Universidade de São Paulo
Instituto Oceanográfico

Igor Uchôa Farias

Eddy-Induced Ekman Pumping in the Vitória-Trindade Ridge Region

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

Advisor:

Prof. Dr. Ilson C. A. da Silveira.

Co-advisor:

Prof. Dr. Paulo Henrique Rezende Calil.

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"The past is just a story we keep telling ourselves."

JONZE, Spike; *Her* (2013)

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List of Abbreviations

AB	Abrolhos Bank
AE	Abrolhos Eddy
BC	Brazil Current
BCS	Brazil Current System
EKE	Eddy Kinetic Energy
EWI	Eddy-Wind Interaction
GOGRO	Global Ocean Gridded Reprocessed Observations
ITE	Intrathermocline Eddy
IWBC	Intermediate Western Boundary Current
MKE	Mean Kinetic Energy
MLD	Mixed Layer Depth
OW	Okubo-Weiss Method/Parameter
ROMS	Regional Oceanic Modelling System
RCE	Royal Charlotte Eddy
RCB	Royal Charlotte Bank
RMS	Root Mean Square
SEC	South Equatorial Current
SLA	Sea Level Anomaly
SODA	Simple Ocean Data Assimilation
SSH	Sea Surface Height
TB	Tubarão Bight
VE	Vitória Eddy
VTR	Vitória-Trindade Ridge
WA	Winding-Angle Method

Abstract

The transport of rich-nutrient waters to the euphotic zone is a key process in controlling primary production in the ocean. One of the most important features that are related to the boosting/damping rates of chlorophyll A in the upper ocean are eddies due to its isopycnal displacements adjustments. In this present study, we investigated the input of eddy-induced Ekman pumping in the density structure of two main mesoscale eddies in the Vitória-Trindade vicinity: Abrolhos Eddy (anticyclonic) and Vitória Eddy (cyclonic). Using an air bulk formula for wind stress that accounts for ocean current feedback, eddy-induced Ekman pumping may be described as the sum of two terms: linear Ekman pumping (Eddy-Wind Interaction) and non-linear Ekman pumping. A three-stage analysis was conducted consisting of (i) statistical characterization of the mesoscale eddies, (ii) idealized parametrization of Ekman pumping using a homogeneous wind field, and (iii) inferences on eddy-induced Ekman pumping in mesoscale eddies isopycnal displacements. Climatological eddy structures indicated total Ekman pumping values higher than 0.1 m day^{-1} for Abrolhos Eddy and lower than -0.1 m day^{-1} for Vitória Eddy at the eddy center, though those values may decreased if averaged in an azimuthal sense. Isopycnal displacements seem to be correlated with the horizontal distribution of both components of Ekman pumping. Abrolhos Eddy showed a mode-water configuration related to Eddy-Wind Interaction. Vitória Eddy showed a thinny structure that seemed to be related to non-linear Ekman pumping values, yielding -1 m day^{-1} with similar variation and magnitude to the vertical velocities of the eddy structure.

Keywords: Brazil Current, Eddy, Eddy-Wind Interaction, Non-linear Ekman Pumping.

Resumo

O transporte de águas ricas em nutrientes à camada eufótica é processo chave no controle da produção primária no oceano. Uma das feições mais intrinsecamente relacionadas ao aumento/diminuição de taxas de clorofila A no oceano superior são os vórtices devido seus ajustamentos isopícnais na coluna d'água. Esse estudo tem como foco investigar o papel do bombeamento de Ekman acoplado à vórtices na estrutura vertical de densidade de dois dos principais vórtices de mesoscala na região da Cadeia Vitória-Trindade: o Vórtice de Abrolhos (anticiclônico) e o Vórtice de Vitória (ciclônico). Usando a fórmula de tensão de cisalhamento do vento acoplada ao oceano, bombeamento de Ekman pode ser dividido em dois termos: o bombeamento linear de Ekman (ou efeito vento-vórtice) e o bombeamento não linear de Ekman. Foi desenvolvido nesse trabalho uma análise que se constituiu de três etapas: (i) caracterização estatística dos vórtices, (ii) parametrização do bombeamento de Ekman usando um campo de vento homogêneo e (iii) inferências no bombeamento de Ekman acoplado à vórtices de mesoescala nos ajustamentos isopícnais da coluna d'água. A climatologia das estruturas de velocidade dos vórtices indicou valores maiores que 0.1 m dia^{-1} para o Vórtice de Abrolhos e -0.1 m dia^{-1} para o Vórtice de Vitória em seus respectivos centros. Porém esses valores podem diminuir se estabelecermos curvas médias ao longo dos raios de cada vórtice. Ajustamento de isopícnais para a formação de vórtices intratermoclínicos parece estar correlacionado com as distribuições horizontais do bombeamento de Ekman. Vórtice de Abrolhos (Vitória) apresentou configuração de vórtice água modal (*thinny*) devido principalmente aos padrões de bombeamento de Ekman linear (não linear). Vórtice de Vitória apresentou valores de bombeamento não linear de até -1 m dia^{-1} com variações e magnitude similar as velocidades verticais da estrutura do vórtice.

Palavras-chave: Corrente do Brasil, Vórtice, Efeito Vento-Vórtice, Bombeamento de Ekman Não Linear.

Chapter 1

Introduction

1.1 Physical-Biological Processes

Phytoplankton communities and other autotrophic marine organisms are responsible for the ecologic role of primary production in the open and coastal oceans. Primary production is the foundation of vital biogeochemical cycles in the global ocean. Unraveling how primary production affects the upper ocean ecosystem is essential if we are to analyze on how the oceanic ecosystems behave and survive as well as organic carbon uptake/fixation is developed at the air-sea boundary layer (McGillicuddy et al., 2007).

Spatial distribution and temporal fluctuations of chlorophyll A, the most well-studied proxy to analyze primary production, are the main focus of several works aiming to understand life in the ocean. Besides observational analysis, the synoptic-data retrieval of satellite imagery and the capability of reproducing highly-complex phenomena by numeric simulations have become the most reliable approaches to analyze and understand primary production (McGillicuddy Jr, 2016). However, in order to understand and model the dynamics of primary production we shall also analyze the hydrodynamics of the ocean, since they are deeply intertwined (Flierl and McGillicuddy, 2002).

The transport of rich-nutrient waters to the euphotic zone is a key process in controlling primary production in the ocean. The most important features that are related to the boosting and damping of chlorophyll A in the ocean directly are eddies and fronts. Studies about the roles of those mesoscale and submesoscale features on primary production have been increasing since the pioneer works of Gower et al. (1980), Jenkins (1988), and Nelson et al. (1989). These works explored vertical nitrate flux and phytoplankton growth in mesoscale eddies that highlighted their importance.

Mesoscale eddies are characterized as a baroclinic flow and are originated by either vorticity waves or instabilities in the geostrophic flow (Gill et al., 1974). These features are typically dozens of kilometers in diameter and temporal scales of weeks to months; Therefore, these eddies are larger than the first baroclinic radius (Chelton et al., 1998)

and longer than the local inertial period. However, submesoscale processes, intrinsically associated to these features, are smaller and short-living. Mesoscale processes are predominantly geostrophic, horizontally non-divergent, and hydrostatic; whereas submesoscale occupies the upper ocean, are mostly non-hydrostatic and strongly non-linear. Moreover, submesoscale processes are intimately related to the energy transfer by larger scale process of the ocean and/or the atmosphere (Lévy et al., 2012).

1.2 Mesoscale Features Structures

One may characterize mesoscale eddies according to their polarization. Those first baroclinic mode flows may be characterized as cyclones and anticyclones. At first order, cyclonic (anticyclonic) eddies are divergent (convergent) flows that generate positive (negative) vertical velocities. Regarding primary production, one can infer that vertical velocities in cyclonic (anticyclonic) eddies may ascend (descend) isopycnals in the water column, transporting water of nutrient-rich (nutrient-depleted) water in nutrients towards the surface (bottom) layers of the eddy. This mechanism may boost (weaken) the primary production inside the vortex for a cyclone (anticyclone), respectively (Gill et al., 1974; McGillicuddy Jr, 2016).

However, according to McGillicuddy Jr (2015), more complex mesoscale eddies are recently found and described. These eddies are second baroclinic mode structures that are characterized by a seasonal thermocline displacement that opposes its permanent thermocline depression/uplift, the latter being ruled by the polarization of the eddy. Those eddies are denominated intrathermocline eddies. In respect of the latter eddies, the displacement of the main thermocline dominates geostrophic velocity and sea level perturbation (η) (Flierl and McGillicuddy, 2002; McGillicuddy Jr, 2015).

The first description of intrathermocline lenses in eddies was the mode-water eddy feature, which, as the regular anticyclonic eddy (McGillicuddy et al., 1999), has a positive amplitude (sea level perturbation) and descendent displacement of the main thermocline. Nevertheless, its seasonal thermocline is steepened, consisting of a thick lens of fluid between both isopycnals. The analogous feature for a cyclone was further classified as "thinny" eddy. In this case, the eddy has a downward displacement of both amplitude and seasonal thermocline; and a positive displacement of the main thermocline. The representation of all four types of eddies are shown in Figure 1.1.

These features have been receiving more attention due to the fact that their structures differ from the regular mesoscale eddies in terms of primary production. Since mode-water eddies have an upward vertical displacement of the seasonal thermocline; nutrient-rich water may be injected in the euphotic zone favoring production. The opposite may occur in the cyclonic eddies, where a decrease of primary production might be observed. Intrathermocline eddies are defined as decay transformations of regular eddies (Figure 1.1)

Figure 1.1: Hypothesized transformations of first baroclinic mode eddies into their second baroclinic mode analogous according to McGillicuddy Jr (2015)

(Flierl and McGillicuddy, 2002; McGillicuddy Jr, 2015); although, the mechanisms of decay for those eddies are not fully understood, an important number of works have pointed out that the main factor for this process is due to the wind regime acting in the surface of those features.

1.3 Eddy-Wind Interaction

The studies of Martin and Richards (2001), McGillicuddy et al. (2007), and McGillicuddy Jr (2015) state that the formation of intrathermocline lenses in eddies are due to the interaction between the wind stress input and the swirling velocity of the eddy. As a matter of fact, Stern (1965) was the first to investigate the interaction of a vortex with Ekman flow in a nonlinear system on theoretical framework.

In order to understand the eddy-wind interaction problem, let us consider a spatially uniform constant wind over an anticyclonic eddy in the southern hemisphere. Not only the effect that the wind's energy input causes on the ocean but also the effect of current's energy input causes on the atmosphere are considered in this computation. This latter assumption is not neglected due to the importance of mechanical damping of the currents in eddies described by various authors (Dewar and Flierl, 1987; Eden and Dietze, 2009; Renault et al., 2016). In order to simplify the problem, let us analyze two opposites flanks of an eddy with swirling velocities directions parallel to a constant wind field. The flank

with same velocity direction as wind field is under less frictional stress than the symmetrical opposite flank with an opposite velocity direction of the wind field. Therefore, there is a gradient of Ekman transport between both opposite flanks of the eddy. This horizontal shear of Ekman transport over the eddy generates, through horizontal divergence, positive vertical velocities that oppose the ageostrophic negative vertical velocities caused by the mesoscale anticyclonic eddy. In Figure 1.2, the representation of this mechanism may be observed.

Regarding primary production response, through eddy-wind Interaction, it is theoretically possible to observe relatively high primary production rates in anticyclones. The same mechanism can be analogously described for cyclones. The horizontal shear of Ekman transport is also observed except with opposite sign as found at the former case. That generates negative vertical velocities that sink surface water to deeper regions of the eddy. Hence, the phytoplankton is advected away from the euphotic zone provoking a decrease of primary production on the eddy due to either mortality or nutrient-limiting growth.

1.4 Eddy-Induced Ekman Pumping

In the work of Martin and Richards (2001), the dynamic and morphometric parameters of an anticyclonic eddy at the Iceland Basin on a Lagrangian survey were described. The preliminary analysis of this eddy showed that, despite its polarization, it received an enhanced flux of nitrate from depth Figure 1.3. The author attributed the vertical velocity balance to two factors: ageostrophic upwelling, caused by non-linearities in the geostrophic flow and the β plane effect; and wind forcing, due to interaction between the eddy and the wind field causing differential Ekman transport in the field.

McGillicuddy et al. (2007) have also investigated the effect of wind over eddies. Analyzing a number of eddies that were detached from the Gulf Current as well-formed rings, the author analyzed the isopycnal displacements and chlorophyll A on the eddies. Mode-water eddies presented the highest values of chlorophyll A, more than the first baroclinic mode cyclones studied. The author also addressed eddy-wind interaction on the Ekman pumping balance.

Furthermore, there has been some discussion on whether the linear Ekman pumping cause by the interaction between the wind and the ocean is, in fact, the main component of the Ekman-layer vertical velocity balance. Mahadevan et al. (2008) stated that non-linear Ekman pumping and other mesoscale processes may be the most important components in the balance. Those submesoscale processes are relevant for flows with relative vorticity (ζ) of the same order as the planetary vorticity (f) Figure 1.4 and 1.5. In addition, time scale plays an essential role in the vertical velocity values within the eddy in terms of submesoscale. If we are to take into account submesoscale process within the domain,

Figure 1.2: Representation of how an uniform wind field on the Southern Hemisphere dynamically interact to an anticyclonic eddy can cause divergence on Ekman transport that may lead to upwelling in the eddy interior.

relative vorticity of both wind and current, parameters that comprise non-linear Ekman pumping, are dependent on the duration of the feature which may yield vertical velocities 10 to 100 times higher as the ones expected by linear Ekman pumping.

However, patchiness of vertical velocities in the eddy center are higher and due to eddy-wind interaction input on linear Ekman pumping. That is, submesoscale processes are located in the periphery of the eddy and its azimuthal averaged values are not as intense as the eddy-wind interaction induced ones (McGillicuddy et al., 2008).

Therefore, in order to analyze the influences of wind stress in the ocean's interior through Ekman pumping; let us understand the Ekman Pumping analytic model. Stern (1965) has denoted the Ekman transport modified by relative vorticity (ζ) in order to take the non-linear advective terms of the Ekman dynamics into account. Hence, the total Ekman Pumping is

Figure 1.3: Potential density anomaly section from CTD stations of an modal-eddy according to Martin and Richards (2001)

Figure 1.4: Representation of non-linear Ekman pumping over an anticyclonic eddy in the northern hemisphere. This effect generates an upwelling and a downwelling lobe due to the relative vorticity coupled with the wind. The Ekman transport horizontal shear occurs due to the decay of velocity towards the center of the eddy according to Mahadevan et al. (2008)

Figure 1.5: Normalized curves of a gaussian anticyclonic eddy for relative vorticity (ζ), zonal velocity, and non-linear Ekman pumping.

$$W_{Total} = \frac{1}{\rho_o} (\nabla \times \frac{\tau}{f + \zeta}), \quad (1.1)$$

where the mean density of seawater is $\rho_o = 1024 \text{ kg m}^{-3}$, the Coriolis parameter for latitude θ and rotation parameter $\Omega = 7.29 \times 10^{-5} \text{ s}^{-1}$ is $f = 2\Omega \cos \theta$; and the surface wind stress with meridional (τ^y) and zonal (τ^x) components is τ . As τ and ζ are horizontally dependent, the chain rule applies to this expression. By operating the curl on Equation 1.1, it becomes

$$W_{Total} = \frac{(\nabla \times \tau)}{\rho_o(f + \zeta)} + \frac{1}{\rho_o(f + \zeta)^2} (\tau^x \frac{\partial \zeta}{\partial y} - \tau^y \frac{\partial \zeta}{\partial x}). \quad (1.2)$$

The first term of Equation 1.2, resulting from the vertical curl of the wind stress is often denominated as linear Ekman pumping (W_τ) (McGillicuddy et al., 2008; Gaube et al., 2015). This term contributes to Ekman pumping responses from the direct interactions between the eddy dynamics and the wind field and it is the term of the Equation 1.2 which accounts for the EWI. On the other hand, the second term of the equation results from derivatives of ζ and the wind stress terms over the squared vorticity ($f + \zeta$). That is, vorticity-induced Ekman pumping. This term is often referred as non-linear Ekman pumping (w_ζ) (McGillicuddy et al., 2008; Gaube et al., 2015). The latter contribution to the W_{total} is the one described in Mahadevan et al. (2008).

As mentioned in section 1.3, the assumption of mechanical damping due to the current

speed on the wind stress computed is considered in this model since we are investigating the EW. Hence, for the surface wind stress τ , we have chosen the simplified bulk aerodynamic approximation described by Bye (1986):

$$\tau = \rho_{air} C_D |U_a - U_o| (U_a - U_o), \quad (1.3)$$

where $\rho_{air} = 1.2 \text{ kg m}^{-3}$ is the mean air density, C_D is the dragging coefficient, and U_a and U_o are the magnitude of surface winds and of ocean currents, respectively. This subtraction is also referred as relative wind magnitude (U_{rel}) (Gaube et al., 2015).

Considering this analysis, it is possible to quantify and characterize the main influences of Eddy-induced Ekman pumping in the upper ocean. It is important to highlight that the observations regarding EW are narrowed to the North Atlantic Ocean and to fully developed rings derived from the Gulf Stream (McGillicuddy et al., 2007; Martin and Richards, 2001). A global inventory with altimetry-derived composite data as well as scatterometer-derived, as in Gaube et al. (2015), was employed. However, the authors focused solely on specific regions of intense eddy formation not having addressed on the vertical eddy structures.

1.5 Brazil Current Eddies and the Vitória-Trindade Ridge

The Brazil Current (BC) is a western boundary current that flows southward along the Brazil continental margin and is generated through the bifurcation of the South Equatorial Current (SEC). That is, BC is the boundary current which closes the western flank of the South Atlantic Subtropical Gyre. (Peterson and Stramma, 1991) inferred the relatively lower transport of the current compared to other western boundary currents because three quarters of the impinging SEC at about 15°S enters the North Brazil Undercurrent.

The mean state of the BC is not simple to describe, specially nearby the latitudes it borders complex topography. Soutelino et al. (2011) characterized the BC at the latitudes of 10°S - 20°S as eddy-dominated and its poleward flow as an interaction with mesoscale eddies that are recurrent along the shelf break of Brazil shelf. Although some works have studied the mesoscale activity of the BC (Silveira et al., 2008; Calado et al., 2010; Soutelino et al., 2011), the mesoscale features of BC system in this latitude range have not been well characterized in terms of recurrence, morphometry, and intensity.

At 17°S and 19°S two mesoscale anticyclones have been described by Soutelino et al. (2011) and named as Royal Charlotte Eddy and Abrolhos Eddy (AE). Those two eddies are named due to the proximity of Royal-Charlotte Bank and Abrolhos Bank as the region of formation of the mesoscale features. South of 20°S , a cyclonic eddy firstly described by Schmid et al. (1995) is recurrent and its formation region is located South of the Tubarão Bight (Figure 1.6). Its formation to date seems to be related to the BC crossing of the

Vitória-Trindade Ridge (VTR).

Figure 1.6: Averaged stream function field from geostrophic velocities of the region retrieved by CMEMS altimetry dataset. Abrolhos Eddy (AE) and Vitória Eddy (VE) are schematized in white arrows. The black labels indicate the topographic features: Abrolhos Bank (AB), Tubarão Bight (TB), and Vitória-Trindade Ridge (VTR).

The VTR is a quasi-zonal seamount chain located at 20°S and it is the encounter region of the Brazil Current (BC) and the Intermediate Western Boundary Current (IWBC) also known as the Brazil Current System. This geologic feature is comprised by approximately 30 seamounts from which 17 reach an elevation of over 2500m. In Figure 1.7, we can observe the features of the VTR.

Due to the high complexity of the VTR and the uncommon characteristics of the BC as a western boundary current, the mesoscale activity of this region is not simple to understand. The recurrent eddies in the region may be interpreted as a result of a flow driven by the Brazil Current instability and highly complex topography (Soutelino et al., 2011). Numerical simulations have also contributed to understand the dynamics of the Brazil Current and the eddy formation from the western boundary current through the influence of local topography (Soutelino et al., 2013).

The VTR region wind regime is driven by South Atlantic Subtropical Gyre. The atmospheric feature migrates at winter when it presents its most important interannual variability (Sun et al., 2017). Stech (1990) characterized the surface wind field in the South Brazil Bight as a frontal system predominant region. Those atmospheric fronts occur over periods of 6 to 11 days and may influence on the flow, specially in the continental margin.

The mean wind regime in the study site is considered, however, as low seasonal fluctuation and it is characterized for being trade wind predominant, with eastern/southeastern winds throughout the year yielded by the large-scale circulation cells as it is presented in

Figure 1.7: The study site and its topography. The acronyms for the submersed banks are: Besnard Bank(BSB), Vitória Bank (VTB), Congress Bank (CGB), Montague Bank(MTB),Jaseur Bank(JSB), Columbia Bank(CLB), Davis Bank (DVB),and Dogaressa Bank (DGB). The acronyms for the seamounts are: Champalim Seamount (CPS) and Columbia Seamount (CLS).

climatologic wind field and the wind rose in Figure 1.8.

1.6 Motivation, Hypothesis, and Scientific Questions

Based on the literature review and introductory theory of the previous section, one may observe that the conformation of the surface winds and the recurrent formation of AE and VE at the Brazil continental margin between 18-21°S is a *première* site for the EWI studies as well as eddy-induced Ekman pumping in general. To the author's knowledge, there is no work regarding about such dynamics so far in the region.

It is important to highlight that there still lacks studies in respect of eddy-induced Ekman pumping in the global ocean, let alone studies with non-propagating eddies, eddies situated within western boundary layer, and eddies formed due to current interaction with complex topography as VTR.

Although the enrichment of phytoplankton biomass has already been studied in VE

Figure 1.8: Climatological windrose retrieved by entire QuikSCAT dataset for the study region (left) and climatological annual field of 10m winds from European Centre for Medium-Range Weather Forecasts or ERA-interim (1980 - 2010) (right).

(Gaeta et al., 1999), little is known about the processes that may result in the fluctuation of phytoplankton within the cyclone. In addition, studies involving the AE and its likely potential for increasing the primary production as well as processes that may influence in the eddy have not been reported so far.

Thus, we may formulate our scientific hypothesis:

"The wind conformation and duration of the mesoscale eddies adjacent to the Vitória-Trindade Ridge generate important vertical velocities and isopycnal displacements to generate intrathermocline lenses via eddy-induced Ekman Pumping, mainly by linear Ekman pumping (Eddy-Wind Interaction)".

Based on this primary hypothesis, we add two supplementary hypotheses regarding production and the non-linearity of Ekman Pumping. The first supplementary hypothesis relies on the recurrence and duration of the eddy-wind interaction to alter the eddy isopycnal configuration. The second hypothesis assumes that the linear Ekman pumping has a more effective influence on the eddy center than submesoscale processes as non-linear Ekman pumping. The two supplementary hypothesis to be tested are:

- Eddy-induced Ekman pumping is intense and recurrent sufficiently to configure the isopycnal displacement of both eddies as intrathermocline lenses.
- Vertical velocities fluctuations in the Ekman Layer depth are governed by linear Ekman pumping rather than non-linear Ekman Pumping.

1.7 Objectives

The main objective of this work is to investigate the relevance of eddy-induced Ekman pumping in the Abrolhos and Vitória eddies with respect to vertical velocity balance and density displacement within the eddy boundaries. Hence, in order to achieve this objective, we defined the following specific objectives.

- To perform a statistical characterization of mesoscale eddies from altimetry data.
- To parameterize the horizontal distribution and azimuthal mean of the eddy-induced Ekman pumping in both eddies as idealized mesoscale features using the statistical characterization.
- To examine the isopycnal displacements of numerically simulated eddies with two distinct wind stress forcing formulations.
- To analyze the fluctuations of vertical velocity and its relation with Ekman pumping in the AE and VE within the eddy boundaries for aforementioned output sets.

Chapter 2

Datasets

2.1 Altimetry Data

In the present work, we used altimetry data in order to analyze the mean geostrophic speed of CB system, the morphometry and frequency of the eddies associated to the VTR crossing, as well as the signatures of the current feedback on surface wind stress on the VTR surroundings through 24 years.

Due to coarse time resolution of altimeters, an efficient option to obtain finer resolution geostrophic velocities data is using combined data from different satellite images. The altimetry dataset used for the analysis is called *Global Ocean Gridded Reprocessed Observations* (GOGRO) acquired by CMEMS (Copernicus Marine Service Information) and it is comprised by a multimission assimilated and reprocessed satellite data. The altimetry product variables are surface height anomalies and derived variables such as zonal and meridional geostrophic velocities. It is a processed dataset from the satellite products of Jason-3, Sentinel-3A, HY-2A, Saral/AltiKa, Cryosat-2, Jason-2, Jason-1, T/P, ENVISAT, GFO, and ERS1/2.

For the data processing, Geophysical Data Records are used by a Precise Orbit Ephemeris (POE) which are assimilated around every 2 months. There is a quality data control, assuring that the best altimetry data is being acquired, and a cross-calibration process, removing any residual orbit or long wavelength error. The final step to the process is the calculation of a Mean Sea Surface (MSS) that is deducted to the Dynamic Topography in order to generate the Sea Level Anomalies (SLA) and derived products.

The reprocessed dataset has a time range from 1993 to current date and it provides a daily $0.25^\circ \times 0.25^\circ$ interpolated and gridded synoptic observation data. In order to dynamically improve the quality of data, values up to -200 m of bathymetry were neglected and shallow areas ($H < 1000\text{m}$) were not focused on our analysis, since the eddies are actively at deeper regions of the VTR region.

2.2 Global Model

In order to obtain a climatological transport of the BC in the study region, we used the global ocean reanalysis MERCATOR Ocean GLORYS2V4. The Mercator Ocean is performed with NEMOv3.1 ocean model in configuration ORCA025-LIM. This global reanalysis is generated and developed by CMES. It is characterized as a $2.5^\circ \times 2.5^\circ$ gridded resolution, with a vertical grid of 75 levels that are set by partial steps of the bottom of the ocean. Regarding the data assimilation procedures in this global simulation, ERA-Interim reanalysis products drive the surface of the model. Assimilated observation are satellite sea surface temperature (SST) from daily AVHRR products and sea level anomaly (SLA) data from CMES product compilation. In addition, *in situ* profiles of temperature and salinity from CORA 4.1 provided by CMES are also assimilated.

This global model was selected to compute the climatological fluctuations of the BC transport due to its heavy assimilation from *in situ* data and satellite imagery. In addition, the geostrophic velocities assimilated in the reanalysis are from the altimetry data aforementioned on section 2.1. Hence, the need for a prior detailed confirmation of the BC transport is virtually purposeless.

2.3 Regional Model

The two simulations we used to evaluate the effect of the current feedback on the wind surface stress on the VTR region are from the Regional Oceanic Modelling System, or simply ROMS, maintained by IRD (*Institut de Recherche pour le développement*) and INRIA (*Inventeurs Du Monde Numérique*).

The version of the ROMS used in this work is the "Coastal and Regional Ocean Community (CROCO)". The numerical model solves the primitives equations of momentum in a rotating system with the Boussinesq approximation, hydrostatic balance, and incompressibility of fluids. The equations are discretized in horizontal curvilinear coordinates in a ARAKAWA-C grid type for horizontal components, whereas the vertical coordinates are terrain-following set for a finer resolution in the upper ocean (Shchepetkin and McWilliams, 2005).

In order to represent both mesoscale and submesoscale processes in the region of VTR, the output of the numeric simulation had been set from a higher resolution 2km horizontal grid nested to a lower resolution 6km horizontal grid both with 30 vertical levels. The vertical levels in the simulation are set with a higher resolution within the first 100 meters. Heat flux, freshwater flux, temperature and salinity, sensible heat flux at the surface, and shortwave radiation are the surface forcing parameters acquired from the Comprehensive Ocean and Atmosphere Dataset (COADS). For wind forcing at the surface of the ocean, a daily climatology of the QuikSCAT satellite data was employed. Finally, the domain was

fed by the SODA or Simple Ocean Data Assimilation climatological state of January 1st and spun up for the year of 1980 using 5-day average surface fluxes and lateral oceanic boundary conditions.

As we are aiming to examine the differences of the ocean current feedback in the VTR region, we ran two simulation of ROMS only differing on the wind stress formulation. The first run of the model is characterized by the standard air bulk formula usually employed in most numeric models where there is no ocean current feedback in the air-exchange. That is, wind stress energy input is only originated by the wind. The equation of the standard air-bulk formula is depicted as follows:

$$\tau_{std} = \rho_{air} C_D |U_a|(U_a), \quad (2.1)$$

where U_a is the magnitude of the wind, as on Equation 1.3.

Moreover, the second model is characterized by the wind forcing with the bulk approximation described on Equation 1.3. In this sense, ocean current feedback is considered in the wind stress computation, enabling the analysis of EWI on the Ekman Pumping values.

For the sake of clarity on presenting the different models, the simulation forced by the bulk approximation of Equation 2.1 will be further named "uncoupled simulation" (or subscripted "A" variables, where A stands for atmosphere). The simulation forced by the bulk approximation of Equation 1.3 will be further named "coupled simulation" (or referred as subscripted "A-O" variables, to denote the atmosphere ocean coupling).

Chapter 3

Mesoscale Eddies of Brazil Current at the Vitória-Trindade Ridge

This first part of the work intends to provide a characterization of the mesoscale activity nearby the VTR region. Although a few studies had been investigated the VE; in terms of its translation equatorwards (Arruda et al., 2013; Campos, 2006) and its hydrographic and kinematic properties (Schmid et al., 1995); a statistical characterization has not been conducted for the main features of VTR, let alone the other mesoscale eddies associated to the BC such as the AE and VE.

Thus, prior to addressing of whether the eddy-induced Ekman pumping is an effective phenomenon in the VTR surroundings, we need to estimate the eddies' frequency and morphometric parameters. We intend to establish a climatological view of the main recurrent mesoscale features in the study area.

The use of SLA data acquired by altimetry satellites has become the main tool for analysis of mesoscale features. As the manipulation of altimetry data had improved, automated eddy detection methods have developed. These detection methods are essential for statistical characterization of geostrophic eddies.

In this section, we aim to provide a statistically significant characterization the two main mesoscale eddies of the VTR surroundings (Abrolhos and Vitória Eddies) using a 24-year altimetry SLA data. In particular, we aim to compute frequency/probability of formation and duration of both eddies monthly.

The chapter is organized in the following subsections: Section 3.1 describes the process of computation in the geostrophic currents in the region using altimetry. In Section 3.2 we explain the detection methods used to accomplish the statistical analyzes. Finally, Section 3.3 we determined the geometric, frequency and dynamic parameters of the two eddies.

3.1 Geostrophic velocity fields from altimetry data

As mentioned before, the altimetry data used is a merged set of different altimetry satellites outputs in order to provide a daily interpolated SSH and SLA data of the oceans. It is developed to perceive features of the order of 100km. Moreover, it fails to resolve smaller scales and has a low accuracy at coastal regions (Chelton et al., 2011). Nevertheless, it is the most reliable synoptic observation to measure geostrophic velocities of mesoscale features.

Since we are aiming to describe the geometrics and periodicity of the eddies in the region, we employed the SLA data series for the analysis, so we could retrieve information from the actual mesoscale features and velocity anomalies (Kurczyn et al., 2012). According to the geostrophic balance, η' (or SLA) horizontal gradients dictate the perturbations of mesoscale movement in the ocean. That is, from the balance between the Coriolis parameter and the pressure gradient, one may compute the geostrophic velocity anomalies as follows:

$$U'_g = -\frac{g}{f_o} \frac{\partial \eta'}{\partial y} \quad (3.1)$$

$$V'_g = \frac{g}{f_o} \frac{\partial \eta'}{\partial x}, \quad (3.2)$$

where g is the gravity acceleration, f is the Coriolis Parameter, and x and y are the eastward and northward distances, respectively. As we can observe, we consider the Coriolis Parameter is constant due to the fact the β -plane model is an irrelevant approximation for the range of latitudes of our study site. Based on those approximations, we are able to analyze the daily geostrophic velocity fields regarding the Brazil Current and the main mesoscale eddies.

To infer the BC path stability in the VTR region, we used the root mean square (RMS) of geostrophic velocities acquired by Equations 3.1 and 3.2 to evaluate the path stability of the over the period of 1993-2016. As we can observe from Figure 3.2, the BC path shows a greater stability south of TB. However, we can infer meandering and/or strong variability of the current in the VTR region, specially when the BC crosses VTR seamounts at 20-21°S. The path stability in the region corroborates the findings of Soutelino et al. (2011) stating that it due to an interaction between the topography of the region and shear of the BC. This interaction implicates in disturbances of the steady state of the current which may contribute to the formation of mesoscale features.

We can also analyze this eddy-dominated region by the mean and anomalous geostrophic flow of BC (Figure 3.1). By inspecting the region of high velocity anomalies, we can infer that TB region is dominated by eddy activity, mostly related to VE. The BC has a stable path near AB, which results in weaker magnitudes in the flow anomaly in Figure 3.1.

Figure 3.1: Mean geostrophic and averaged geostrophic anomalous velocities around VTR retrieved from GOGRO altimetry dataset from 1993-2016. The mask represents the limit of the shelf break (100m bathymetry) described in this region by Palóczy et al. (2016) .

Figure 3.2: The BC path variability pictured by the RMS of the geostrophic current values acquired by altimetry dataset from 1993-2016. The mask represents the limit of the shelf break (100m bathymetry) described in this region by Palóczy et al. (2016) .

3.2 Detection Methods

Since we are aiming to develop a statistically relevant analysis of the eddies, we need to use a statistically significant sampling data. Therefore, it is necessary to automatically detect the mesoscale eddies and measure their characteristics due to the long-term aspect of the dataset. We used two eddy detection methods on the SLA time series (1993-2016). The Okubo-Weiss (OW) (Okubo, 1970; Weiss, 1991) method was employed to detect the mesoscale activities and compute their frequency and duration. For the morphometric measurements such as: mean axis length, area, swirling speed, eddy kinetic energy (EKE), and amplitude; we employed the Winding-angle Method (WA) (Guo, 2004).

The OW method was selected to detect the recurrence of these features since it is an algorithm that relies its efficiency on physical criterion based on a balance among strain and relative vorticity components of the deformation tensor. Even though the method has been currently assessed by various authors (Arruda et al., 2013; Chaigneau et al., 2008; Chang and Oey, 2014), its limitations are well known (Chelton et al., 2011). They are: (i) the method uses the second derivatives of SLA, therefore it enlarges the error in the analyzed field; (ii) the threshold of the area of the mesoscale features is often misplaced or erroneously measured; finally, (iii) the method tends to overestimate the number of mesoscale features, though erroneous detections are easily reorganized as not timely and spatially coherent.

Nevertheless, the OW is one of the only algorithms that uses dynamic parameters to analyze mesoscale features in satellite products. Therefore, the method was chosen to evaluate only the frequency/duration parameters of the eddies.

The WA method was chosen to analyze the geometric features of the eddies since it uses a geometric approach. Consequently, the retrieval of the dynamic parameters was more reliable than the OW method due to the issues regarding the threshold of the eddy area formerly mentioned. On the other hand, the WA efficiency performance was not as successful for the temporal parameters due to data interpolation computed by the altimetry dataset. That is, a deformation of the eddy due to interpolation may interfere in duration parameter acquiring efficiency. This issue has been previously reported by Chaigneau et al. (2008).

The selection of specific algorithms for different section of the mesoscale features was an adaptation of the work of Chaigneau et al. (2008) from which used both algorithms and computed the statistical efficiency of both performances. In the following section, we described the two methods used for the analysis.

3.2.1 Okubo-Weiss Method

The OW method is a detection algorithm that computes the dominance between strain and relative vorticity in the ocean as it balances those components in opposite signs. The

method detects mesoscale activity when there are patches of predominant relative vorticity in the field. In other words, mesoscale features are highlighted when the vertical curl of the geostrophic currents are higher in magnitude than their deformation rate. The algorithm is described as:

$$OW = S_s^2 + S_n^2 - \omega \quad (3.3)$$

where the S_s is the shear component of strain (or shearing deformation rate), S_n is the normal component of strain (or stretching deformation rate), ω is the relative vorticity. These three terms of the equation are derivatives of the geostrophic velocities which are determined as:

$$S_s = \frac{\partial V'_g}{\partial x} + \frac{\partial U'_g}{\partial y} \quad (3.4)$$

$$S_n = \frac{\partial U'_g}{\partial x} - \frac{\partial V'_g}{\partial y} \quad (3.5)$$

$$\omega = \frac{\partial V'_g}{\partial x} - \frac{\partial U'_g}{\partial y} \quad (3.6)$$

We can infer that negatives values of OW refer to high relative vorticity which highlight mesoscale eddies that are actively rotating at those areas; as we can observe from the equation 3.3. A common threshold delimited for an effective detection of mesoscale eddies is $OW_o = -0.2\sigma_W$, where σ_W corresponds to the standard deviation of the OW at each computed field.

In Figure 3.3 we can observe the detection of anticyclonic Abrolhos Eddy and the cyclonic Vitória Eddy in a event with both features occurring (Arruda et al., 2013). The two relative minimums of the Okubo-Weiss parameter depict the occurrence both eddies in the field. The σ_W contour lines delimit their areas. As aforementioned, the contours do not precisely measure the morphometric parameters of such features. Using the geostrophic velocities to support it, the delimitation of the eddy is not efficiently covering the area of intense rotation of the two eddies.

3.2.2 Winding-angle Method

As mentioned before, in a geostrophically balanced ocean, the northward and eastward gradients of SLA dynamically represent the geostrophic oceanic velocities. In other words, η' lines (SLA) on the contour field are geostrophic current streamlines. Therefore, we define an eddy, in a more strict sense, as a closed SLA streamline.

The WA method aims to detect eddies by locating and clustering closed streamlines in the fields of SLA. It firstly locates the center of the eddy as the innermost closed

Figure 3.3: Okubo-Weiss field of an three-month averaged with the two mesoscale features in the VTR surroundings. The dipole Abrolhos-Vitória is enhanced in the relative minima of OW delimited by the gray contour lines. This is an averaged field over the period from of August to October of 1997 corresponding to the duration of the eddies. The arrows represent the geostrophic velocities acquired by the altimetry dataset.

streamline of a moving window. The edge of the eddy is detected as the outermost closed streamline of the determined feature.

In order to understand how the method performs, let us consider the horizontal streamline beginning at point P1 and followed by the adjacent segments with spatial difference corresponding to the step of the SLA field ($\frac{1}{4}^\circ$). The WA of the streamline corresponds to the cumulated sum of the angles between each angle of consecutive pair of the segments (Figure 3.4).

It is possible to observe the schematic representation of the method. The algorithm computes the angle (α_j) winding the backward and forward segments of the point in the streamline. If the α_j is equal or higher than 2π , the method classifies the streamline as closed.

In order to obtain a more precise and consistent output for the geometric parameters, we establish some thresholds regarding the detection:

- After the WA application, we computed the mean swirling velocity (U_{θ_m}) and mean radii (R_m) of both eddies in order to estimate the mean and standard deviation of the turnover time (T_{eddy}) which is denoted as:

$$T_{eddy} = \frac{R_m}{U_{\theta_m}}. \quad (3.7)$$

- The turnover time of the eddy was used to discard the eddies which did not meet the criteria (duration lower than $T_{eddy} + 3\sigma_{T_{eddy}}$).

Figure 3.4: Representation of Winding-angle detection method (WA) for a segmented streamline. Each α_j angle represents the signed angle between the forward and backward segments of a point at the streamline. According to Chaigneau et al. (2008).

- The method was employed to detect the formation of the eddies at the areas previously known as formation sites in the literature: east of AB for AE and south of the VTR and east of the TB for VE (Schmid et al., 1995; Soutelino et al., 2013).
- An event characterization was obtained by averaging the eddies' parameters and properties over the period at which the turnover time passed the threshold criterion of 3 standard deviations.

3.3 Eddy Characteristics

Although Vitória Eddy translation events were previously reported by the works of Campos (2006) and Arruda et al. (2013); we chose not to cover the tracking of this eddy phenomena in order to simplify our analysis. From the results of the detection algorithms, we retrieved the following characteristics with the WA method: amplitude, mean apparent radius, swirling speed, Rossby number, eddy amplitude, and eddy kinetic energy (EKE). In addition, we retrieved duration, monthly probability of formation (P_{form}), and monthly probability of duration (P_{dur}) with the OW.

Regarding the frequency parameters acquired, the probability of formation was denominated as a ratio of the number of events that were formed at a specific month over the total number of events. This expression gives us a sense of an annual distribution of the mesoscale features' formation. It can be expressed as follows:

$$P_{form} = \frac{n_m}{\sum_{i=1}^{12} n_i}, \quad (3.8)$$

where the n is the number of events formed in a month.

The monthly probability of duration is an analogous expression to 3.8. However, it

takes into account the months of which the eddy was formed and perdured as well. In other words, this probability computes the ratio between the events that formed or persisted in a specific month over the sum of the samples of recurrences. It is important to highlight that for one eddy event that persisted more than one month, one sample for each month of eddy persistence was taken into account. The expression is described below:

$$P_{dur} = \frac{p_m}{\sum_{i=1}^{12} (p_i)}, \quad (3.9)$$

where p is the number of eddy events that occurred (were formed or persisted) in a month.

For the dynamic characteristics, in order to present a robust sample of the eddy events; the parameters were computed as a mean of the specific duration of each specific eddy event. The mean for each parameter was formulated as:

$$S_{mean} = \frac{\sum_{i=1}^n S_i}{n} \quad (3.10)$$

where the S_{mean} is the mean parameter value, S_i is the parameter value at a specific eddy event day and n the number of days of the eddy event. The eddy characteristics were formulated and computed according to the work of Kurczyn et al. (2012).

The mean apparent radius (R_{app}) of the eddies was estimated by the area retrieved from the geostrophic streamlines in the SLA field. As previously mentioned, the last close geostrophic contour was retrieved by the WA detection method as a vortex edge. Once the streamline was retrieved, the area of its respective polygon was fit to a idealized ellipsis and then the apparent radius was computed. The mean diameter formula is described as

$$R_{app} = 2\sqrt{\frac{Area}{\pi}}. \quad (3.11)$$

As for swirling velocity, Rossby number, and duration; the maximum parameter was chosen to represent the eddy event. The swirling velocity corresponds to the geostrophic speed inside the eddy. The duration is the number of days between the eddy formation and its dissipation. The Rossby number is the dimensionless ratio of horizontal velocity (swirling speed) over the length of the feature (apparent radius) multiplied by the Coriolis parameter. The expression is described as follows:

$$Ro = \frac{U_{\theta}}{R_{app}|fo|} \quad (3.12)$$

Regarding EKE ($\text{m}^2 \text{s}^{-2}$) values, it was represented as a spatial average of the parameter within the eddy boundaries. The EKE is formulated as

$$EKE = 0.5\sqrt{(U_g'^2 + V_g'^2)}. \quad (3.13)$$

where U'_g and V'_g are the zonal and meridional velocity anomalies of the geostrophic balance. These speed parameters are computed as the total speed fields subtracted from their respective temporal means of the entire available SLA data (1993-2016). Thus, EKE becomes a proxy of eddy intensity in the field.

Lastly, the amplitude of an eddy (A_{eddy}) is the absolute difference between the sea level anomaly of the eddy center and its edge. The amplitude is described as

$$A_{eddy} = |\eta'_{center} - \eta'_{edge}|, \quad (3.14)$$

where η'_{center} and η'_{edge} represent the sea level anomaly of the eddy center and eddy edge, respectively. Due to the choice to not deepen our study into eddy propagation, we did not compute the non-linearity parameter as Kurczyn et al. (2012), since propagation speed is a parameter into the equation.

3.4 Eddy Analysis

After taking into consideration the eddy duration higher than $T_{eddy} + 3\sigma_{T_{eddy}}$ as a cut-off and averaging the parameters of the eddies; 56 Vitória Eddy events and 28 Abrolhos Eddy events were detected in 24 years of SLA data. Therefore, AE is less frequent than VE in the Vitória-Trindade Ridge surroundings, at least by taking in consideration a strict duration cut-off for mesoscale eddies mentioned above. The mean annual eddy occurrence for the AE and VE are 1.2 and 2.3 respectively. The mean parameters of both eddies are displayed at Table 3.1.

Dynamical parameters such as swirling speed, apparent radius, and EKE present linear proportionality to the amplitude of the eddy (Figure 3.5). We can observe strong correlation between the amplitude parameters and the swirling speed and EKE of the eddies. However, moderate to low correlation coefficients were found between amplitude and apparent radius, specially for the VE. This is may be due to the meandering of BC and its interference with the eddy events regarding the stream function field. The correlations are consistent with the correlations presented by Kurczyn et al. (2012).

As mentioned before, we computed two different types of annual probabilities for AE and VE. The P_{dur} and P_{form} were computed and plotted in bar graphs in Figure 3.6. The maximum probability of formation of AE was at February (17.8%) whereas Vitória Eddy forms mostly in May (14.5%). The lowest frequencies for AE (VE) are June and July (August and December) with both 3.5 % (3.5%) of probability.

We can roughly observe a seasonal patterns for both eddies formation if we observe the P_{dur} graph. VE events have two periods of low probability (mid-winter and late spring months) and May as the high probability period (fall). AE has its probability peaks mostly in summer (February and March) and spring (October and November) and has

Table 3.1: Eddy geometric and dynamic parameters for Abrolhos and Vitória eddies.

Eddy	R_{app}	U_θ	R_o	A_{eddy}	EKE	$Duration_{mean}$	$T_{eddy}(\sigma_{T_{eddy}})$	$Duration_{max}$
Vitória	49.1 ± 6.1 km	15 ± 8 cm s ⁻¹	0.06	1.8 ± 1.0 cm	1.6×10^{-2} m ² s ⁻²	32 days	4 (3) days	129 days
Abrolhos	49.2 ± 6.1 km	11 ± 3 cm s ⁻¹	0.75	1.1 ± 0.5 cm	0.7×10^{-2} m ² s ⁻²	32 days	6 (4) days	70 days

Figure 3.5: Linear correlation of three eddy properties versus amplitude for all eddy abundance (AEs in red; VEs in blue): (left) amplitude versus swirling speed, (center) amplitude versus Apparent Radius, (right) amplitude versus EKE. Solid lines represent linear fits. The Pearson Correlation Coefficient and the asymptotic significance (or probability value) are denoted in the upper part of the plots as "pearson r " and "p value", respectively.

Figure 3.6: Probability of formation (upper panel) and of duration (lower panel) for the Abrolhos Eddy (red bars) and Vitória Eddy (blue bars).

its lowest frequencies in January and May (summer and fall months, respectively).

By observing the probability of duration bar graph (Figure 3.6), we can examine the seasonality pattern more clearly. The highest probability of a AE event to happen is in winter months whereas they are the least likely to have a VE event. Highest duration frequencies of VE are also in April and October as formation frequencies. To understand the relationship between the eddy frequencies and the BC, we computed a monthly climatology of BC's transport in the region using a global model that not only has the same time-period employed in the altimetry data but also assimilates the CMES altimetry data set in the reanalysis.

In order to confirm the model in a simplistic manner since we are dealing with a simulation that assimilates the observation analyzed in this section, we generated, for the whole time series of the model/observation, an averaged field of velocities for comparisons. The confirmation is represented in Figure 3.7. The magnitude values are similar for both fields as well as the location of maximum speed of the BC. The paths of the current in the global model are slightly different from the observation where the jet tends to flow attached to the bathymetry even at TB. Nevertheless, there is still a significant path of the BC going through the seamounts as presented in the observations.

Figure 3.7: Climatology fields for altimetry observations (left) and the MERCATOR model (right) averaged from 0 to 50m.

The confirmation allowed us to generate a monthly climatology of transport in two latitudes of the region where the influences of VE and AE would be minimized. Hence, the analysis of transport would be far from the effect of enhancement/depletion of transport may be caused by the eddies. The latitudes chosen to compute the sections were selected from a region of higher and lower latitude of both eddies (17°S or North Section and 22,4°S or South Section). The climatological transport plot is shown on 3.8.

Furthermore, we computed the Pearson correlation coefficient for the probabilities of formation and duration versus the monthly transports to infer the similarity between the four climatology parameters. The values of correlation are displayed in the table 3.2. The formation of the eddies does not seem to be related to the transport of the current, if we consider that the correlation coefficients suffice to elucidate a cause-effect process.

Figure 3.8: Brazil Current's monthly climatology transport from the MERCATOR global model in two areas of the VTR ridge.

Table 3.2: Table of correlation coefficients between the probabilities and the current transports for both eddies.

AE	North Section	South Section
P_{form}	0.54	0.49
P_{dur}	0.70	0.75

VE	North Section	South Section
P_{form}	-0.40	-0.54
P_{dur}	-0.63	-0.51

Values of any correlation with the P_{form} are not greater than 55% of correlation. If we are to consider the BC transport as the main factor influencing the eddies generation, the probability of duration of the AE is robustly related to the transport of the BC (75% and 70% for the South and North region respectively). On the other hand, the VE events are as strongly correlated with the BC transport. There is a moderated correlation value of 63% between the probability of duration of the VE with the transport of the BC when measured in the North Section.

In respect of the eddies' morphometry and their dynamical parameters, we compared the computed results with the literature. The first description of VE by Schmid et al. (1995) is based on a 55-day VE event in 1991. According to the study, the VE event had a mean diameter of 100 km and swirling velocities of 0.4 m s^{-1} at the western flank and 0.5 m s^{-1} at the eastern flank that was attributed to the direct influence of the BC. In addition, the eddy event had a mean Rossby number of 0.15. Since the observational data was collected 2 years before the first year of altimetry data, we were not able to

establish a confirmation of the detection methods with the observed variables as well as a bias computation.

The values obtained in this work are based on objective criteria and agree well with the previous study of Schmid et al. (1995). The discrepancies in terms of velocity data are probably related to the comparison between a single event study which employed in-situ data and a climatological approach used here.

3.5 Chapter Highlights

From the usage of 24-year altimetry data series, we developed a thorough analysis of the two most recurrent eddies in the Vitória-Trindade region for dynamic, frequency and morphometric characteristics. Despite the polarization, the eddies presented similar characteristics in respect of dynamic and morphometric parameters, which are consistent with the previous descriptions of these features in literature (Schmid et al., 1995; Campos, 2006; Soutelino et al., 2013).

Both AE and VE are persistent through the year, though VE annual mean occurrence and total events in the analysis are twice greater than the AE's values. Although the VE is more prone to occur, probability of duration and formation rates are centered in one single period of the year: May for formation and May, June and July for duration. On the other hand, the events present two peaks of higher probability of formation and duration in the beginning and end of the year (mostly February and November).

The BC volume transport is correlated with the Abrolhos Eddy persistence at the level of confidence of 95%, whereas Vitória Eddy does not seem to be correlated with the transport variation neither in formation or duration. Finally, this characterization is essential for our further analysis where we parametrize both eddies and employed a artificial wind field for horizontal distribution of Ekman pumping in a eddy domain.

Chapter 4

Eddy-induced Ekman Pumping Analysis for an Idealized Gaussian eddy

In this chapter, we developed a parametric analysis for idealized Abrolhos Eddy and Vitória Eddy structures using the parameters obtained in the statistical characterization (3. This allowed us to have a sense of how the ocean current feedback influences the wind stress energy input in the studied features in a climatological sense. Using this type of approach, we can improve the understanding on how linear and non-linear Ekman pumping horizontal distribution is set for a simple idealized model. This chapter is based on the work of Martin and Richards (2001) which also computed the Ekman pumping of a potential modal-water eddy horizontally, as well as the work of McGillicuddy et al. (2007), which introduced the differences in ocean current feedback in idealized parametrizations. This chapter is organized as follows. In the section 4.1, we described the wind stress field. In section 4.2, we present the equations for the idealized eddy and the dynamical parameters used. In section 4.3, we discuss our output results and how they can be interpreted. Finally, in section 4.4, we draw our preliminary conclusions of the chapter.

4.1 Wind Stress Field Setting

As we aim to infer Ekman pumping fields as a simplified model computed from an idealized eddy, we opt here to use a constant northeastern wind field. Moreover, as our focus is to depict the ocean current feedback, we employed the air bulk formula for wind stress, as presented by Equation 1.3.

It is straightforward to understand how ocean current feedback is relevant in the input of curl of wind stress if we analyze the differences of wind stress magnitude obtained by the bulk formulas presented in this study (See Equations 2.1 and 1.3). In Figure 4.1, the left panel exhibits a constant field of northeastern wind stress computed from the standard air bulk formula (Equation 2.1). The resultant flow is irrotational. On the other

hand, if we are to compute the wind stress from Equation 1.3 over a mesoscale eddy, the presence of currents in opposite direction on the north-south (or east-west) of the edge of the vortex would result in a rotational wind stress field (right panel). Hence, the wind stress distribution may favor Ekman Pumping despite a homogeneous wind field blowing over the ocean surface.

Thus, we chose a wind field constant in direction and magnitude to compute the Ekman pumping fields for the idealized eddy calculations presented next. The magnitude of the wind components for both zonal and meridional were 6 m s^{-1} which yielded a mean total magnitude of 8.5 m s^{-1} . Assuming a constant air density ρ_{air} of 1.3 kg m^{-3} and mesoscale geostrophic speeds of $O 10^{-1} \text{ m s}^{-1}$, based on the values of Table 3.1, we can estimate that the wind stress magnitudes are around 0.10 Pa with a range of fluctuations of $O 10^{-3} \text{ Pa}$ as depicted in Figure 4.1 .

The setting of magnitude and direction of the wind field was based on a simplification of the climatological windrose from Figure 1.8. One may infer an reasonable value for the field as a southwesternward wind regime with 8.5 m s^{-1} of magnitude. It is important to highlight the smoothing that may occur for the quantitative analysis since we are dealing with a simplified static model with idealized eddies and wind fields.

Figure 4.1: Artificial wind stress fields settings for the parametrized computations. On the left, we can observe a constant wind stress field from the standard bulk formula with no variation of magnitude or direction. On the right, it is straightforward to observe the ocean current feedback clearly changing the magnitude of the wind stress field, which allows it to have a scalar curl.

4.2 Swirling Velocity Configuration

For the sake of simplicity, we assumed a Gaussian eddy based upon the values acquired by the Statistical Characterization (See Table 3.1). The velocity profile is based on the work of Martin and Richards (2001) where they adjusted an idealized eddy. The same equation for swirling velocity and stream function are, respectively:

$$U_{\theta}(r) = U_{\theta_{max}} \left(\frac{r}{R} \right) \exp\left[\frac{1}{2} \left(1 - \frac{r^2}{R^2} \right)\right] \text{ and} \quad (4.1)$$

$$\psi = U_{\theta_{max}} R \exp\left(\frac{1}{2}\right) \left[1 - \exp\left(-\frac{r^2}{2R^2}\right) \right], \quad (4.2)$$

where the $U_{\theta_{max}}$ is the maximum speed of the eddy, R is the maximum radius of the eddy, r is the euclidean distance of the feature. The grid used for the computations were developed as coordinate based according with the mean region of formation. The parameters of both eddies are maximum velocities and radii of the eddies. For the AE, the mean maximum velocity is 0.11 ms^{-1} and the mean radius is 49 km. The VE has a 0.15 ms^{-1} with also 49 km of radius. The equations described above set the radius at the edge of decay from the maximum velocities to zero in the Gaussian streamlines.

In order to detail the patterns of the eddy velocity; we plotted the stream function and velocity of the AE in Figure 4.2. We set the maximum velocities to be in the same location of the eddy radius.

4.3 Parametrization Results

Using the AE and VE averaged values and Equations 4.1, 4.2, we developed the parametrization models for each eddy. We obtained eddy azimuthal velocity, total Ekman pumping W_{Total} , linear Ekman pumping W_{τ} , and non-linear Ekman pumping W_{ζ} by the idealized eddy structure and a constant wind field.

In Figures 4.3 and 4.4, the parametrized model is depicted. Regarding W_{τ} , AE presented a positive velocity of 0.14 m day^{-1} at the center of the eddy and small to virtually null values outside the edge of maximum velocity. For the VE, W_{τ} field presented the opposite pattern. At the center of the eddy, two lobes of opposites velocities are also yielded. However, the higher-magnitude lobe presents velocities of -0.13 m day^{-1} . Those horizontal patterns corroborate the theory of the EWI or linear eddy-induced Ekman pumping where there is stronger vertical velocities patches in the center of an anticyclonic eddy near the ekman layer depth (d_{Ek}). The symmetrical opposite may be described for a cyclonic eddy with negative linear Ekman pumping values in the center of the eddy (Dewar and Flierl, 1987; McGillicuddy Jr and Robinson, 1997; Flierl and McGillicuddy, 2002).

As for the W_{ζ} fields, both computations presented a dipole-like structure with stronger

Figure 4.2: Transectional patterns of stream function and swirling velocity of the Abrolhos Eddy experiment. We may observe the maximum speeds at the edge of the eddy represented at 49 km.

values on the edge of the eddy and oriented by the wind mean direction. Such pattern was previously described by Mahadevan et al. (2008) where the shear of relative velocity multiplied by the components of wind stress would generate dipole feature along the eddy. The values of inferred vertical velocity in those regions were higher than the ones computed in the linear Ekman pumping fields. As we can observe from Figures 4.3 and 4.4, in the edges of AE and VE located on a axis perpendicular to the wind-field mean direction (Figure 4.1), the vertical velocities are greater than $\pm 0.5 \text{ m day}^{-1}$. The position of the positive and negative vertical velocity lobes are dictated by the polarization of the eddy.

In a strictly steady and horizontal-limited analysis, we may infer that non-linear Ekman pumping plays a more important role in the vertical velocities patterns than the linear Ekman pumping. In the subplot of W_{Total} of Figures 4.3 and 4.4 it is possible to observe that the distribution of W_{ζ} is dominant (dipole structure) for both eddies. The magnitudes of inferred vertical velocities for the non-linear Ekman pumping are one order of magnitude greater than the ones from W_{τ} .

However, as mentioned in the introductory section, it has been debated that regarding the effectiveness of both terms of the total Ekman pumping equation. The work of McGillicuddy et al. (2007) has emphasized the dominance of linear Ekman pumping value in the center of the eddy even though it is usually smaller than non-linear Ekman pumping in magnitude. The work states that the azimuthal average of both terms depicts

Figure 4.3: Eddy-induced Ekman pumping for parameterized AE using the dynamic parameters acquired in the statistical characterization. The cyan limit represents the maximum velocity contour.

Figure 4.4: Eddy-induced Ekman pumping for parameterized VE using the dynamic parameters acquired in the statistical characterization. The cyan limit represents the maximum velocity contour.

a more honest estimate of how the resultant vertical velocities behave. It is based on the hypothesis that eddies are rotating features, and an appropriated sampling method that assess the actual effectiveness of the mean fluctuations of the Ekman pumping along the eddy radius is the azimuthal mean.

Since W_ζ fields are usually a dipole structure with a significant symmetry, an azimuthal mean would cancel the two poles of opposite vertical velocities. That effect flattens the vertical velocity variations to nearly null rates along the radius of the eddy. According to this depiction (Figure 4.5), the magnitude of the vertical velocities in the dipole in an idealized Gaussian eddy and constant wind field would not matter due to canceling effect of the rotation of the eddy. On the other hand, in respect of the linear Ekman pumping values, there are strong positive (negative) vertical velocity in the center of the AE (VE) parametrizations. This would not be affected by the azimuthal averaging due to the fact that these values are summed with smaller opposite velocities in the eddy center, resulting in consistent velocities as depicted in Figure 4.5.

In terms of azimuthal averaging, the values of non-linear Ekman pumping are close to null in both panels of Figure 4.5. However, we can observe significant values for linear Ekman pumping in both eddies parametrizations. The analysis of AE yielded values of 0.05 m day^{-1} and for the VE we can observe values of approximately -0.04 m day^{-1} . Those values dominate the patterns of W_{Total} in both models.

Taking into consideration, as statistically-significant parameters, the mean duration of the eddies; we can compute the mean vertical velocities in the d_{Ek} for both eddies. Since both eddies had a mean duration of 32 days (see Table 3.1), the vertical displacements in the center of the eddy for AE and VE experiments would be $\cong 4.5 \text{ m}$ and $\cong -4 \text{ m}$, respectively. It is important to highlight that the values of vertical velocities inferred by Ekman pumping are smaller in comparison with the works of Martin and Richards (2001); McGillicuddy et al. (2007); McGillicuddy Jr (2015) where the order of magnitude found were the one order of magnitude higher than in the present study. However, this low values represent only a climatological based analysis. In other words, the values of both eddy swirling velocity and wind speed may be higher in magnitude which would dramatically enhance the effect of Ekman pumping decay first-baroclinic-mode eddies into ITEs. If we take into consideration a 0.45 m s^{-1} and swirling velocity and 50 km radius for the same wind field as described by Schmid et al. (1995); we should find Ekman pumping values approximately three times greater than the ones presented, let alone the inferences in wind speed and direction. In Figure 4.6, we do not only observe the higher absolute values of linear Ekman pumping in the eddy center but also and important influence of non-linear Ekman pumping in the eddy's edge. The mean values of -0.42 m day^{-1} in the eddy center for a 55-day duration eddy, as described in the work of Schmid et al. (1995), may yield isopycnal displacements of $\cong -23\text{m}$.

Figure 4.5: Azimuthal means of Ekman pumping terms in meter per day. The origin of the x axis represent the center of the parametric eddy. The upper plot depicts the Abrolhos Eddy Fluctuation of both terms of Ekman Pumping whereas the Vitória Eddy plot is represented at the bottom plot.

Figure 4.6: Azimuthal mean of Ekman pumping components with the morphometric parameters based on the observations of Schmid et al. (1995). The origin of the x axis represents the center of the parametric eddy.

4.4 Chapter Highlights

The employment of a parametric model to compute the inference of vertical velocities driven by Ekman balance in two idealized eddy patterns emerged from the works of Martin and Richards (2001); McGillicuddy et al. (2007). We developed two distinct eddy experiments for the parametrization using statistically robust values retrieved using from the altimetry characterization of the Abrolhos and Vitória eddies as described in Chapter 3. The equation for the idealized Gaussian eddy employed is due to the work of Martin and Richards (2001). From this equation we could set a stream function field that restricted the maximum velocities magnitudes at the edge of the eddy radius for both experiments.

Vertical velocities resembled the patterns described in the literature: the anticyclonic eddy experiment (AE), where quasi-geostrophy states a strong negative vertical velocity value, yielded positive vertical velocities at the base of the Ekman Layer due to its relative vorticity and the wind field regime. Analogous results were obtained in the cyclonic eddy (VE). Positive vertical velocities driven by quasi-geostrophy are expected, yet Ekman Pumping generates negative vertical velocities. We evaluate that total Ekman pumping

velocities are consistent yet lower with the order of magnitudes described in the works of Martin and Richards (2001). In addition, for both experiments, the terms of W_τ showed greater values in magnitude than the ones yielded by W_ζ when averaged in an azimuthal manner as stated McGillicuddy et al. (2007). Nevertheless, W_ζ presented absolute values of vertical velocity order of 10 higher than W_τ , as reported Mahadevan et al. (2008).

Some aspects that are not taken in consideration regarding the parametrization are: (i) asymmetry of the eddy structure, (ii) daily-varying wind fields, and (iii) time stepping of the model. The former item of the list presents an important issue that is not considered in the idealization of the eddy. Since we are dealing with Gaussian idealizations, the effects of shear of the velocities in the eddy are well-behaved and smoothed. What is more, we are not taking in consideration the background flow (BC) that may influence in the eddy symmetry as well. With respect to non-linear Ekman pumping, it takes into account the horizontal gradients of relative velocities multiplied by the wind stress components. Hence, asymmetry in the eddy velocity structure caused by eddy decay, wind forcing, bathymetry interaction that may infer in greater pulses of velocity in the W_ζ that are not depicted in the analysis.

Moreover, the wind stress fields are constant in magnitude and static in direction for simplicity of the parametrization, as mentioned previously in the first section of this chapter. Such idealization filters any potential curl values of wind stress that may add or decrease in the total Ekman pumping field, specially in the linear Ekman pumping values. Therefore, the input of the eddy-induced Ekman Pumping that would be yielded by solely wind shear are filtered out of this analysis.

Those two aspects referred above (i and ii) are intrinsic to time dependency (iii); time stepping may play a considerable role in a analytical analysis that computes non-linear Ekman pumping due to that fact that we can consider a variable wind stress field and a decaying velocity structure of the ocean. As we compute our model from fields that do not vary in time, we cannot expect to see the cumulative values inferred by the Ekman pumping equation. In respect of W_ζ , those differences may be much more important due to variations of relative vorticity that may be caused by differences of eddy velocities from time. Hence, an improved analysis would take into consideration ageostrophic circulation, varying winds and time-dependent model to provide a more realistic output.

For further evaluation of this phenomenon, one should investigate the outputs of ocean current feedback as eddy-induced Ekman pumping in a three dimensional model coupling the vertical and horizontal density fields of the eddy domain as well. This would be a variation of the work of McGillicuddy Jr (2015). With the vertical step in the parametrization, one can present the differences in velocity, isopycnal displacements and even primary production.

For this specific region, a more representative analysis might be a altimetry-averaged interpolation of the eddies presented, similar to the global analysis of Gaube et al. (2015).

As we are using observations, this will present a quantitative result of eddy-induced Ekman pumping in the Vitória-Trindade region.

Finally, Ekman pumping in these experiments yielded smaller vertical velocities, yet we can observe consistent results corroborating a possible eventuality of intrathermocline lens eddies in the Vitória-Trindade Ridge region. This analysis allowed us to understand in a simplified manner how the terms of Ekman Pumping behave distinctively and how isopycnal displacements and vertical velocity patterns might be altered by those terms. Using numerical simulation, we may add fluctuations to the ocean current and wind field as well as non-linearity terms not related to Ekman theory. With this in mind, we can grasp and analyze eddies in complex settings.

Chapter 5

Analysis of the ROMS Simulations

In Chapter 3, the characterization of the two mesoscale features recurrent in the VTR region was employed regarding frequency and morphometry. With this in mind, in Chapter 4, we built a parametric model that allowed us to understand the main horizontal distribution of Ekman pumping and its inferences in vertical velocity within the Ekman depth layer, we also examined the effectiveness of the linear component (Eddy-Wind Interaction) over the non-linear of Ekman pumping inside the eddy domain (McGillicuddy et al., 2007, 2008). Those analyses enlightened us to investigate the phenomenology of ITEs and more realistic inputs on eddy-induced Ekman pumping using a numerical model. As vertical structure is also available for our analysis, we will be able to analyze not only the density but also the vertical velocity structure and evolution of AE and VE in the VTR region.

Thus, in this chapter, we analyze two simulations using ROMS for the same area and time frame and employing similar boundary and initial conditions, which differ only by the wind stress forcing applied to the surface of the ocean. As mentioned in Chapters 3 and 4, we further explore the wind stress computed by the air bulk formula given in Equation 1.3. This air bulk formula represents the wind stress resulted from atmospheric winds and ocean current feedback inherent to this air-sea interaction. In this work, this simulation will be coined as "coupled simulation" or subscribed as "*A-O*" due to the atmosphere and ocean deduction. The second simulation was carried out using the usual air bulk formula presented in Equation 2.1, that only takes into account the wind velocities in the wind stress computation. We coined this simulation "uncoupled simulation" or subscription of "*A*".

Since the first theoretical studies of surface currents of eddies changing the sign of resulting wind stress curl Dewar and Flierl (1987), several works investigated the role of eddy-induced linear Ekman pumping in regional (Martin and Richards, 2001; McGillicuddy et al., 2007; Li et al., 2014) and global scale (Anderson et al., 2011; Gaube et al., 2015). Correspondingly, the importance of eddy vorticity on Ekman pumping has been explored by Stern (1965). Moreover, a number of works treating non-linear Ekman

pumping has also been reported in literature (Mahadevan and Tandon, 2006; Mahadevan et al., 2008).

Besides the surface-induced wind stress curl, another eddy-induced Ekman pumping term has been reported by Chelton et al. (2004). This component with regard to SST may cause damping/boosting in the curl values due to crosswind gradients of SST on eddies. However, we will not include this component as we are aiming to restrict our attention only to the problem of a surface-induced Ekman pumping analysis.

The employment of the aforementioned settings of simulations aimed to analyze the potential differences of ocean current feedback by the mesoscale eddies. As we are dealing with a complex numerical model in the same area and conditions, we use the results of this analysis as an inference of the effective recurrence of ITEs as well as the impact of eddy-induced Ekman pumping in the VTR region.

This chapter is organized as follows. Section 1 presents the confirmation of the simulations. In section 2, we present the vertical analysis of the eddy density and stratification structures. The evaluation of the vertical velocity in the eddy domains are depicted in section 3. Section 4 summarizes the conclusions and future work of this thesis.

5.1 Model Output Confirmation

Prior to the simulation analysis, the confirmation of the numeric simulation for a specific phenomenon is vital. According to (Oreskes et al., 1994), validation and verification of numerical models in natural system is a task virtually impossible to accomplish due to the fact that simulations need to be closed and their outputs are intrinsically not singular. That restrict us to "confirm" the model rather than validate or verify it. That is, there is a relative evaluation to the phenomenon of interest, which is denominated confirmation. Thus, we aimed to confirm the BC transport value throughout the year and the surface EKE/MKE horizontal distribution for both simulations in the VTR region.

It is important to highlight that since our forcing sets are primarily climatological (See Section 2.3), we cannot consider our simulations dated but as a representative sample of the VTR dynamics. Therefore, we compared the values of BC volume transport and EKE/MKE with those in literature obtained from climatology. Our model spin-up reached energy stability after 4 years. After this period, where the model simulations are energetically stable, we employed our analyses at the 5th year of the simulation outputs.

Firstly, we analyzed whether the BC transport is consistent with the literature values. We chose an artificial transect to depict the mean path and transport values of the BC, as showed in Figure 5.1. We observe the mean sectional structure of the BC and the variation of volume transport along the year from the coupled simulation from ROMS. We can observe that the BC volume transport is consistent with the values described in literature (Evans et al., 1983; Stramma et al., 1990). In addition, the similarity of

fluctuations within the year with the MERCATOR reanalysis presented in Figure 3.8 is also significant. Nevertheless, our dynamical analysis would be restricted to the second month of the simulation.

Figure 5.1: Evaluation of the Brazil Current transport, velocities and pathways for the coupled simulation output. The upper panel presents the transect chosen for the feature analysis. The middle panel shows the mean velocity section of the Brazil Current System at the upper ocean. The lower panel presents the BC's transport time series and the comparison with previous studies. Courtesy of Paulo H. Calil (Helmholtz-Zentrum Geesthacht Zentrum für Material- und Küstenforschung)

To analyze the formation and decaying of the eddies with no initial conditions influence, we established a threshold of one month after the first day of the simulation. Moreover, since AE and VE presented mean duration of approximately 30 days (see Table 3.1), we considered a month-long subset of the simulations would suffice for our analysis. In addition, each wind stress condition drove the surface current dynamics distinctively, leading us to select a recent month so the simulation would not become significant different, and, hence, we would still be able to consider the eddies analogous in each simulation. As a result, we selected February for all of the analysis conducted in this chapter.

Furthermore, some studies have already addressed the differences in energy balance and wind work input using the two wind stress forcing employed in this present work.

According to Eden and Dietze (2009), Kang et al. (2016), and Renault et al. (2016); if we take into account the feedback caused by ocean current velocities; the wind stress input works as an "eddy-killing" process. That is, a stabilization of surface current is observed when comparing the air-sea interaction forcing of coupled over an uncoupled simulation. Eddy kinetic energy (EKE) decreases whereas the mean kinetic energy (MKE) has stronger paths in comparison with a balance where ocean current feedback is neglected (Eden and Dietze, 2009). The equations describing MKE and mEKE are, respectively, presented below:

$$MKE = \frac{(\bar{U}^2 + \bar{V}^2)}{2}, \quad (5.1)$$

$$mEKE = \langle \frac{(U'^2 + V'^2)}{2} \rangle, \quad (5.2)$$

where $\bar{U}$ and $\bar{V}$ are the mean zonal and meridional velocities, whereas $U' = U - \bar{U}$ and $V' = V - \bar{V}$ are the zonal and meridional velocity anomalies. In the present case, we averaged the year-long daily time series from both simulations; and from this averaging we computed the velocity anomaly. MKE allow us to analyze the current mean path and intensity. mEKE fields give us inferences on the mesoscale activity in the region: distribution, intensity and variability.

In Figure 5.2, we may observe two aspects. First the apparent agreement with what was reported by Eden and Dietze (2009) and Kang et al. (2016) that ocean current feedback acts as a mechanical damping for the exchange of kinetic energy between the mean state and its perturbations (first and second columns). As it is shown in Figure 5.2, the averaged eddy kinetic energy is less pronounced in the coupled simulation whereas the mean kinetic energy field presented stronger paths through the seamounts of VTR than the uncoupled simulation.

We also plotted climatological (1993-2016) fields of mEKE and MKE from the GOGRO altimetry data (Figure 5.2). There is a significant agreement among the MKE fields, specially between the coupled simulation (MKE_{A-O}) and the altimetry-derived field. However, the simulated state of MKE is weaker beside the Abrolhos Bank (18-19°S) meaning a weaker BC in this region for both simulations. The BC path intensifies south of 21°S in all three maps.

Regarding mEKE fields, there is a agreement with a high energy patch around TB (20-21°S) representing the VE activity throughout the year.

mEKE and EKE distributions and values in altimetry product are in better agreement with the coupled simulation than the uncoupled one. This is probably related with the stability effect that ocean current feedback according to Eden and Dietze (2009) and Renault et al. (2016).

Figure 5.2: Upper panel: Mean Kinetic Energy (MKE) field (averaged in 50 m) in $\text{cm}^2 \text{s}^{-2}$ for (a) the coupled experiment (MKE_{A-O}), (b) uncoupled experiment (MKE_A), and (c) the altimetry based climatology (1993-2016) (MKE_{GOGRO}). Lower panel: averaged Eddy Kinetic Energy (mEKE) fields (averaged in 50 m) in $\text{cm}^2 \text{s}^{-2}$ for (a) the coupled experiment ($mEKE_{A-O}$), (b) uncoupled experiment ($mEKE_A$), and (c) the altimetry based climatology (1993-2016) ($mEKE_{GOGRO}$)

5.2 Intrathermocline Eddy Analysis

We now investigate whether ITE (or second-baroclinic mode eddies) are features that occur in the VTR region using two different outputs of the regional simulations, each with a different wind stress forcing (See Chapter 2). The following section follows the work of McGillicuddy Jr (2015) where the author isolates the evolution of mesoscale intrathermocline eddies under the wind stress formulation ocean current feedback. McGillicuddy Jr (2015) developed eddy density structures through numerical modeling to investigate the density displacement that are typical of ITEs and, compared them with ITE observations found in the Gulf Current.

Thus, the analysis will be described as follows: (1) the tracking the mean eddy center locations for AE and VE; (2) to depict cross-bathymetry sections for every time step of the simulation; (3) the evaluation of the isopycnal displacements highlighted on the sections; and (4) the averaging of the sections to infer the main differences between the modeled eddies in both model outputs. For the latter stage of the analysis, we will also compute the density anomaly (ρ') and squared Brunt-Väisälä frequency (N^2).

5.2.1 Cross-bathymetry Density Evolution

Firstly, we focused our analysis on tracking the center of the mesoscale eddies in our field through the month. We aimed to frame, in both simulations, Abrolhos Eddy and a Vitória Eddy events. According to the statistical analysis we developed in Chapter 3, those eddies have mean life cycles of 32 days. Therefore, we should be able to locate those eddies in the monthly mean fields, even if there is some translation. We consider the monthly average horizontal velocities at 100m depth. At this depth we can observe the dynamics of the eddy far from the surface and where is effectively influenced by the Ekman pumping. Figure 5.3 shows the monthly-averaged velocity field at 100 m for February in both experiments. It is straightforward to select reasonable eddy centers for the analysis when it is possible to track the trajectories of both eddies. For the AE (VE), we established the center at 18.5°S,35.9°W (21°S,38.9°W) for coupled output and the 18.5°S,35.8°W (21.2°S,39°W) for the uncoupled simulation output.

After the eddy tracking, we were able to generate the cross-bathymetry section to investigate the recurrence and sharpness of density displacements of ITEs. The sections were taken at a one fixed latitude, and, the longitude as central location, a 2-degree wide range (Figures 5.4-5.11). The cross-bathymetry section will be fixed spatially to encompass the vertical density structure of the eddy as well as its evolution in the region.

This approach will guarantee that the isopycnal displacements that we focused are provided by the eddy/Ekman pumping dynamics rather than any other non-linearity by bathymetry, ocean currents, and internal waves.

The analysis provided us 29 sections (February) for each eddy and for each experiment.

Figure 5.3: Monthly averaged velocity fields (100m) for the month of February for coupled (left panel) and uncoupled model output (right panel), respectively.

The mesoscale dynamics is fairly similar in both simulations for this month. Therefore, we were able to carry our analysis considering both eddies as analogous features in the model output pairing. We analyzed the coupled and uncoupled outputs simultaneously. In doing so, it was possible to observe the evolution of isopycnal displacements concomitantly.

Our analysis was restricted to 4 days due to the great quantity of images produced. In Figure 5.4, we may observe the initial day of the analysis for the Abrolhos Eddy. The density configuration through the water column is similar with downward isopycnal displacements explained by geostrophic dynamics in an anticyclone (see Chapter 1).

However, as we move on to the eddy evolution after 10 days (Figure 5.5) when the eddy center is not perfectly aligned with the center of the section; we begin to observe sharper inferences of negative vertical velocities in the uncoupled output between 36°S and 35°S (density section in the lower panel). It is important to highlight that there is a sinusoidal displacement at shallower depths (30 to 40 m) which we assume to be interferences of non-linearities previously foreseen in the chapter. The reason of this assumption is the persistence of the pattern even in the initial condition of the analysis for both model outputs.

At mid-month evolution of the experiment, we observe the steeper dome-like feature within upper 50m of the coupled simulation output, which is a typical configuration of mode-water anticyclones (Figures 1.3 and 1.1). There is only an attenuation of the isopycnal displacements of the uncoupled simulation output (Figure 5.6). After 20 days (Figure 5.7), the AE translates further from the section and both density configurations at the region of mixed layer depth (MLD) are flattened out.

The VE is less symmetric and spatially longer than the AE. The initial condition for both outputs (Figure 5.8) did not depict a well-formed eddy structure, yet the vertical

Figure 5.4: Initial daily depiction of the density structure of Abrolhos Eddy. The top row of the figure shows the coupled output (A-O) whereas the bottom row depicts the uncoupled setting (A). Left column shows the velocity field at 100 m. The right columns shows the density variations in the water column within the established range of the section. The mean density deducted from the field is 1024 kg m^{-3} .

Figure 5.5: Density structure depiction of Abrolhos Eddy after 10 days. The top row of the figure shows the coupled output (A-O) whereas the bottom row depicts the uncoupled setting (A). Left column shows the velocity field at 100 m. The right columns shows the density variations in the water column within the established range of the section. The mean density deducted from the field is $\rho_o=1024 \text{ kg m}^{-3}$.

Figure 5.6: Density structure depiction of Abrolhos Eddy after 15 days. The top row of the figure shows the coupled output (A-O) whereas the bottom row depicts the uncoupled setting (A). Left column shows the velocity field at 100 m. The right columns shows the density variations in the water column within the established range of the section. The mean density deducted from the field is $\rho_o = 1024 \text{ kg m}^{-3}$.

Figure 5.7: Density structure depiction of Abrolhos Eddy after 20 days. The top row of the figure shows the coupled output (A-O) whereas the bottom row depicts the uncoupled setting (A). Left column shows the velocity field at 100 m. The right columns shows the density variations in the water column within the established range of the section. The mean density deducted from the field is $\rho_o = 1024 \text{ kg m}^{-3}$.

structure for density are different. The velocity field of the coupled simulation presents slightly lower values in magnitude than the uncoupled one. We may address this difference to the mechanical damping, previously referenced, caused by ocean current feedback (Eden and Dietze, 2009). However the mesoscale activity presented in the coupled velocity field is rather filamentous which explains the rapid variations in the density section of the coupled run. It is important to highlight the clear meander from the BC path influencing the VE dynamics. This does not only increase the velocity in the eastern flank of the eddy, evidenced in the velocity maps and the density section by geostrophic adjustment; but also infer the non-linearity of the eddy structure.

As we move to the experiment for its evolution of 10 days (Figure 5.18), when the eddies are more developed but not entirely aligned to the center of the section, it is possible to observe sharper subductions/obductions in the coupled output regarding the eddy (west of the section center) and the BC (east of the section). Moreover, the thinny isopycnal configuration in the coupled model are already perceivable. For 15-day evolution, the coupled experiment presents a well formed thinny structure besides the possible interactions with BC and TB bathymetry (Figure 5.10). As the eddy translates further from the section center location, the density configuration becomes similar to the both simulation outputs though the ITE signature is still observed in the section plotted on the upper panel of Figure 5.11.

5.2.2 Averaged Sections Analysis

To better depict the isopycnal displacements evidenced in the Section 5.2.1 for AE and VE, we employed a Lagrangian timely average in both of the eddy events for both model outputs. The averaging was based on the location of the eddy center throughout the month time series.

As represented in Figure 5.12, the AE in the coupled model output, showed a sharper positive isopycnal displacement within the MLD while a negative displacement at deeper depths. The mode-water eddy structure is presented for the AE and the denser waters are uplifted into depths of 25 to 50m in the coupled experiment. This behavior is as pronounced in the uncoupled experiment, where there is a sinusoidal pattern in shallower layers of the section already covered above, but no doming for a mode-water structure in the anticyclone.

For the lower pair of plots in Figure 5.12, density anomaly sections are depicted. This anomaly was computed as follows: (i) we computed the mean vertical density profile of the delimited section; and (ii) we subtracted this mean profile for the entire section of interest. Those sections are temporally averaged sections of these spatial anomaly patterns. This part of the analysis works allowed us to distinguish the ITEs structures as cores of positive and negative density values.

Figure 5.8: Initial daily depiction of the density structure of Vitória Eddy. The top row of the figure shows the coupled output (A-O) whereas the bottom row depicts the uncoupled setting (A). Left column shows the velocity field at 100 m. The right columns shows the density variations in the water column within the established range of the section. The mean density deducted from the field is $\rho_o = 1024 \text{ kg m}^{-3}$.

Figure 5.9: Density structure depiction of Abrolhos Eddy after 10 days. The top row of the figure shows the coupled output (A-O) whereas the bottom row depicts the uncoupled setting (A). Left column shows the velocity field at 100 m. The right columns shows the density variations in the water column within the established range of the section. The mean density deducted from the field is $\rho_o = 1024 \text{ kg m}^{-3}$.

Figure 5.10: Density structure depiction of Abrolhos Eddy after 15 days. The top row of the figure shows the coupled output (A-O) whereas the bottom row depicts the uncoupled setting (A). Left column shows the velocity field at 100 m. The right columns shows the density variations in the water column within the established range of the section. The mean density deducted from the field is 1024 kg m^{-3} .

Figure 5.11: Density structure depiction of Abrolhos Eddy after 20 days. The top row of the figure shows the coupled output (A-O) whereas the bottom row depicts the uncoupled setting (A). Left column shows the velocity field at 100 m. The right columns shows the density variations in the water column within the established range of the section. The mean density deducted from the field is $\rho_o = 1024 \text{ kg m}^{-3}$.

For the AE anomaly sections, we can observe a core of positive anomaly in both sections that represent the sinusoidal isopycnal displacement. However there is a stronger core in the coupled model output ranging from 25 to 75 meters, which represents the doming of the mode-water eddy structure. The lower negative cores of anomaly are depict the quasi-geostrophic negative velocities theorized in a anticyclonic mesoscale eddy for both model outputs.

Moreover, for the VE experiments, we analogously averaged density and the maps of velocity in the same aforementioned manner (Figure 5.13). In this particular experiment, the dynamics are more complex due to BC shearing in the eddy domain, the eastern flank of the eddy is also influenced by the BC path. Nevertheless, it is still possible to perceive the baroclinic eddy structure pattern in both of the sections of the model outputs: there is a enhanced thinny structure in the coupled output; whereas we can observe a quasi-geostrophic vertical velocity typical of cyclones interacting with the BC current in the uncoupled output section.

This is highlighted in the spatial density anomalies displayed in the lower panel of the Figure 5.13. While the uncoupled output presents a dipole-like anomaly structure due to the interaction of the BC current, there is attenuation on the positive anomaly core and a strengthening of the negative density anomalies. This configuration represents the intensification of negative velocities reaching depths of nearly 100 m within a cyclonic eddy.

5.2.3 Submesoscale Motion

Going further with our analysis, to infer stratification and homogeneity of the water column, we averaged squared Brunt Väissälä frequency along the sections of the eddy for both experiments in the same manner described in Section 5.2.2. In addition, mixed layer depth (MLD) were plotted to infer the effectiveness of eddy-induced Ekman pumping and damping/boosting of upwelling/downwelling activity, respectively. The sections for both AE and VE are depicted in Figure 5.14.

For MLD, we used the criterion employed by Monterey and Levitus (1997) where the MLD is delimited by the depth where the density changes by 0.125 kg m^{-3} from the surface values in the vertical density profile.

Figure 5.12: Lagrangian timely averaged Abrolhos Eddy of (a) eddy velocity, (b) density cross-section, and (c) spatial density anomaly for the coupled (left column) and uncoupled model (right column) outputs.

Figure 5.13: Lagrangian timely averaged Vitória Eddy of (a) eddy velocity, (b) density cross-section, and (c) spatial density anomaly for the coupled (left column) and uncoupled model (right column) outputs.

Figure 5.14: Lagrangian timely averaged Rossby number maps and stratification sections (N^2) for Abrolhos Eddy (a) and Vitória Eddy (b). The black line delimits the mixed layer depth.

Regarding N^2 , the AE sections were markedly similar. We can observe, however, a more pronounced dome in the center of the eddy in the coupled model output (Figure 5.14a), where the MLD shoaling is more pronounced. The mixed layer is shallower in the center of the eddy for both of the outputs. Nevertheless, there is a inference of upwelling within the mixed layer that is sharper in the coupled output. The d_{Ek} behaved similarly in both output, with values varying from 50 to 60 m of depth.

For the VE experiment, the MLD values and stratification around the eddy center (delimited by the black line in Figure 5.14b) are more discrepant. While there is a upward movement represented by the MLD in the uncoupled, we can observe the a valley strucutre in the center of the eddy for the coupled output. Based on N^2 , we can state that stratification on the coupled output is greater and the mixed layer is deeper. The d_{Ek} is shallower than in the AE, yet still deeper than the MLD estimate with values about 50 m depth.

We also plotted maps of Rossby number for both experiments. As the work of Mahadevan et al. (2008) points out, the role of submesoscale process through non-linear Ekman pumping in this case, is possibly relevant at flows with strong relative vorticity (ζ), i.e., those which compete, in magnitude, with the planetary vorticity (f). The authors infer that the loss of geostrophy is counterbalanced by the ageostrophic circulation on density variations in the presence of straining. In Figure 5.14, we can also observe the Rossby number distribution within the eddies. None of the AE experiments yield values greater than 0.4 of Rossby number for the averaged map. That is, planetary vorticity is at least three times larger than relative vorticity. VE presented higher values than AE event, but we only observed values of 0.5 or more at the coupled model output. As aforementioned, the coupled model model output showed a VE event with more associated ageostrophic flow than the uncoupled model output, which explains the higher values of Rossby number. Therefore, submesoscale phenomena is more likely to be relevant in the VE site.

5.3 Vertical Velocity Analysis

For the concluding part of the regional model analysis, we investigated the importance of the two terms of the total Ekman pumping in the formations of ITEs, focusing on vertical velocity and density horizontal distribution. We employed a similar tracking evaluation of the eddies as in Section 5.2.

We fitted ellipses that encompass the entire area of the eddies along their main path analogously to the sectional analysis (Figures 5.15-5.20) in order to observe the horizontal distributions of vertical velocity, density anomalies (at 25 m and 40 m of depth) and linear and non-linear Ekman pumping within the ellipse boundary.

We observed the positive spatial density anomaly in the AE core for both model out-

puts 5.15. This was already evidenced in the sectional analysis presented in Figures 5.4 to 5.7. However, the coupled simulations output yield larger positive spatial density anomalies. This corroborates with the statement that AE does have a mode-water formation if we take the ocean current feedback into account.

Analyzing the terms of Ekman pumping for the AE, we can affirm that non-linear Ekman pumping presents a strong similarity in both coupled and uncoupled simulation outputs even though uncoupled non-linear Ekman pumping is more intense in magnitude. We can observe the dipole structure that was parameterized in the Chapter 4 that may be annulled by the azimuthal averaging. This distribution, however, is summed with other smaller-scale filaments of positive and negative velocities that are located in the edge of the AE experiments.

For the coupled model output, the linear Ekman vertical velocity component, though smaller in magnitude than the non-linear Ekman pumping or vertical velocity, are relevant. The positive velocity core of W_τ following both the eddy center and the positive anomaly of density in the AE experiment evolution is represented in Figures 5.15-5.17.

The vertical velocity for the coupled simulation is smaller in magnitude than the uncoupled one. In addition, there is also a dipole of vertical velocity structure mainly in the center of the eddy similar to the non-linear Ekman input. This is better evidenced in Figure 5.15.

Positive spatial density anomalies are also evidenced for the VE experiments for both model outputs (Figure 5.18). However, there is a attenuation of this positive core in the eddy center for the coupled simulation, which is mainly depicted in Figures 5.19 and 5.20. In addition, at the eddy edge, strong negative spatial density anomalies are noted around the eddy center.

Regarding Ekman pumping terms, the fields of W_ζ in the coupled and uncoupled simulation outputs are not comparable. This may be due to the submesoscale activity which may overcome in the VE scenario, as our Rossby number maps suggest. Linear Ekman pumping in the coupled simulation output is negative in the eddy center and mostly positive in the flanks of the eddy. The negative velocity core seems to be the input attenuating the positive density dome in the eddy center.

Figure 5.15: Abrolhos Eddy evolution of horizontal velocity and distribution of vertical velocity (w), spatial density anomaly ($\rho - \langle \rho \rangle$), linear Ekman pumping (W_τ), and non-linear Ekman pumping (W_ζ). These sets represent the day 10 of February for the coupled (top plot) and uncoupled (bottom plot) model outputs. On the left column, the horizontal velocity fields are represented. For the first and second columns on the right (ellipsis maps), vertical velocity and spatial density anomaly are depicted for 25 and 40 m depth in two first rows. The last row shows the linear and non-linear Ekman pumping distribution.

Figure 5.16: Abrolhos Eddy evolution of horizontal velocity and distribution of vertical velocity (w), spatial density anomaly ($\rho - \langle \rho \rangle$), linear Ekman pumping (W_τ), and non-linear Ekman pumping (W_ζ). These sets represent the day 15 of February for the coupled (top plot) and uncoupled (bottom plot) model outputs. On the left column, the horizontal velocity fields are represented. For the first and second columns on the right (ellipsis maps), vertical velocity and spatial density anomaly are depicted for 25 and 40 m depth in two first rows. The last row shows the linear and non-linear Ekman pumping distribution.

Figure 5.17: Abrolhos Eddy evolution of horizontal velocity and distribution of vertical velocity (w), spatial density anomaly ($\rho - \langle \rho \rangle$), linear Ekman pumping (W_τ), and non-linear Ekman pumping (W_ζ). These sets represent the day 20 of February for the coupled (top plot) and uncoupled (bottom plot) model outputs. On the left column, the horizontal velocity fields are represented. For the first and second columns on the right (ellipsis maps), vertical velocity and spatial density anomaly are depicted for 25 and 40 m depth in two first rows. The last row shows the linear and non-linear Ekman pumping distribution.

Figure 5.18: Vitória Eddy evolution of horizontal velocity and distribution of vertical velocity (w), spatial density anomaly ($\rho - \langle \rho \rangle$), linear Ekman pumping (W_τ), and non-linear Ekman pumping (W_ζ). These sets represent the day 10 of February for the coupled (top plot) and uncoupled (bottom plot) model outputs. On the left column, the horizontal velocity fields are represented. For the first and second columns on the right (ellipsis maps), vertical velocity and spatial density anomaly are depicted for 25 and 40 m depth in two first rows. The last row shows the linear and non-linear Ekman pumping distribution.

Figure 5.19: Vitória Eddy evolution of horizontal velocity and distribution of vertical velocity (w), spatial density anomaly ($\rho - \langle \rho \rangle$), linear Ekman pumping (W_τ), and non-linear Ekman pumping (W_ζ). These sets represent the day 15 of February for the coupled (top plot) and uncoupled (bottom plot) model outputs. On the left column, the horizontal velocity fields are represented. For the first and second columns on the right (ellipsis maps), vertical velocity and spatial density anomaly are depicted for 25 and 40 m depth in two first rows. The last row shows the linear and non-linear Ekman pumping distribution.

Figure 5.20: Vitória Eddy evolution of horizontal velocity and distribution of vertical velocity (w), spatial density anomaly ($\rho - \langle \rho \rangle$), linear Ekman pumping (W_τ), and non-linear Ekman pumping (W_ζ). These sets represent the day 20 of February for the coupled (top plot) and uncoupled (bottom plot) model outputs. On the left column, the horizontal velocity fields are represented. For the first and second columns on the right (ellipsis maps), vertical velocity and spatial density anomaly are depicted for 25 and 40 m depth in two first rows. The last row shows the linear and non-linear Ekman pumping distribution.

Figure 5.21: AE's timely averaged horizontal distribution of vertical velocity (w), density anomaly ($\rho - \langle \rho \rangle$) and Ekman pumping components (W_τ , W_ζ) for (a) coupled and (b) uncoupled model outputs. The first and second rows present the vertical velocity and density anomaly for 25 and 40m, respectively.

Figure 5.22: VE's timely averaged horizontal distribution of vertical velocity (w), density anomaly ($\rho - \langle \rho \rangle$) and Ekman pumping components (W_τ , W_ζ) for (a) coupled and (b) uncoupled model outputs. The first and second rows present the vertical velocity and density anomaly for 25 and 40m, respectively.

Figure 5.23: Azimuthal averaging of vertical components of a Vitória Eddy event timely averaged. The upper plot depicts the coupled model output (a). The lower plot depicts the uncoupled model output (b)

The vertical velocity distribution has patterns of dipole similar to the distribution in the AE. Also, the values in magnitude are greater in the uncoupled simulation. However, the azimuthal inferences may allow us to give inferences on the effective values of vertical velocities and Ekman pumping in the eddy domain.

To observe these features in an averaged sense, we employed the Lagrangian averaging used in the previous section to observe the effect of ocean current feedback influence in the coupled and uncoupled model outputs during the entire month. In Figure 5.21, we can observe the averaged fields of vertical velocity, density anomaly and Ekman pumping components for the AE. It is straightforward to characterize the differences between the coupled and uncoupled model outputs, the positive density anomaly core is enhanced and reaches shallower depths in the eddy for the coupled model output. In addition, this positive density anomaly is located at the same position of the positive velocity core in the linear Ekman pumping map. For the uncoupled eddy structure, the same positive density anomaly region is not as pronounced. Vertical velocities patterns and non-linear Ekman pumping distribution and magnitudes are very similar in both model outputs. This corroborates the hypothesis that linear Ekman pumping is an important component in the mixed layer depth displacement.

Regarding VE for both model outputs (Figure 5.22), we observe the opposite behavior of the density anomaly distribution. The coupled model output presents a depression of the positive density anomaly in the eddy center differing from typical regular cyclone structure that is presented in (Figure 5.22b) of the uncoupled model output. This damping of positive vertical velocity suggests the influence of the negative Ekman pumping values yielded in the eddy center. However, the asymmetry of the VE velocity structure and the strong interaction with the Brazil Current motivate us to hypothesize that non-linear Ekman pumping plays a more important role in the isopycnal displacement structure.

Similarly to the analysis we obtained in the Chapter 4, we analyzed the vertical velocity components of VE in the azimuthal averaging perspective. For the month time series of both model outputs (Figure 5.23), we averaged the components of Ekman pumping as well as the vertical velocity in two levels of the water column. The levels for the analysis were chosen from the limits of the density anomaly sections depicted in Figure 5.13. The depths selected were 25 m and 125 m. In this sense, we analyzed the variation of vertical components in the eddy boundaries with the strong influence of Ekman pumping (25 m) and at a level with less Ekman pumping intensity (125 m).

The employment of the azimuthal average clarifies the resulting values of vertical velocity in the center of the eddy. For the coupled model output, we can observe values lower than -0.75 m day^{-1} for the surface level (25 m) in the center of the eddy, whereas values of nearly -0.5 m day^{-1} are yielded in the eddy center. The symmetric opposite occurs in the uncoupled simulation, where values of positive vertical velocity are yielded in the eddy center. However, the values of vertical velocity at 125 m are higher than the

25 m vertical velocity. Ekman pumping and the vertical velocity relation is not strictly linear, that is, they values are not resulting sums or subtraction of values. However, it is possible to observe that the patterns of non-linear Ekman pumping and the difference of vertical velocities between the two levels (25 and 125m) are very similar.

It is straightforward to understand the direct influence of non-linear Ekman pumping in the resulting vertical velocity input in the eddy boundaries. On the other hand, linear Ekman pumping is flattened out which is counterintuitive with the findings of Martin and Richards (2001) and the parametrization developed in Chapter 4. Regarding the uncoupled model, the quasi-geostrophy driven velocity seem to prevail in the eddy center. The non-linear Ekman pumping is smaller in magnitude and of opposite sign (0.2 m day^{-1}) with the coupled simulation values.

5.4 Chapter Highlights

In the presented chapter, we carried out an analysis to investigate the occurrence of ITEs in the mesoscale eddies of VTR surroundings and the influences of such density configuration in Ekman pumping terms. The analysis consists in using a two simulation outputs of a regional model (ROMS) with the same initial conditions and time frame, only differing by the wind stress forcing applied. In one simulation output, we employed the commonly use air bulk formula for wind stress (Equation 2.1) and coined this simulation as "uncoupled". On the other hand, the second simulation, coined "coupled", is forced by a daily wind stress field computed by the ocean current feedback air bulk described in Equation 1.3.

The model outputs were confirmed in volume transport of the BC and EKE/mEKE surface fields by literature reported values and altimetry dataset. The mean values of BC volume transport are similar to the ones found in literature. In addition, both simulations presented a fair agreement with the altimetry kinetic energy climatologies. As a matter of fact, the coupled simulation presented better agreement which was justified by the mechanical damping described by Eden and Dietze (2009) in model outputs that take ocean current feedback into account.

The first stage of the analysis looked further in the isopycnal displacements typically driven by the second-baroclinic mode eddies dynamics. For the coupled model outputs, the ITE density configurations could be observed by the cross-sectional density plots. Mode-water and thinny eddy structures were evidenced in the Abrolhos Eddy and Vitória Eddy events for the coupled simulation, respectively. VE presented a spatial density anomaly from 25 m to 125 m of depth, this downward isopycnal displacement characterizes the thinny cyclone configuration. What is more, a stronger positive anomaly for AE in the coupled simulation highlights the mode-water eddy isopycnal displacement in the shallower depths (25m to 50 m of depth). We could also observe a displacement in the

MLD comparing the uncoupled simulations: AE presented displacement in the order of 10^0 m, whereas we could observe differences in MLD of order of 10^1 in the VE event.

After that, we establish a elliptic threshold to evaluate the vertical velocities components as well as horizontal density distribution in the eddy boundaries. For Abrolhos Eddy in the coupled model output, the density anomalies seemed to agree with the distribution of linear and non-linear Ekman pumping. Positive density anomalies accompanied the positive values of linear Ekman pumping that are due to EWI, yet the maximum values of positive W_τ are 0.19 m day^{-1} . Moreover, the dipole structure of W_ζ also seems to follow the positive density anomaly values. This dipole structure is also highlighted in the vertical velocities of 25 m and 40 m in the eddy, but the vertical velocities present significantly higher velocity values. This patterns are weaker or non-existent throughout the evolution of the AE in the uncoupled simulation which corroborates the importance of ocean current feedback.

Regarding VE, we can also observe the positive density anomaly in the eddy structure for both output simulations. However, there is a attenuation of the values in the coupled model output. This corroborates with the negative values of vertical velocities yielded by EWI. However, non-linear Ekman pumping seems to be strongly influencing the values of density and vertical velocity. We employed azimuthal averaging plots for both model outputs. VE thinny structure seem to be related to non-linear Ekman pumping as the values of W_τ are lower than 0.1 m day^{-1} and W_ζ patterns are very similar to the vertical velocities subtraction from top and bottom levels of the isopycnal displacements as -1 m day^{-1} in the eddy center.

Chapter 6

Summary and Conclusions

In the present study, we investigated the eddy-induced Ekman pumping in the density structure of two main mesoscale eddies in the Vitória-Trindade surroundings. Ekman pumping can be divided in two terms: linear and non-linear Ekman pumping. The first term depends predominantly on the wind stress curl field, whereas the non-linear Ekman pumping depends in the gradients of relative vorticity as well as wind stress. However, we can only account the input of currents in the wind stress actual energy input if one takes the ocean current feedback into account. The effect of a mesoscale eddies on the wind stress curl is differential dragging at the opposite flanks of the eddy which results in values of curl of opposite sign to the quasi-geostrophic vertical velocity of such eddy. This effect is coined as Eddy-Wind Interaction.

Prior to the analysis of eddy-induced Ekman pumping, we statistically characterized the two mesoscale eddies using altimetry data in order to establish a AE and VE climatology. Two detection methods were used in this characterization: the first method used dynamical criterion to obtain frequency parameters (Okubo-Weiss) and the second used geometrical delimitation to obtain morphometric and, consequently, dynamical parameters (Winding-Angle). The analysis employed showed that the Eddies are persistent through the year. The AE and VE are similar in dynamical and morphometric aspects but are not similar in frequency. The author of this work states the strong correlation between VE events and the BC volume transport throughout the year.

After the employment of the statistical characterization, we parametrized, with the values obtained, the eddy-induced Ekman pumping in idealized gaussian eddies in order to understand the horizontal distribution of both vertical components in the eddy boundaries. We employed a homogeneous wind field based on the climatology directions of the VTR region and computed a wind stress field using the air bulk formula of ocean current feedback (Equation 1.3). This allowed us to analyze EWI as well as non-linear Ekman pumping distribution.

The values yielded by non-linear term were 10 times lower than the one yielded by linear term. However, W_{ζ} presented a dipole structure that in a symmetrical eddy is

annulled if we take in consideration a azimuthal averaging perspective, whereas W_τ values are enhanced in the eddy center which corroborates with the theory of EWI. Although the values of W_τ are important in this analyses in comparison with W_ζ ; the values are still smaller than expected in literature. This is probably due to the climatological term this parametrization is employed. As we are using geostrophic velocities retrieved by a statistical characterization, the values of velocity and radius tend to be lower resulting in smaller values of Ekman pumping.

As the last part of our analysis, we used a regional numerical model to address the differences in the density structure of mesoscale eddies when ocean current feedback is taken into account. We employed wind stress forcing with two air bulk formulas for each simulation output and analyzed the evolution of isopycnal displacements and vertical components in AE and VE for both model outputs. Ocean current feedback plays an important role in the decaying of first baroclinic-mode mesoscale eddies into intrathermocline eddies. AE presented a mode-water configuration mostly related with linear Ekman pumping in the coupled simulation output, as positive density displacements and could be observed following W_τ core in the time evolution of the eddy. VE presented a thinny isopycnal displacement configuration in the coupled output simulation. This, however, is probably related to the eddy-induced non-linear Ekman pumping as it is highlighted in a azimuthal averaging sense. Linear Ekman pumping presents no apparent important influences in the azimuthal averaging.

From our results and discussions, we must confirm/deny the preliminary hypothesis that we emphasized in the introductory chapter of this work. Based on the results of our regional model, we confirm that the wind regime and the mesoscale feature dynamics of the Vitória-Trindade Ridge generates intrathermocline eddies via eddy-induced Ekman pumping. Also, linear Ekman pumping component has relevant role in the total Ekman pumping input regarding ITEs formation, specially for Abrolhos Eddy events. For AE events, linear Ekman pumping generates/enhances dome of positive density displacements. For Vitória Eddy, the positive density displacements generated are mainly by quasi-geostrophic vertical velocities of a cyclonic that are flattened by the linear Ekman pumping. As a matter of fact, non-linear Ekman pumping presence higher influence on the VE structure than linear Ekman component. It is important to highlight that Vitória Eddy yielded strong submesoscale/ageostrophic circulation associated within its flanks that is worth further investigation.

We believe that additional studies should be carried out for the eddy-induced Ekman pumping and intrathermocline eddy dynamics. First, studies of primary production and nutrient distribution, not accomplish by this work due lack of time, are of utmost importance. The isopycnal displacement caused by ITEs in VTR surroundings, specially near the mixed layer depth, may provide high-nutrient waters to shallower depths and cause phytoplankton blooms in the case of an AE event. Mortality may also occur as

the symmetrical opposite phenomenon in the VE boundaries. Moreover, studies of the submesoscale input in the primary production are also essential if we are to investigate the primary production processes that occur in our study region. Finally, a more quantitative analysis of quasi-geostrophic vertical velocities in the eddy structure must be conducted in order to filter the actual influence of each term of Ekman pumping in mesoscale eddies and compare it with the vertical velocity components.

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Prof. Dr. Ilson C. A. da Silveira

Igor Uchôa Farias