stable EKE states favored mode water formation. The analysis of the BC stability and EKE on SASTMW formation done by Sato and Polito (2014) showed that this is not the case on the South Atlantic, since the correlations found were small. Further studies with SASTMW and the BC with longer datasets could address those issues more thoroughly.

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