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Seasonal and Momentum Balance of the Atlantic Equatorial Undercurrent

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UNIVERSITY OF MIAMI

SEASONAL MASS AND MOMENTUM BALANCE OF THE ATLANTIC EQUATORIAL UNDERCURRENT

By
Athanasia Papapostolou

A DISSERTATION

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ATLANTIC EQUATORIAL UNDERCURRENT

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An eastward subsurface current, the Equatorial Undercurrent (EUC), is part of the equatorial oceanic circulation in all oceans: permanent in the Atlantic and Pacific and seasonally present in the Indian due to the monsoonal wind circulation. This work focuses on the seasonal variability of the Atlantic EUC that supplies the equatorial upwelling with colder, salty and nutrient-rich waters, being therefore of significant importance for the surface heat budget and primary productivity in the Eastern Tropical Atlantic (ETA). The present study aims to answer two major questions: (i) what is the seasonal upwelling transport in the (ETA) that is related to the EUC transport and (ii) what is the forcing that controls the seasonality of the EUC. Methodologically, the first question is addressed in view of the seasonal mass balance while the second question is addressed by the seasonal momentum balance, both estimated with a unique data set of observations collected over the last 15 years, especially during the Tropical Atlantic Climate Experiment (TACE). Time series from the Prediction and Research Mooring Array in the Tropical Atlantic (PIRATA) and moorings that were deployed during TACE, observations from research cruises, Argo profiling floats measurements, drifter and altimetry measurements provide the core of the observational data used in this study. The 5-year daily output of a regional high resolution (1/4°x1/4°) model simulation of the Tropical Atlantic is also used as a testbed for various methodological sensitivity tests and as a comprehensive tool, when
observations are not sufficient to resolve certain aspects of the balances. There are significant differences between the western and the eastern parts of the ETA basin, in terms of the roles of zonal (mainly related to the EUC downstream mass losses) and meridional convergence in supplying the observed seasonal upwelling. In the western ETA, the upwelling shows a semiannual character with maxima occurring in July and December, while in the eastern region the upwelling cycle is mainly annual with a maximum in August. In both regions, the 50-300m EUC supplies significant part of the upwelling: 60% in the western box and 90% in the eastern box of the upwelling transport are supplied by zonal convergence, approximately ¾ of which is attributed to the EUC in both cases. On the other hand, it is the near surface meridional mass divergence (0-30m) that drives the near surface upwelling in the ETA, according to the traditional view of equatorial upwelling, confirmed by observations (drifters) and the model. From a momentum balance perspective, mid-basin in the ETA at 10ºW, non-linear advection terms are very important and seasonal changes in EUC strength can only be accurately reproduced when the non-linear advection terms are considered, in contrast with the linear dynamics that have been used to explain the EUC seasonality in numerous studies. However, below the EUC core, the seasonal zonal momentum balance is mainly linear, with the zonal pressure gradient being primarily responsible for accelerating and decelerating the lower EUC. Although the observational estimates of the mass and momentum balances involve methodological challenges, the model's ability to reproduce accurately the seasonal cycle of the EUC helps significantly in understanding the main processes controlling the seasonal cycles of upwelling and the EUC in the ETA. This work is one of the few observational studies that quantifies the seasonal cycle of upwelling in the ETA, one of the few attempts to study the
zonal momentum balance on a seasonal time scale, and the first observationally-based zonal momentum balance study in the equatorial Atlantic.
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Chapter 1 Introduction

The equatorial oceanic circulation demonstrates the presence of Equatorial undercurrents (EUCs) in all oceans. The EUCs are subsurface currents, flowing towards the east while the overlying surface South Equatorial Current (SEC) flows towards the west. The EUCs are permanent features of the equatorial circulation in the Atlantic and Pacific Oceans, and are seasonally present in the Indian Ocean due to the reversal of the monsoonal circulation. The key role of the Atlantic EUC is to supply the upwelling along the equator with cold, salty and nutrient rich waters that originate from the subtropics. The upwelling of the colder and nutrient rich waters along the equator enhances the phytoplankton bloom and therefore the primary productivity, highlighting the importance and broader impacts of upwelling and its relation to the EUC.

Over the last decade, research in the Tropical Atlantic has focused on understanding processes controlling sea surface temperature (SST) variability on intra-seasonal to interannual time scales, and its climate implications, referred to generally as Tropical Atlantic Variability (TAV) ([Xie and Carton, 2004]). While in the Pacific the presence of the El Nino Southern Oscillation (ENSO) has motivated many modeling and observational studies near the equator, in the equatorial Atlantic the number of studies to this day remains smaller. As an indication, the Prediction and Research Moored Array in the Tropical Atlantic (PIRATA), which is the source of the longest time series of equatorial hydrographic measurements, was implemented only in 1997 ([Servain et al., 1998; Boulès et al., 2008]), almost a decade after its Pacific counterpart, the Tropical Atmosphere-Ocean (TAO) array ([McPhaden et al., 1998]. The PIRATA array has had Acoustic Current
Doppler Profiles (ADCP) measuring zonal meridional currents in the upper 150m only at one location at 23°W since 2002, but the first long term current measurements at 10°W and 0°E were collected during the Tropical Atlantic Climate Experiment (TACE, 2007-2011). The TACE experiment returned approximately 4 years of EUC measurements, allowing for the first time a robust estimate of the transport seasonal cycle at multiple longitudes (23°W, 10°W and 0°E) using a combination of equatorial and off-equatorial moorings. The TACE current measurements form the basis for addressing the main research questions posed in this study.

The most striking feature in the Tropical Atlantic is the seasonal development of the equatorial cold tongue. During the peak of the upwelling season in boreal summer, a cold front of thermocline waters appears at the surface north of the equator (e.g. [Hastenrath and Lamb, 1977; Legeckis and Reverdin, 1987]). Recent studies ([Hormann and Brandt, 2007; Perez et al., 2012; Hormann et al., 2013; Hummels et al., 2013; Schlundt et al., 2014]) have highlighted the seasonal and interannual variability of the cold tongue and investigated features that are associated with the cold tongue front on multiple time scales. Such features include the intra-seasonal Tropical Instability Waves (TIWs) (e.g. [Jochum et al., 2004; Grodsky et al., 2005; Wu and Bowman, 2007; von Schuckmann et al., 2008; Lee et al., 2012; Perez et al., 2012], the seasonal phytoplankton bloom and primary productivity (e.g. [Monger et al., 1997; Pérez et al., 2005; Grodsky et al., 2008; Wang et al., 2013] and the Atlantic Niño, or zonal, mode (e.g. [Ruiz-Barradas et al., 2000; Keenlyside and Latif, 2007; Brandt et al., 2011]). Fundamental to understanding the variability of SST in the tropical Atlantic are processes related to upwelling, vertical mixing, and horizontal advection. In this study, we are particularly interested in the role of
the EUC in supplying seasonal upwelling to the near surface layers, since it acts as the main source for thermocline waters that are upwelled along the equator.

The EUC supplies the eastern part of the Equatorial Atlantic with waters that originate from the subduction zones in the subtropics. In Figure 1-1, taken from [Schott et al., 2004], the salty and recently ventilated waters are subducted in the blue shaded areas and can reach the equator either through the ocean interior or be directed to the western boundary where they then turn eastward into the EUC ([Zhang et al., 2003]). Three areas of upwelling (green shaded areas) are shown in Figure 1-1: (i) the equatorial upwelling region from the coast of Africa in the Gulf of Guinea all along the equator, (ii) the coastal upwelling zone off and north of the equator called the Guinea Dome (GD), and (iii) another coastal upwelling zone off and south of the equator, the Angola Dome. Waters that are upwelled in the equatorial Atlantic are moved toward the poles by off-equatorial Ekman
transports and returned to the subduction zones. This circulation pattern, known as the "Subtropical cells" (STCs; [McCreary and Lu, 1994]; [Liu et al., 1994]; [Schott et al., 1998]; [Zhang et al., 2003]), is asymmetric in the Atlantic ocean. The asymmetry in the STCs is attributed to the meridional Overturning Circulation (MOC), which results in most of the EUC - and the thermocline waters upwelled on the equator - originating from the southern hemisphere [Fratantoni et al., 2000];[Hazeleger et al., 2003]).

The TAV, especially the Atlantic Cold Tongue (ACT) region, is associated with poor seasonal predictability skill in climate models and poor representation of the mean state SST (e.g. [Richter and Xie, 2008; Richter et al., 2014] and references therein). As shown in Figure 1-2 from [Richter and Xie, 2008], climate models generally have warm SST biases in the eastern tropical Atlantic (ETA) compared to observations. The warm SST bias in the ETA is often attributed to wind stress biases during boreal spring that limit

Figure 1-2 : From [Richter and Xie, 2008]. Annual mean SST (°C) averaged between 2°S and 2°N for selected CMIP models (colored lines d-o), solid black (a) is ICOADS observations and solid gray (c) is the ensemble mean.
considerably the upwelling and as a consequence the simulated SSTs in the ETA are warmer than observed (\cite{Richter2008, Wahl2011, Richter2014}). Another known weakness of these models is their inability to capture the EUC strength and penetration into the Gulf of Guinea (GG). Usually the simulated EUC is too weak to penetrate all the way to the east in the GG and terminates at a longer distance from the African coast than shown in observations (\cite{Pacanowski1981, Wacongne1989, Blanke1993, Johns2014}).

Given the continuous time series observations in the EUC and accurate representation of the seasonal cycle of the EUC obtained in TACE, this work aims to further investigate links between the seasonal upwelling and EUC transport. The first question this study attempts to answer is how the seasonal cycle and transport of the EUC are related to the upwelling transport. Linked to this question is how the EUC responds to the seasonal wind stress forcing across the basin.

![Figure 1-3: Same as Figure 1-1 in Veronis, 1959. Depicts the “overturning cellular motion in a confined non-rotating basin.](image)

To understand how the surface forcing induced by the wind stress affects the EUC, we first consider the simple representation of the EUC given by \cite{Veronis1959}. As shown in Figure 1-3, easterly winds pile up water in the westernmost part of a confined basin, which sets a zonal pressure gradient. The pressure gradient at the west drives the water at
a depth below the direct influence of the surface wind towards the east, resulting in the EUC. The EUC strength in the time-averaged state should reflect the strength of the surface wind stress: where the westward surface wind stress is the strongest, the zonal pressure gradient will be greater (sea surface slope steeper) and the EUC will be stronger. But is the ZPG, set by the wind stress alone, the only forcing term that explains the seasonal variations of the EUC?

In variety of studies on the EUC seasonal cycle, before and after TACE (e.g. [Bourlès et al., 2002; Schott et al., 2003; Brandt et al., 2006, 2014; Hormann and Brandt,
it has been shown that the EUC exhibits a semi-annual seasonal cycle in the ETA, though the months of extrema differ depending on the location and the strength of the local winds. The seasonal core vertical displacement of the EUC is the most prominent feature of the variability on a seasonal time scale. As shown from observations at 23°W, the EUC core shoals to a minimum depth during boreal spring and plunges to greater depths during the rest of the year, reaching its maximum depth in boreal fall [Lumpkin and Garzoli, 2005; Brandt et al., 2006]. At 10°W, the EUC core also shoals in boreal spring and deepens during boreal fall and important features are: (i) a semi-annual transport cycle, confirmed by the recent study of [Johns et al., 2014] (Figure 1-4) and (ii) a deep extension of EUC during boreal summer, also present at 23°W (Figure 1-4). Figure 1-4 also reveals two maxima in the EUC core strength: one during boreal spring and one during boreal fall. At 0°E, the core is at its upper most location and stronger during April-May and then has a weaker secondary maximum during October.

Figure 1-5 shows the seasonal cycles of the westward surface wind stress provided by the Cross-Calibrated Multi-Platform (CCMP) Ocean Surface Wind Vector Analyses product [Atlas et al., 2011] on the equator as a function of longitude [Johns et al., 2014]. Comparing the seasonal cycle of the wind stress (Figure 1-5) and the EUC transport strength (Figure 1-4) one can observe that the EUC is weaker when the easterly wind stress
is maximum during late boreal spring and summer. Therefore, the local wind stress appears to be insufficient to explain the EUC seasonal cycle. On the other hand, enhanced upwelling into the mixed layer during boreal summer, in response to maximum Ekman divergence (caused by the maximum wind stress), could explain the EUC weakening. Zonal mass losses related to upwelling are consequently very important, and are expected to contribute to an understanding of the seasonal cycle of the EUC. A full understanding of the zonal momentum balance along the equator is required to determine what causes the EUC to accelerate and decelerate with season.

The relation of the seasonal cycle of the EUC with the different forcing terms that contribute to the momentum balance has long been under investigation in variety of studies mainly in the Pacific Ocean and mainly using models ([Bryden and Brady, 1985, 1989; Verstraete and Vassie, 1990; Hebert et al., 1991; Johnson and Luther, 1994; Qiao and Weisberg, 1997; Yu and McPhaden, 1999a]. To this day, a detailed analysis of the momentum balance of the EUC in the ETA using observations has not been conducted. Therefore, to answer what is forcing the EUC seasonal variability, this work attempts an analysis of the seasonal momentum balance.

The dissertation is structured in two main parts, each one related to the two overarching questions it attempts to answer. In the first part, included in chapter 2, this work addresses the question of how the seasonal cycle of the EUC transport and its downstream changes are related to the seasonal upwelling in the ETA. In the second part, in Chapter 3, this work addresses what forces the seasonality of the EUC and therefore the strength of the upwelling. The answers related to these questions are summarized in chapter 4 where future work is also discussed. In addition to the TACE data and various observational datasets, a
parallel analysis of both mass and momentum balance of the EUC is carried out using the output of a regional high resolution model (NEMO) and compared to the observational analysis.
Chapter 2  Seasonal upwelling in the eastern Tropical Atlantic and its relation to zonal circulation during the Tropical Atlantic Climate Experiment

2.1  Introductory remarks on the seasonal upwelling

Upwelling in the equatorial Atlantic is produced by meridional wind-driven divergence in the surface layer associated with the easterly trade winds, and possibly also by zonal divergence of the surface westward flow along the equator in the South Equatorial current (SEC). The traditional view is that the corresponding subsurface convergence that supplies the upwelling comes from meridional geostrophic flow toward the equator ([Wyrtki, 1981]; [Roemmich, 1983]) - forced by an eastward pressure gradient on and near the equator - and partly by zonal convergence within the EUC. These circulation components form the basis of the so-called "Subtropical cells" (STCs; [McCreary and Lu, 1994]; [Liu et al., 1994]; [Schott et al., 1998]; [Zhang et al., 2003]), in which off-equatorial Ekman transports are eventually returned by subduction in the subtropics leading to equatorward flow in the thermocline. This flow can either reach the equator within the ocean interior, or be directed to the western boundary where it then turns eastward into the EUC ([Zhang et al., 2003]).

In the Atlantic Ocean, the STC is asymmetric because of the presence of the Meridional Overturning Circulation (MOC), which results in most of the EUC - and the thermocline waters upwelled on the equator - originating from the southern hemisphere [Fratantoni et al., 2000];[Hazeleger et al., 2003]).

Vertical velocities in large scale oceanic flows are 4-5 orders of magnitude smaller than the horizontal velocities and cannot be determined directly using the instrumentation that
is routinely used to measure currents with the accuracy of a few cm/s. Observational estimates of equatorial upwelling have therefore relied instead on indirect techniques.

[Broecker et al., 1978], using bomb radiocarbon distributions and a box model between 15°N and 15°S, estimated near-surface upwelling in the equatorial Atlantic of 17 Sv. Shortly thereafter [Wunsch, 1984], also using radiocarbon in the same area and a quasi-geostrophic model, estimated upwelling of 7-10 Sv. More recently, helium isotope (³He or ⁴He) disequilibrium between the atmospheric and oceanic mixed layer has been used to infer upwelling rates ([Klein and Rhein, 2004]; [Rhein et al., 2010]). Using these techniques, [Rhein et al., 2010] estimated upwelling rates across the full width of the tropical Atlantic between 4°N and 4°S of 19±7 Sv (June) and 24±5 Sv (September), at times when equatorial upwelling is expected to be near its seasonal maximum. In another study, [Kadko and Johns, 2011] used ⁷Be as a tracer in the cold tongue region of the equatorial Atlantic to estimate local upwelling rates.

A more common practice in inferring upwelling estimates is to use a combination of current measurements, hydrography and surface winds ([Roemmich, 1983; Bryden and Brady, 1985; Gouriou and Reverdin, 1992; Meinen and Mcphaden, 2001; Meinen et al., 2001; Vauclair et al., 2004]). [Roemmich, 1983] reported that near the surface (above potential density surface of 25.5 kg/m³), 10 Sv of upwelling are needed to compensate for Ekman and geostrophic divergence between 8°N-8°S in the Atlantic. From a number of cruises conducted between 1983-1984, [Gouriou and Reverdin, 1992] reported a mean upwelling rate of 11-12 Sv between 1.5°N-1.5°S, for an area of the equatorial Atlantic between 40°W-5°W. Using moored current meter data from an array at 28°W and the continuity of mass equation, [Weingartner and Weisberg, 1991] estimated a mean vertical
velocity of $0.6 \cdot 10^{-5}$ m/s above the Equatorial Undercurrent (EUC). [Molinari et al., 2003] similarly used Acoustic Doppler Current Profiler data from repeat sections along 35ºW and find vertical velocities of $3.2 \cdot 10^{-5}$ m/s above 50m. All the above studies however report large uncertainties.

In the Pacific, [Wyrtki, 1981] estimated 50 Sv of upwelling between 5ºN-5ºS by separating the ocean into two layers: a surface layer where transports were estimated using Ekman divergence and a subsurface layer where meridional transports were estimated using geostrophic currents. [Bryden and Brady, 1985], using a diagnostic model for the equatorial Pacific and a combination of mooring data and hydrographic sections, calculated an upwelling transport of 22 Sv for the region between 5ºN-5ºS and 150ºW-110ºW. [Halpern and Freitag, 1987] reported a value of 40 Sv between moorings at 180ºW-90ºW on the equator and a mean vertical velocity of $2 \cdot 10^{-5}$ m/s at 110ºW. [Qiao and Weisberg, 1997] also using mooring data and mass divergence to report an average vertical velocity of $2.3 \cdot 10^{-5}$ m/s between 142ºW-128ºW and 1ºN-1ºS. Finally, [Meinen et al., 2001] in a box between 5ºS-5ºN, 155º-95ºW found a mean upwelling transport across 50m of $24 \pm 3$ Sv, using observations from 1993-1999 and box mass budget model.

Using models, [Philander and Pacanowski, 1986b] estimated mean upwelling in the equatorial Atlantic between 30-10ºW and 2.5ºN-2.5ºS of 13.7 Sv across 50m, and [Schott and Böning, 1991] show upwelling between 2.5ºN-2.5ºS west of 30ºW to the coast of Brazil and across 72m of 13.3Sv. A more recent modeling study by [Hormann and Brandt, 2007] reported relatively weak upwelling of 3.2 Sv across a potential density surface of 25.4 kg/m³ (~60m) for the eastern equatorial Atlantic region between 2.5ºN-2.5ºS and west of 23ºW.
Overall, there is quite a spread in the upwelling transport estimates derived from the variety of available methods. Most of the available studies have focused on annual mean upwelling rates, even though significant changes in equatorial upwelling are expected in association with changes in the seasonal winds and the development of the boreal summer cold tongue in the central and eastern Atlantic.

To increase the understanding of the circulation in the equatorial Atlantic and its role in regional climate variability, the Tropical Atlantic Climate Experiment (TACE) was initiated in 2007, during which equatorial and near equatorial moorings were deployed to monitor the seasonal and interannual variability of the EUC for a 4 year period ([Perez et al., 2013; Brandt et al., 2014; Johns et al., 2014; Kolodziejczyk et al., 2014]). A main goal of the program was to determine the response of the EUC to local and remote wind forcing and how its mass transport and seasonal cycle changed across the basin. The results of this program have provided robust new estimates of the EUC strength and seasonal cycle at three longitudes in the basin (23ºW, 10ºW and 0ºE) that form a key part of the results needed to perform the present study.

In this work, the TACE results are combined with other data sources to construct a seasonal mass balance in two areas of the equatorial Atlantic and produce indirect estimates of the upwelling transport. Because of the unique role that the EUC plays in supplying the equatorial upwelling, it is of interest to understand how exactly the zonal variations in the EUC transport are related to the upwelling transport. This work is, to the authors’ knowledge, one of the few observational studies that quantifies the seasonal cycle of upwelling in the tropical Atlantic. The output of Nucleus for European Modeling of the Ocean Océan PArallélisé (NEMO-OPA) regional high resolution model is also used. The
model captures very well the seasonal cycle of the EUC transport and model results are compared to the observational results and to assess the sensitivity and accuracy of the different methods.

Figure 2-1: TMI SST and TACE mooring locations. Colors show the TMI derived mean SST (in °C) during the cold tongue season of 2009 (April-May). The dashed black line shows the equator, the two boxes used for the mass balance estimates are also shown.

2.2 Data

2.2.1 Moored observations

The following analysis, uses zonal transport estimates derived from current observations obtained as part the Tropical Atlantic Climate Experiment (TACE) that took place from September 2007 to June 2011. Acoustic Doppler Current Profilers (ADCPs) were deployed at three locations (23°W, 10°W and 0°E) on the equator (0°N), as well as off the equator (0.75°N-0.75°S) to provide continuous current time series of the EUC (Figure 2-1). A time line of the moored observations at each longitude and of all other data products used in this analysis is shown in Figure 2-2. The off-equatorial moorings at 23°W were positioned at 0.75°S and 0.75°N and all the moorings at this longitude were maintained by GEOMAR (Kiel, Germany) [Brandt et al., 2014]). At 10°W, the off-equatorial moorings were located at 0.75°S and 0.75°N and were maintained by the University of Miami (US), while the equatorial mooring was maintained by the Institut de Recherche pour le Développement (IRD, France) ([Kolodziejczyk et al., 2014];[Johns et
The off-equatorial 0°E mooring was located south of the equator at 0.75°S and at this longitude the moorings (0°N and 0.75°S) were both maintained by the University of Miami [Johns et al., 2014]. Although the off-equatorial moorings were all deployed as part of the TACE experiment, the moorings on the equator are part of the Prediction and Research Moored Array in the Tropical Atlantic (PIRATA, [Bourlés et al., 2008]). In this study, the time series of the 40-hour low-pass filtered current velocities at the above-mentioned mooring locations are used primarily to quantify zonal transports, as described below in the Methods section. These data have already been extensively described by [Johns et al., 2014] and [Brandt et al., 2014], and Figure 2-3 (left panel) shows the composite seasonal cycle of the zonal velocity at the equator at all three longitudes derived
from the 3-4 years of continuous data at each site. Additionally, the temperature and salinity data (T/S) from the PIRATA moorings are used to quantify measurement errors associated with the dynamic heights estimated from the T/S gridded Argo product that are then used to calculate geostrophic currents (see below). Details on the error quantification in the dynamic height calculations using the PIRATA T/S profiles can be found in the Appendix B2.

2.2.2 Gridded Argo and altimetry products

Following [Johns et al., 2014], the Scripps Institution of Oceanography (SIO, http://sio-argo.ucsd.edu/) monthly gridded Argo product [Roemmich and Gilson, 2009] is used. The SIO product showed the best verification against PIRATA mooring data among several available Argo products. The SIO Argo data set has monthly T/S profiles at 1°x1° horizontal spatial resolution and 58 vertical levels (near the surface to approximately 2000m). The 4-D (longitude, latitude, depth and time) gridded Argo T/S monthly profiles from 2004 to 2011 are used to calculate geostrophic currents from dynamic heights (DH) within the area of interest (40°W to 10°E and from 5°N to 5°S). Equatorial geostrophy [Bryden and Brady, 1985; Picaut and Tournier, 1991; Lagerloef et al., 1999; Perez and Kessler, 2009] is used to estimate near-equatorial meridional currents from the Argo DH fields, following the approach in [Lagerloef et al., 1999]. The DH derived from the SIO Argo T/S profiles near the surface is compared to the absolute dynamic topography (ADT) from AVISO (Archiving, Validation and Interpretation of Satellite Oceanographic data) to test what reference depth is most suitable for the application of equatorial geostrophy. The
2.2.3 Shipboard measurements

Shipboard CTD and ADCP data from several cruises, shown on Table 2-1, are used to ground-truth and optimize transport estimates of the EUC derived from the mooring arrays. These data have already been used in several previous studies of the equatorial Atlantic circulation ([Bourlès et al., 2002];[Bourlès et al., 2008]; [Brandt et al., 2006]; [Brandt et al., 2014]; [Hummels et al., 2013]; [Johns et al., 2014]; [Kolodziejczyk et al., 2009]; [Kolodziejczyk et al., 2014]; [Perez et al., 2013]). In this work, an adapted form of the "optimal width" method of [Johns et al., 2014] and [Brandt et al., 2014] is used, to define the seasonal cycle of both eastward and westward components of the circulation at each end of the equatorial boxes. A detailed description of the calculation is provided in the Methods part of this section.

Table 2-1: ADCP and CTD data availability for the analysis of this work.

<table>
<thead>
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<th>LON</th>
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<td>10ºW</td>
<td>Le Suroît EGEE1</td>
<td>June 2005</td>
</tr>
<tr>
<td>23ºW</td>
<td>Ron Brown</td>
<td>July 2006</td>
<td>10ºW</td>
<td>Le Suroît EGEE2</td>
<td>September 2005</td>
</tr>
<tr>
<td>23ºW</td>
<td>L’Atalante</td>
<td>February 2008</td>
<td>10ºW</td>
<td>L’Atalante EGEE3</td>
<td>June 2006</td>
</tr>
<tr>
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<td>May 2009</td>
<td>10ºW</td>
<td>Antea EGEE4</td>
<td>November 2006</td>
</tr>
<tr>
<td>23ºW</td>
<td>Endeavour EN463</td>
<td>June 2009</td>
<td>10ºW</td>
<td>Antea EGEE5</td>
<td>June 2007</td>
</tr>
<tr>
<td>23ºW</td>
<td>Meteor M80/1</td>
<td>November 2009</td>
<td>10ºW</td>
<td>Antea EGEE6</td>
<td>September 2007</td>
</tr>
<tr>
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<td>November 2009</td>
<td>10ºW</td>
<td>Endeavour EN463</td>
<td>June 2009</td>
</tr>
<tr>
<td>20ºW</td>
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<td>February 2010</td>
<td>0ºE</td>
<td>Thalassa EQUA00</td>
<td>August 2000</td>
</tr>
<tr>
<td>23ºW</td>
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<td>April 2010</td>
<td>0ºE</td>
<td>Le Suroît EGEE1</td>
<td>June 2005</td>
</tr>
<tr>
<td>23ºW</td>
<td>Maria Merian</td>
<td>June 2011</td>
<td>0ºE</td>
<td>Le Suroît EGEE2</td>
<td>September 2005</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>0ºE</td>
<td>L’Atalante EGEE3</td>
<td>June 2006</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>0ºE</td>
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<td>June 2007</td>
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<td>0ºE</td>
<td>Antea FR18</td>
<td>September 2008</td>
</tr>
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<td></td>
<td></td>
<td>0ºE</td>
<td>Endeavour EN463</td>
<td>June 2009</td>
</tr>
</tbody>
</table>
2.2.4 Drifters and Argo floats

The shallowest depth that we obtain currents from the moorings or from shipboard ADCP data is typically 30m; above that depth we use a merged drifter and Argo float data set to estimate the near surface currents. The satellite-tracked drifters are from the Global Drifter Program, part of National Oceanic and Atmospheric Administration’s (NOAA) Global Ocean Observing System, and we use drifter data from the tropical Atlantic for the period from June 1997 to October 2013. Surface drifts from Argo floats are used in order to increase the near surface data coverage and are taken from the YoMaHa’07 data set ([Lebedev et al., 2007]), for the period from July 1997 to March 2015. We use the techniques described by [Lumpkin et al., 2013] and [Lumpkin and Johnson, 2013] to treat the drifter data set and follow the approach used by [Perez et al., 2013] to merge the drifters with the floats. Six-hourly winds at 10m height from the National Centers for

![Figure 2-3: Seasonal zonal velocity at the equatorial mooring locations. Left panels show the seasonal cycle derived from observations and right panels the seasonal cycle derived from the model at the 23°W (top), 10°W (middle) and 0°E (bottom)](image-url)
Environmental Prediction/National Center for Atmospheric Research (NCEP/NCAR) Reanalysis 2 are used to correct the float and drifter data set for wind-slip. This correction is applied as \( \tilde{u}_{\text{corr}} = \tilde{u}_{\text{uncorr}} - A \cdot \tilde{W} \), where for the drifters we use only those that have retained their drogues, with \( A = 7 \cdot 10^{-4} \) \cite{Niiler:Paduan:1995}, and for the floats we use \( A = 1.87 \cdot 10^{-2} \) following \cite{Perez:et:al:2013}.

2.3 Model

To complement the observational analysis and test our observational methods, a high resolution regional numerical model output is used. The NEMO \cite{Madec:2008} output used here comes from a simulation for the Tropical Atlantic region (20°S to 20°N and 60°W to 15°E) similar to the one used by \cite{Jouanno:et:al:2011b, Jouanno:et:al:2011a, Jouanno:et:al:2013}. The simulation uses a horizontal 1/4°x1/4° grid and 75 vertical levels, the primitive equations of the model are solved on an Arakawa C-grid and the model is run for the period 1990-2012, forced with the DRAKKAR Forcing Set (DFS5.2; \cite{Brodeau:et:al:2010}) and the European Centre for Medium-Range Weather Forecasts (ECMWF) Interim Re-Analysis (ERA-Interim) winds. At the boundaries it was forced with 5-day outputs from a North Atlantic simulation \cite{Treguier:et:al:2012} run for 1958-2012. The horizontal advection was parameterized using an Upstream Biased Scheme (USB) and vertical diffusion was parameterized using a Generic Length Scale Scheme (GLS; \cite{Umlauf:Burchard:2003}; \cite{Umlauf:Burchard:2005}) with k-\( \varepsilon \) turbulence closure and CANUTO-A stability condition, daily model outputs of the basic fields (zonal, meridional, vertical velocities, temperature and salinity), are used for the period 2008-2012.
The model was chosen because it reproduces the seasonal variability of the EUC better than most other currently available models of the tropical Atlantic (see [Johns et al., 2014] for a discussion of model comparisons to the observed EUC during TACE). Figure 2-4 shows the meridional structure of the zonal velocity from the observations and the model at all locations. The model captures reasonably well the meridional structure of the EUC, except that the model EUC is slightly stronger than observed at all longitudes and shows stronger penetration into the Gulf of Guinea (e.g., at 0°E in Figure 2-4) than the observations. The model also reproduces very well the main features of the seasonal cycle of the EUC, showing (i) a clear semi-annual cycle with two maxima at 23°W with primary

![Figure 2-4: Mean zonal velocity from shipboard ADCP (cruises, left panels) and from the model (right panels). From top to bottom: sections at 23°W, 10°W and 0°E](image-url)
maximum during spring and a secondary maximum during fall that is consistent with the observations (Figure 2-3), (ii) a semi-annual cycle with two maxima at 10ºW in observations, at approximately the same times of the year as at 23ºW, that is weakly reproduced in the model and (iii) an annual cycle at 0ºE in both model and observations and (iv) the presence at 23ºW and 10ºW and absence at 0ºE of deep EUC extensions during summer.

2.4 Methods

2.4.1 Mass balance in equatorial boxes

The schematic in Figure 2-5 shows the different components of transport that we estimate in this paper in two equatorial boxes, a "western" box between 23ºW and 10ºW and an "eastern" box between 10ºW and 0ºE. Applying continuity of volume to these equatorial boxes:

\[
W(z_1) - W(z_2) = -\int_{z_1}^{z_2} \left[U(x_E, z) - U(x_W, z)\right]dz - \int_{z_1}^{z_2} \left[V(y_N, z) - V(y_S, z)\right]dz \tag{Eq. 2-1}
\]

In Eq. 2-1, \( W \) is the upwelling transport at the upper \((z_1)\) and lower \((z_2)\) surfaces of the box, \( U = \int_{y_S}^{y_N} u(y, z)dy \) is the zonal transport per unit depth at the bounding longitudes \((x_E \text{ and } x_W)\) of the box, and \( V = \int_{x_E}^{x_W} u(x, z)dx \) is the meridional transport per unit depth at the northern \((y_N)\) and southern \((y_S)\) boundaries of the box. The first term on the right-hand side (rhs) of Eq. 2-1 represents the zonal transport divergence, and the second term is the meridional transport divergence. The zonal transport \( U \) can be further broken into an
eastward component, associated with eastward velocities only \((u>0)\), and a westward component associated with westward velocities only \((u<0)\). This break-up of the zonal transport aims to separate the EUC-related (eastward) transport from the total zonal transport, and is necessary to obtain optimized estimates from the mooring arrays of each of these contributions to the total zonal transport. Mathematically, this can be expressed by

\[ U_{\text{total}} = \int_{y_1}^{y_2} u_{\text{EUC}}^+(y,z) dy + \int_{y_1}^{y_2} u_{\text{WESTWARD}}^-(y,z) dy, \]

where \(u_{\text{EUC}}^+\) is \(u>0\) and \(u_{\text{WESTWARD}}^-\) is \(u<0\), respectively.

The box budgets are applied over a subsurface layer extending from \(z_2=300 \, m\) (the deepest common measurement level from the mooring arrays) to \(z_1=50 \, m\) depth, and over latitude limits from \(2^\circ\text{S} \) to \(2^\circ\text{N}\). Estimates for some of the transports are also computed.
using latitude limits of 1.2°S to 1.2°N, as will be described later. The choice of 50 m for the upper surface of the box is based on several considerations that will become clearer as the results are presented. Firstly, the meridional transport divergence is estimated by applying equatorial geostrophy at the northern and southern faces of each box, and therefore needs to be limited to depths below which there is a significant contribution to the meridional flow by ageostrophic wind-driven (Ekman) effects; secondly, 50 m is near the depth of maximum upwelling in the model, and we believe also in the real eastern equatorial Atlantic regime being studied here. For the calculations here, the upwelling (or downwelling) transport at 300 m is assumed very weak (zero). This assumption also evaluated within the context of the model and confirmed as it results in weak upwelling transport of the order of <1 Sv for most of the year.

2.4.2 Zonal transports from moored time series

To determine the zonal transports $U$ at the edges of the equatorial boxes, we apply the optimal width method (hereafter OW) as it is described and applied by [Johns et al., 2014] and [Brandt et al., 2014] (hereafter mentioned as J14 and B14). Briefly, this method determines - from a set of high-resolution shipboard ADCP transects at each longitude - the optimal widths to apply to discrete mooring velocity profiles to recover the net zonal transport over a specified latitude range. In J14 and B14, the EUC transport using the OW method is estimated for the EUC by following $U_{n}(x_{0}, z, t) = \int_{y_{N}}^{y_{S}} u(x_{0}, y, z, t) dy$. The transport is approximated by:

$$U_{\text{tr}(EUC)}(x_{0}, z, t) = \sum_{n} W_{n} u^{+}(x_{0}, y_{n}, z, t) \text{(Eq. 2-2)}$$

The $u^{+}(x_{0}, y_{n}, z, t)$ in Eq. 2-2 refers to the eastward (EUC, positive) velocities of a zonal
velocity profile at a mooring location at $x_0$ longitude and $y_n$ latitude, while $W_{n+}$ is the optimal width that better describes the meridional extend of the EUC at that location. J14 and B14 used the OW concept to estimate optimal widths $W_{n+}$ as best fit values of a least squares minimization using all the cruises around a certain longitude as their test bed. The reader is referred to the above papers for details of the method and error assessments. For the eastward zonal transports, the same optimal widths used in J14 and B14 studies (Table 2-2) are used here as well.

For the westward transports, widths that range from 0.1° to 1.4° degrees of latitude with a step of 0.1° (14 different widths) are used to estimate transports for all possible combinations of widths at each mooring location, by applying:

$$U_{tr(WESTWARD)}(x_0, z, t) = \sum W_n u^-(x_0, y_n, z, t)$$ (Eq. 2-3)

with $u^-(x_0, y_n, z, t)$ being the westward velocities at each mooring location for each available cruise section and $W_n$ the widths. Examples of how the transport profiles for the westward flow contribution are reproduced by the method are shown in Appendix A and similar examples for the eastward transport can be found in J14 (their Figs. 4 and 5). The

Figure 2-6: As in Fig. 6 of [Johns et al., 2014] but for the westward zonal transport. In all panels the comparison shown is between the westward transport estimates using the OW method and the actual transport from the sections (see text for more details on the rms differences of each case).
resulting optimal widths for the westward transport contributions are shown in Table 2-2. Applying the OW method to recover the westward transports taken directly from the sections, the mean value and standard error (ste) of the transports are recovered for all cases: (i) at 23ºW the mean westward transport is \(-1.5\pm0.6\) Sv for both the OW reconstructed and actual section transports, (ii) at 10ºW the westward transport is \(-2.2\pm0.6\) Sv for the 2 mooring reconstruction using the OW and the sections, while at the same longitude the westward transport is \(-2.1\pm0.6\) Sv using the merged mooring profile in OW and \(-2.2\pm0.6\) Sv from the sections alone, and (iii) at 0ºE the westward transport value is \(-0.6\pm0.3\) Sv for the OW reconstruction and \(-0.7\pm0.2\) Sv from the sections’ transport. The successful recovery of the westward transports using the OW is better illustrated in Figure 2-6. For both 23ºW and 10ºW cases in Figure 2-6, the scatter plots show a mild spread around the 1-1 ratio line, with the 0ºE case being the only one that could potentially include biases due to the small number of cruises (8) mainly during spring-summer months included in the calculations. However, the root mean square (rms) differences between the actual transport during each cruise and the transport estimated by the OW reconstruction, provide an actual metric for the uncertainties in the instantaneous zonal transports derived from the moorings. The rms difference for the westward transports from the OW method is 0.1 Sv for the case of 23ºW, 0.4 Sv for both cases of the merged and 2 mooring reconstructed profiles at 10ºW, and 0.2 Sv for 0ºE Figure 2-6). The rms differences of the westward transports due to the OW reconstruction when compared to the standard errors are in all cases more than 50% smaller.

In J14 and B14 the optimal widths are used in combination with the mooring zonal velocity time series profiles, with \(u>0\), to calculate EUC transport time series. Here, the
optimal widths for the westward transports are applied in mooring the zonal velocity time series, where \( u < 0 \), to estimate the times series of westward transport at each mooring location. Therefore, it is possible in the present work to account for the total zonal transport (EUC + westward), which is necessary to compute the box mass budgets as accurately as possible. The OW method is applied to estimate the time-varying zonal transports between 1.2°S to 1.2°N, a latitude range that almost fully includes the width of the EUC and was also used in the above-mentioned studies. The application of OW for a wider latitude range, as from 2°N to 2°S, is not possible since when it is tested for the westward transports, it produces biased transport estimates. Through visual inspection, the optimal widths that were estimated for the westward transports were not accurately reproducing the cruise section transport between 2°N to 2°S. However, for the calculation of meridional transport the ±2° off the equator limits are used and the possible errors in the mass budget associated with using only ±1.2° limits on the zonal transports are on the order of ~1 Sv based on comparison within the models’ framework and ~0.6 Sv according to the sections’ transport, both discussed in more detail in section 2.5.2.

Table 2-2: Optimal Widths used in this study to calculate zonal transports from the mooring records. The EUC values are the same used in [Johns et al., 2014] and [Brandt et al., 2014].

<table>
<thead>
<tr>
<th>MOORING LOCATIONS</th>
<th>OPTIMAL WIDTHS (° OF LATITUDE)</th>
<th>3 MOORINGS</th>
<th>2 MOORINGS</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>EUC WESTWARD</td>
<td>EUC WESTWARD</td>
<td></td>
</tr>
<tr>
<td>23°W 0.75°N</td>
<td>0.76 0.73</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>23°W 0°N</td>
<td>0.74 1.07</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>23°W 0.75°S</td>
<td>0.79 0.87</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
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<td>1.31 1.1</td>
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</tr>
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<td>-</td>
<td>-</td>
</tr>
<tr>
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<td>1.39 1.3</td>
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</tr>
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<tr>
<td>0°E 0.75°S</td>
<td>- -</td>
<td>0.99 1.07</td>
<td></td>
</tr>
</tbody>
</table>
2.4.3 Meridional geostrophic transport from Argo data

To estimate geostrophic transports across the meridional faces of the equatorial boxes at 2°N and 2°S, we use the SIO Argo analysis to determine dynamic height profiles relative to a reference level of 1000 m. Three different reference levels were tested: (i) the deepest SIO Argo measurement depth at 2000db (1979m), (ii) 1000db (1041 m), also used by [Lagerloef et al., 1999; Meinen and Mcphaden, 2001; Meinen et al., 2001]; (iii) 500db (541m) used routinely in [Bryden and Brady, 1985; Perez and Kessler, 2009]. The results for all three reference depths are very similar and the estimated transports do not vary

Figure 2-7: Seasonal climatology comparison between the Absolute Dynamic Topography Anomalies (ADTA, from AVISO shown as black line with error bars) and the Dynamic Height Anomalies (DHA, from SIO Argo) referenced at 2000db (blue), 1000db (red) and 500db (green) at 2°N (upper panels) and 2°S (lower panels) for each longitude at 23°W (left), 10°W (middle) and 0°E (right). The error bars on the ADTA represent only the statistical variability.
significantly between these choices within statistical errors. Figure 2-7 and Figure 2-8 show a comparison between the Argo-derived surface dynamic height anomalies (DHA) and AVISO absolute dynamic topography anomalies (ADTA) along 2°N and 2°S for each one of the reference depths. In Figure 2-7, there is a very good agreement at all locations in both the magnitude and the seasonal variability of the DHA and ADTA climatologies. The comparison between the DHA referenced at 1000db and the ADTA as a function of longitude (Figure 2-8), shows that the seasonal changes in surface topography are also captured very well. In this work, we choose the 1000db as the “level of no motion”, although from our analysis and Figure 2-7 reference depths of 500db and below provide a good approximation. The 1000db level produces reasonable near surface estimates of the seas level topography when compared to the AVISO products: the rms difference between
the Argo surface dynamic height anomaly referenced at 1000db (DHA) and AVISO absolute dynamic topography anomaly (ADTA) are 1.6 cm. Furthermore, the 1000db level is the one used by [Lagerloef et al., 1999] for the equatorial geostrophy approximation that is also used here. A more quantitative analysis and details related to this choice can be found in Appendix B2.

Within a few degrees off the equator, it is necessary to apply a modified form of the geostrophic balance that includes aspects of both \( f \)-plane (Eq. 2-4) and \( \beta \)-plane (Eq. 2-5) geostrophy [Picaut and Tournier, 1991]; [Lagerloef et al., 1999]; [Perez and Kessler, 2009]; [Perez et al., 2012]; [Perez et al., 2013]):

\[
\begin{align*}
fu_g &= -g \frac{\partial D}{\partial y}, \\
fv_g &= g \frac{\partial D}{\partial x} \quad \text{(Eq. 2-4)} \\
\beta u_g &= -g \frac{\partial^2 D}{\partial y^2}, \\
\beta v_g &= g \frac{\partial^2 D}{\partial x \partial y} \quad \text{(Eq. 2-5)}
\end{align*}
\]

In Eq. 2-4 and Eq. 2-5, \( u_g \) and \( v_g \) are geostrophic velocities and \( D \) is dynamic height. The dynamic height gradients \( \frac{\partial D}{\partial x}, \frac{\partial D}{\partial y} \) and curvature \( \frac{\partial^2 D}{\partial y^2}, \frac{\partial^2 D}{\partial x \partial y} \), are estimated using a central finite differences scheme on the \( 1^\circ \times 1^\circ \) Argo grid.

Following [Lagerloef et al., 1999], Gaussian weight functions are used expressed by:

\[
W_\beta = e^{-\left(\frac{\theta}{L}\right)^2} \quad \text{and} \quad W_f = 1 - e^{-\left(\frac{\theta}{L}\right)^2}
\]

where \( \theta \) is the latitude, \( L \) is a meridional decay length scale.

To allow a smooth transition between the \( \beta \)-plane and \( f \)-plane forms, then \( W_\beta + W_f = 1 \), such that \( W_\beta = 1 \) on the equator (\( \beta \)-plane geostrophy) and \( W_f \approx 1 \) away from the equator (\( f \)-plane geostrophy).

The meridional scale \( L \) is chosen to be 2.2\(^\circ\), the same as the optimal value found by
[Lagerloef et al., 1999], and also consistent with the ~2° values found in a similar analysis by [Perez et al., 2012, 2013] for the equatorial Atlantic. The resulting weights applied to the $\beta$-plane and $f$-plane geostrophic velocity estimates are $W_\beta = 0.43$ and $W_f = 0.57$ at 2° north/south. The results are not very sensitive to choices of $L$ that vary by about 20% from this value (from 1.8° to 2.6°), but application of pure $f$-plane geostrophy at 2° north/south would generally be expected to lead to overestimates of the actual meridional transports. We will also show later that application of this same formulation to the model, with $L = 2.2°$, provides a good estimate of the absolute meridional transport in the model at 2° north/south in the subsurface (50-300 m) layer.

2.5 Results

2.5.1 Upwelling transport inferred from subsurface mass divergence

To estimate upwelling through mass divergence we apply Eq. 2-1 using $z_1 = 50$ m and $z_2 = 300$ m, with the assumption that $W(z_2) = 0$. This is referred to hereafter as the “mass balance” upwelling. In Eq. 2-1 the zonal mass convergence is the vertically integrated transport difference between the western and the eastern boundaries of the box. Positive values of zonal convergence indicate an eastward decrease in the EUC (eastward) component of the zonal transport, and/or a westward decrease for the case of the westward transports (i.e., meaning that more westward is entering the box from the eastern limit than leaving from the western limit). Similarly, positive values of meridional mass convergence indicate net equatorward flow into the box.

Figure 2-9 shows time series of the zonal transport for the 50-300m depth layer after applying the OW method at each longitude. The need to include the westward transport is very clearly illustrated in Figure 2-9 by the difference between the EUC transport and the
total zonal transport. For example, in December 2009 and 2010 at 10°W the westward transport is strongly offsetting the total zonal transport resulting in net westward flow. Similar behavior is shown a little earlier in the year at 0°E and at 23°W a little later in the year. The absence of eastward flows below the EUC during boreal fall-winter could explain the reversal of the total transport in the subsurface box that we see in Figure 2-9 at all longitudes. A reversal of the total zonal flow suggests a reversal of the local zonal pressure gradient at that depth ([Kolodziejczyk et al., 2009]). The magnitude of the transports estimated using the OW method is at large confirmed by the available cruise transports at
the same time as the mooring record at all longitudes. The annual mean westward transport and standard error from 50-300m for the time series shown in Figure 2-8 is -2.1 ± 0.9 Sv at 23ºW, -2.4 ± 0.9 Sv at 10ºW and -2.1 ± 0.7 Sv at 0ºE. It is interesting that the value of the westward transports at 10ºW appears to be the largest, suggesting the strongest westward flow to be in mid-basin. However, considering the limitations due to the gap on the mooring time series at 10ºW, this record could be biased. The mean EUC transports and standard deviations are 10.3 ± 2.6 Sv at 23ºW, 8.4 ± 1.9 Sv at 10ºW and 5.4 ± 1.4 Sv,

Figure 2-10: Seasonal zonal transport using the mooring currents (in Sv/m). The panels (a)-(c) show the seasonal cycle of the zonal transport at each mooring location using the OW method and the lower panels (d)-(e) the zonal transport convergence at the western box ((d) is the zonal transport at 23ºW minus the zonal transport at 10ºW) and the eastern box ((e) is the zonal transport at 10ºW minus the zonal transport at 0ºE).

showing the strongest transport at the west.

Figure 2-10 shows the seasonal cycles of the zonal transport profile U as a function of depth at each longitude. The seasonal cycles represent monthly mean values of the zonal
transport profile (note that throughout our work any reference to a seasonal cycle refers to a monthly mean climatology unless otherwise mentioned). The panels (d) and (e) of Figure 2-10 show the zonal transport convergences within the western and eastern boxes. At all longitudes, from 30-150m the flow is purely eastward, dominated by the EUC, and below 150m the flow is westward from November to May at 23°W, from October to March at 10°W and almost all year long at 0°E. There is very close resemblance to the zonal velocity seasonal cycles on the equator shown in Figure 2-3 and the transports in the upper panels (a)-(c) of Figure 2-10 - the most striking feature of which is the deep extension of the EUC at 23°W and 10°W during late boreal summer-fall and the absence of this feature at 0°E.

The general weakening of the EUC from west to east is shown by the positive convergence between approx. 50-120m in the lower panels of Figure 2-10. The maximum EUC transport loss occurs in the western box during July-August and a little later during August-September in the eastern box, approximately at the EUC core depth (Figure 2-3, left panels). The absence of the deep extension at 0°E is the reason why in the eastern box there is more pronounced mass convergence than in the western box, which extends all the way to 300 m and signifies strong downstream mass losses. Transport divergence occurs weakly near the surface (seasonally in the western box and year-round in the eastern box), and below ~120 m during boreal winter and spring. The latter is associated with slight differences in the seasonal timing and intensity of the deep EUC extension and the westward flow periods that precede or follow it.

The seasonal meridional geostrophic transport at the northern (2°N) and southern (2°S) boundaries of the western and eastern boxes, from the surface to 300m is shown in Figure 2-11. In both boxes the near-surface meridional geostrophic transport is northward at 2°S
(Figure 2-11 b, e), with a brief exception during November in the eastern box where the flow reverses. At the northern side (Figure 2-11a), the meridional geostrophic transport in the western box is southward for most of the year while for the eastern box (Figure 2-11d) it remains southward for only half of the year (April to September and mid-November to January). These near surface reversals of the geostrophic flows imply a reversal of the zonal pressure gradient at these times, which can also be seen in Figure 2-8.

The meridional geostrophic convergence at the western box (Figure 2-11c) shows a net near surface convergence throughout the year resulting from the strong equatorward flows at the northern and southern boundaries and is much stronger than the respective

![Images of seasonal meridional geostrophic transport using Argo derived DH (in Sv/m). Panel (a) shows the western box meridional transport at the northern boundary 2°N, the southern boundary 2°S (c) and the meridional geostrophic transport convergence into the box (e). Right panels ((b), (d) and (f)) are same as the left for the eastern box.](image-url)
convergence (seen mainly from April to September) in the eastern box (Figure 2-11f). Near the surface this convergence will be exceeded by the surface wind-driven ageostrophic transport, resulting in the expected due to the prevailing easterlies, net near surface meridional divergence, that is later presented using drifter-derived currents. A direct estimate of this ageostrophic contribution to the meridional component cannot be accurately estimated that close to the equator using the classical Ekman theory [Perez and Kessler, 2009].

At depths below ~70m, there is convergent meridional flow in the western box from April-August and generally divergent flow from September-March. In the east, the flow is convergent except during boreal fall. In both regions, the seasonal convergence/divergence pattern over the 50-300 m depth range closely mirrors the meridional flow pattern at 2°S, where southward (off-equatorial) transport in boreal fall is a common feature.

Our seasonal upwelling estimates at 50 m derived from the subsurface (50-300 m) mass balances are shown in Figure 2-12 and Figure 2-13, where the net transports at each face of the equatorial boxes are also shown. In the western box (Figure 2-12), during April to September the total zonal flow shows little to no convergence and the upwelling transport is supplied almost entirely by equatorward meridional geostrophic convergence. From September to February, the reverse is true: there is strong zonal transport convergence within the EUC that predominantly supplies the upwelling, while the meridional convergence is comparatively weak. The weakest upwelling occurs in boreal spring (actually an implied downwelling of about 2 Sv), when both the meridional and zonal transports are weakly divergent. The seasonal upwelling cycle for this region has a semi-annual character, with maxima of 7.6 and 8.7 Sv in July and December respectively, and
minima in March-April (~2 Sv) and October (1.2 Sv).

The mean zonal transport convergence between 50-300m is 1.7±0.8 Sv, of which 1.5±0.4 Sv is due to EUC and 0.4±0.5 Sv due to westward zonal mass convergence, while the mean meridional geostrophic convergence is 1.3±0.8 Sv. On an annual mean basis, both zonal and meridional mass flows in the western 50-300m box play almost an equal role in supplying the upwelling, which has a mean annual value of 3.6±1.3 Sv (Table 2-3). The estimation of the error bars on the transports shown in the mass balance figures (shaded areas and bars) are described in detail in Appendix B.

Figure 2-12: Seasonal cumulative transport between 23ºW-10ºW and from 50-300m (in Sv). Left and right panels show the zonal transport at the left and right sides of the western box respectively while the middle upper and lower panels show the meridional geostrophic transport at the north and south sides of the western box respectively. The middle panel shows the total transport convergence (black solid line), the zonal transport in blue and the meridional geostrophic transport convergence in magenta. In all panels shaded areas and vertical bars represent errors.
In the eastern box (Figure 2-13), the inferred upwelling cycle is mainly annual in character, with maximum (7.3 Sv) in August and minimum (-2.2 Sv) in January. Two distinct regimes can be observed in terms of the contributions by zonal and meridional convergence: from November to June the meridional convergence is larger while from July to November the zonal convergence is dominant. For the November-June period, the zonal flow is divergent from mid-November to February, implying downstream mass gain and driving weak downwelling transport, while from March to July the zonal flow is weakly convergent and the upwelling transport is supplied by meridional geostrophic convergence. From July to November, the upwelling transport that peaks during August is supplied almost exclusively by the strong zonal mass convergence that peaks a month later than the
upwelling, while being offset by a weak divergence in the meridional geostrophic transport. The strong zonal convergence in the east from July to mid-November is associated with the presence of the deep extension in the EUC transport at 10°W and its absence at 0°E during summer (Figure 2-10), as well as the stronger westward inflow from the eastern boundary at 0°E from June to October/November. For the eastern box, the inferred annual mean upwelling is 3.0±0.6 Sv, almost all of which (~90%) is supplied by 2.8±0.6 Sv of zonal transport convergence (3.1±0.4 Sv from the EUC and -0.3±0.3 Sv of westward flow). The rest ~10% of 0.3±0.4 Sv is supplied by the meridional geostrophic inflow. Overall the EUC mass convergence appears to play a dominant role in supplying the seasonal upwelling in the eastern box, relative to the meridional convergence: it almost exclusively supplies the upwelling at 50m.

Table 2-3: Annual mean subsurface transports in the observations using the mass balance in the observations and in the model. We have included in the model’s case only the values for 2°N-2°S meridional integration limits.

<table>
<thead>
<tr>
<th>23°W-10°W Transport (Sv)</th>
<th>Observations</th>
<th>Model</th>
</tr>
</thead>
<tbody>
<tr>
<td>Zonal</td>
<td>1.9±0.7</td>
<td>4.5±0.5</td>
</tr>
<tr>
<td>Meridional (total)</td>
<td>-</td>
<td>0.9±0.5</td>
</tr>
<tr>
<td>Meridional Geostrophic</td>
<td>1.4±0.7</td>
<td>0.3±0.5</td>
</tr>
<tr>
<td>Upwelling</td>
<td>3.6±0.9</td>
<td>5.3±0.4 (50-300 m)</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>10°W-0°E Transports (Sv)</th>
<th>Observations</th>
<th>Model</th>
</tr>
</thead>
<tbody>
<tr>
<td>Zonal</td>
<td>2.8±0.6</td>
<td>3.7±0.6</td>
</tr>
<tr>
<td>Meridional (total)</td>
<td>-</td>
<td>-1.8±0.7</td>
</tr>
<tr>
<td>Meridional Geostrophic</td>
<td>0.3±0.4</td>
<td>-2.5±0.5</td>
</tr>
<tr>
<td>Upwelling</td>
<td>3.0±0.6</td>
<td>1.7±0.4 (50-300 m)</td>
</tr>
</tbody>
</table>
2.5.2 Model mass balance

The same mass balance is performed using the model (displayed in Figure 2-14) and compared with the observational estimates. Additionally, the model is used to evaluate some of the approximations and assumptions that are used in the observational analysis. In the model, the zonal and meridional convergences are calculated directly from the model's velocity field on the faces of each box, and the upwelling transport in each box is calculated from the model's actual vertical velocity at each depth level. In Figure 2-14, the net upwelling transport is displayed relative to 50m, therefore it is the difference between the

![Figure 2-14: Seasonal mass balance. Upper panels show the observational estimates, lower panels show the model estimates, while left panels show cumulative transport (50-300m) estimates in the western box and the right panels the same transport estimates in the eastern box. For the observations, the black line with error bars shows the inferred upwelling transport from 50-300m for which we assumed is zero at 300m and therefore should be the same at the upwelling transport at 50m and it explains why the upwelling transport at 50m is shown in the model’s case (black line with error bars in the lower panels).]
upwelling transport at 50m and at 300m ($W_{50-300m}$). The $W_{50-300m}$ transport is equivalent to the upwelling transport estimated from the observations, without assuming 0 transport at 300m. The error bars shown in the model’s mass balance transports are estimated following the same statistical procedures as for the observations (Appendix B.1 (a)). It is worth mentioning that for the case of the EUC 50-300m transports estimated from the model, we find integral time scales of variability comparable to the ones in the observations: 30 days for the transport at 23°W (33 days in the observations), 30 days for 10°W (27 days in the observations) and 14 days for 0°E (15 days for the observations).

The lower panels in Figure 2-14 show that in both western and eastern boxes the upwelling seasonal cycles in the model have considerably less variability than the corresponding observational estimates. This results from the fact that the zonal and meridional convergences tend to have nearly opposite phase seasonal cycles in the model. There is nevertheless a remarkable similarity in the seasonality of the meridional transport convergences between the model and the observations, as minima and maxima occur at the exact same times of the year. The same is true for the zonal convergences in the eastern box, but less so in the western box (the model-derived zonal convergence has lesser amplitude there and a minimum in July rather than April). The seasonality of the zonal transport convergence in the eastern box, while it remains convergent in the model all year long, similar to the equivalent observational estimate. The western box in the model shows also a convergent zonal flow all year long but with different seasonality than the observations from January to September. The November-December maximum in the zonal transport convergence in the western box is present in both the observations and model and has a comparable magnitude of 6-8Sv.
The mean upwelling transport at 50m in the model is $4.6 \pm 1 \text{ Sv}$ in the western box and remains almost constant from January to March, then reaches a maximum in June, minimum in August and another maximum later in October. This semi-annual character of the model’s upwelling in the west ($W_{50m}$) resembles the one in the observations, although in the observations the maxima occur in July and December, 1-2 months later. In October,

![EUC and westward Transport at 23°W](image1.png)

![EUC and westward Transport at 10°W](image2.png)

![EUC and westward Transport at 0°E](image3.png)

Figure 2-15: The comparison between the seasonal zonal transport in the model (bars) and the moorings (diamonds) at 23°W (top panel), at 10°W (middle panel) and at 0°E (bottom panel). The EUC transport in the model between 2°N-2°S is depicted using blue bars and the EUC transport between 1.2°N-1.2°S is depicted using cyan bars. The westward transport in the model is depicted using red bars for 2°N-2°S, and magenta bars between 1.2°N-1.2°S. In all cases the mooring derived equivalent EUC and westward transports between 1.2°N-1.2°S are shown with grey and yellow diamonds respectively.
there is clear minimum in upwelling and the amplitude of the seasonal cycle in much greater. In the eastern box, the mean upwelling transport at 50m in the model is 3.1±0.9 Sv and there is also evidence of semi-annual cycle, like in the western box. However, in the eastern box the model’s upwelling has slightly more pronounced amplitude than in the west and peaks clearly in June and in November. The minima and maxima in the upwelling in the east in the model occur at different months than in the observations. The observational upwelling in the east peaks in April and August (2 and 3 months later than the maxima in the model), strong throughout spring and summer and, as in the west, the amplitude of the seasonal cycle is much greater than the one in the model.

Within the model’s framework, there are three questions related to the observational methods that can be evaluated. The first is whether the zonal transports in the model between latitude limits of 1.2ºN to 1.2ºS - the same limits covered by the OW-derived zonal transports in our observational estimates - produce zonal convergence that are consistent with those estimated over the full limits of the equatorial boxes from 2ºN to 2ºS. Figure 2-15 shows the seasonal cycles of the EUC and westward transport in the model using both sets of limits versus our OW-derived transports. The seasonality in all transports is comparable, however, the transports from the model are larger at almost all times and all locations, mainly due to the stronger overall EUC in the model. The differences in the zonal transports when using the different meridional limits of integration can be as large as ~4 Sv, but the EUC transport differences are offset by the differences in the westward transports, such that the differences in the total zonal transports are not as pronounced. Table 2-4 summarizes the annual mean and standard error of the zonal transport convergences for the different limits of meridional integration. The total zonal transport
convergence between 50 and 300m is ~1 Sv greater in each of the equatorial boxes when using 2° vs. 1.2° limits, with eastward and westward transports contributing nearly equally to the difference in both cases. Although this ~1 Sv of zonal transport is larger than the standard error of the total transport in both cases of meridional limits, it is still 3-4 times smaller than the actual transport and we conclude that the difference in the limits does not change the mass balance significantly. Based on the model, we anticipate that our observational estimates of the zonal transport convergence and total upwelling could be underestimated by ~1 Sv. Testing this hypothesis for the observations, within for example the cruises’ framework, is only possible for the western box where a more adequate number of cruises is available. Using the cruises at 23°W and 10°W, one can derive a reasonable seasonal cycle of zonal transports. We find a mean difference of 0.6 Sv in the total zonal transport when different meridional limits (2° vs. 1.2°) are applied. For the eastern box, we are not able to make a meaningful test because there is barely an adequate number of cruise sections at 0°E, and the seasonal distribution of them is biased towards summer months (see previous section and Appendix A.

Table 2-4: Mean zonal transport convergences in the model using two different meridional limits of integration, 2°N-2°S for a closed mass balance and 1.2°N-1.2°S.

<table>
<thead>
<tr>
<th></th>
<th>23°W-10°W Transport (Sv)</th>
<th>2°N-2°S</th>
<th>1.2°N-1.2°S</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>EUC</td>
<td>3.6±0.3</td>
<td>3.2±0.3</td>
</tr>
<tr>
<td></td>
<td>Westward</td>
<td>0.80±0.2</td>
<td>0.3±0.2</td>
</tr>
<tr>
<td></td>
<td>Total Zonal</td>
<td>4.5±0.5</td>
<td>3.6±0.4</td>
</tr>
<tr>
<td>10°W-0°E</td>
<td>Transports (Sv)</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>EUC</td>
<td>3.0±0.4</td>
<td>2.5±0.3</td>
</tr>
<tr>
<td></td>
<td>Westward</td>
<td>0.7±0.2</td>
<td>0.1±0.2</td>
</tr>
<tr>
<td></td>
<td>Total Zonal</td>
<td>3.7±0.6</td>
<td>2.6±0.5</td>
</tr>
</tbody>
</table>
The second question that is tested using the model, is whether the meridional geostrophic transport between 50-300m is a good approximation for the total meridional transport. Meridional geostrophic velocities are computed in the model using the same formulation of equatorial geostrophy as used in the observations, where the model's absolute pressure field (SSH plus vertical hydrostatic pressure) is used in Eq. 2-4 and Eq. 2-5 rather than dynamic height. Figure 2-16 shows the vertical structure of the seasonal cycles of the total meridional transports at the northern and southern sides of the boxes. Figure 2-16 and can be compared to Figure 2-11, where the seasonal meridional geostrophic transports shown are estimated from the Argo derived DH. What is

Figure 2-16: Seasonal total meridional transport using the model (in Sv/m). Similar to Figure 2-11, left panels show the western box total meridional transport at the northern boundary 2°N (top), the southern boundary 2°S (middle) and the meridional transport convergence into the box (lower panel). Right panels same as the left for the eastern box.
immediately evident from Figure 2-16 and Figure 2-11 is that below approximately 50m, the meridional geostrophic transports follow very closely the total meridional transports. However, differences occur above 50 m that reflect the influence of wind-driven ageostrophic flow. At the northern sides of both western and eastern boxes, the poleward flow that results from the Ekman divergence due to the trade winds is generally confined to depths shallower than 30m, while at the southern sides this poleward flow penetrates to approximately 50m (or even possibly deeper at the western box). This asymmetric structure is consistent with what is known from previous studies about the meridional structure of the Tropical Cells ([Perez et al., 2013] and references therein). As expected, the poleward wind driven flows lead to a net divergence in the surface layer, which extends to about 50 m depth in the western box and to about 30 m in the eastern box. It is the sum of this

Figure 2-17: Seasonal upwelling transports as a function of depth. Upper panel shows the seasonal upwelling transport in the western box using the direct vertical velocities from the model and the lower right panel the same for the eastern box.
poleward wind-driven flow and the (generally) equatorward geostrophic flow in the surface layer that partially determines the magnitude of the equatorial upwelling. Based on the model, the seasonal upwelling profile determined from the model’s vertical velocities (Figure 2-17) is maximum at about 50m in the western and a little shallower (30-40m) in the eastern box upwelling. In the upper panel of Figure 2-17, in the western box, maximum upwelling contours of 6-7 Sv are shown exactly at 50m depth, while in the lower panel that represents the eastern box the maximum upwelling contours of 3-4 Sv are found in the 30-50m depth range. The depth location of the maximum upwelling is the main reason why the subsurface mass budget is estimated between 50-300m, while also important is the fact the wind-driven ageostrophic effects extend down to at least the depth at which one can locate the lower limb of the TC in the tropical Atlantic [Perez et al., 2013].

To quantify the overall accuracy of the geostrophic approximation below 50 m, the seasonal cycles of the 50-300 m meridional transport convergence determined from geostrophy versus the absolute model velocities are shown in Figure 2-18. The seasonal cycles in both boxes are in good agreement, however the geostrophic meridional convergence slightly underestimates the total meridional transports at all months of the year. The maximum differences are typically ~1 Sv and the annual mean differences for the western (eastern) boxes are 0.6 Sv (0.7 Sv), respectively. The exact reasons for which even below 50m there is a disagreement between the total meridional and the meridional transports are unknown, however there are possible explanations. First, the ageostrophic component of the meridional flow could penetrate deeper than 50m, at times of the year when the westward surface stress due to winds is strong. [Chereskin and Roemmich, 1991] for example examined basin wide ageostrophic transports at 11°N in the Atlantic Ocean to
report values of penetration depth for the wind driven transport of 100m. While their latitude of interest is far from the 2°N - 2°S regime our study focuses on, we can perform same type of comparison between the panels (c) in Figure 2-16 and Figure 2-17. For example, in July one can easily observe that the depth, at which the meridional geostrophic transport is almost the same as the total meridional, is below 75m. Also, non-linear terms with possible contributions are not accounted for in the linearized momentum equation used in equatorial geostrophy ([Lagerloef et al., 1999]) and furthermore, the blend between the two f-plane and β-plane regimes may not be optimal. Nonetheless, the comparison

![Figure 2-18: Comparisons of total meridional to meridional geostrophic transports and different upwelling transports. Upper panels show the comparison between the total meridional transport convergence and the meridional geostrophic transport convergence estimated using the L99 method. The lower panels show the comparison between the upwelling transport between 50-300m, the upwelling transport between 0-50m and the upwelling between 50-300m as the sum of meridional geostrophic and total zonal transports. Left panels represent western box transport, right panels represent eastern box transports.](image-url)
shows that the geostrophic method we are using in the observations gives very reasonable results in the model when it used to approximate the total meridional transport.

The third question addressed is whether the assumption of \( W=0 \) at 300m is valid in the model, and what kind of uncertainty in the observational upwelling estimates may result from this assumption. The lower panels in Figure 2-18 show the model upwelling at 50m versus the value that is obtained from the total horizontal convergence in the 50-300 m layer, which is equal to the upwelling transport at 50 m minus that at 300 m (i.e., the left-hand side of Eq. 2-1). The difference between these two curves therefore represents the model upwelling at 300 m. The annual mean \( W_{50-300m} \) upwelling in the west has a smaller value of 5.3±0.4 Sv (Table 2-3) compared to the one at 50m of 5.8±0.3 Sv (Table 2-5), due to ~0.5Sv of annual mean upwelling transport at 300m that is almost constant year-round.

In the east, the annual mean \( W_{50-300m} \) is 1.7±0.4 Sv (Table 2-3) compared with 2.2±0.3 Sv at 50m (Table 2-5), indicating a similar annual mean upwelling transport at 300m of ~0.5 Sv. However, in this case the seasonality of \( W_{50-300m} \) is rather different with a peak transport two months later in August that exceeds the \( W_{50m} \) value, associated with downwelling across 300m of >1 Sv in August and September (see Figure 2-17). These values of the upwelling at 300m on the order of +/- 1 Sv suggest that the assumption of no vertical flow across the bottom, although necessary to apply our methods, can be responsible for over- or underestimating the inferred seasonal upwelling through mass balance in the observations.
Table 2-5: Annual mean near surface transports from drifters (0-30m) using the Lumpkin and Johnson 2013 method and bin average seasonal climatology and the model (0-30m and 0-50m).

<table>
<thead>
<tr>
<th></th>
<th>23°W-10°W Transports (Sv)</th>
<th>10°W-0°E Transports (Sv)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Zonal</td>
<td>Meridional</td>
</tr>
<tr>
<td>Drifters</td>
<td></td>
<td></td>
</tr>
<tr>
<td>(0-30m)</td>
<td>-0.2±0.2</td>
<td>5.6±0.5</td>
</tr>
<tr>
<td>Bin average</td>
<td>0.1±0.4</td>
<td>4.2±1.4</td>
</tr>
<tr>
<td>MODEL</td>
<td>0-30m</td>
<td>-0.1±0.3</td>
</tr>
<tr>
<td></td>
<td>0-50m</td>
<td>0.8±0.4</td>
</tr>
</tbody>
</table>

2.5.3 Other upwelling estimates

2.5.3.1 Equatorial Divergence from TACE moorings

Another upwelling estimate can be calculated using only the TACE mooring zonal and meridional currents to infer vertical velocities at the center of the boxes. The vertical velocity $w$ is consequently estimated by integrating the continuity equation from a depth $z_{\text{max}}$ where we assume $w=0$ to any given depth $z$: $w(z) = \int_{z_{\text{max}}}^{z} \left( \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} \right) dz$ (Eq. 2-6). Here $z_{\text{max}}=300\text{m}$ as for the mass balance inferred upwelling.

In Eq. 2-6, the $\frac{\partial u}{\partial x}$ term is estimated by linear regression using the equatorial $u$ velocity profiles at all 3 mooring locations. Fitting a quadratic curve between the $u$ equatorial profiles was also tested and produced comparable results to the linear regression, we therefore concluded that there was no apparent reason to choose a quadratic fit instead of linear. One should keep in mind that the moorings are separated with almost 10 degrees of latitude and although this “rough approach” just described is the best estimate with the available mooring configuration, on the other hand it could contain biases relative to a more local estimate. Since 10°W is almost a mid-point in the linear fit, it can be considered a
more representative location to derive the divergence estimate from moorings and should be less prone to fit biases than 23ºW or 0ºE. Therefore, in this work $\frac{\partial u}{\partial x}$ is only estimated at 10ºW and consequently only $\frac{\partial v}{\partial y}$ is estimated at this same location. The $\frac{\partial v}{\partial y}$ term is calculated using the off-equatorial (0.75ºN-S) moorings’ meridional currents. The resulting $\frac{\partial v}{\partial y}$ is located to the center of the distance between moorings, the equator and then Eq. 2-6 is applied. The resulting vertical velocity at 10ºW is shown in Figure 2-19 along with the associated error bars, described in detail in Appendix B.4.

The equatorial upwelling velocity at 10ºW (Figure 2-19) has a structure that shows upwelling close to the surface all year long with a strong maximum during boreal summer and downwelling during boreal fall. Comparing this vertical velocity with the zonal velocity, it is evident that when the upwelling is maximum on the equator then the EUC weakens.

Assuming that at 10ºW the upwelling is a good representation of the upwelling over the whole 23ºW-0ºE longitude range, multiplying $w$ at 50m to the appropriate zonal and meridional width should give the associated upwelling transport (in Sv). As zonal width, we take the zonal distance between the moorings (23º); the meridional width is the distance between the off equatorial moorings (1.5º). However, this results in very high transport values, especially during the summer maximum. For example, during July the scaled upwelling transport at 10ºW is 33.3±17.7 Sv, when the standard errors are estimated from as the vertically integrated sum of standard errors at each depth level, or 33.3 ± 13.5 Sv where the error is the sum of standard errors at each depth level. Meanwhile, the sum of
the west and east box mass balance upwelling transport in July is 13.7±6.9 Sv, almost half the size for what is essentially a smaller meridional width box. Testing this in the model, again taking the month of July as an example, the mass balance derived upwelling transport at 50m is 7.3 Sv, and the value obtained by multiplying the model's vertical velocity at the equator by the same zonal and meridional width of 23° and 1.5°, is 8.2 Sv. This comparison shows that the mass balance upwelling is in good agreement with the scaled upwelling transport derived from the direct $w$ in the model (1.2 Sv of difference or ~15%) and the
mass balance derived seasonal upwelling transport provides a good estimate. Therefore, our seasonal $w$ derived from local mass divergence at 10ºW should contain biases related to the methodology used that yields very large error bars on the monthly estimates of Figure 2-19, resulting in only a few monthly estimates significantly different from zero, and is not as representative of an estimate as the mass balance one.

2.5.3.2 Drifter based upwelling estimates

Another estimate of equatorial upwelling can be made from surface divergence derived from drifter data. To accomplish this the drifter data are mapped into a rectangular grid that has an optimized zonal resolution instead of a regular grid, where the zonal width is approximately 4° and the meridional resolution is chosen to be 1°; that way in almost all grid boxes there is an adequate amount of drifter days per month of the year. To estimate the seasonal climatology of the near surface currents simple bin averages of the data per month of the year in each grid box result in the climatological values (hereafter called the “bin-average” method). We compare this with the methodology described in [Lumpkin and Johnson, 2013] where in each grid point the data is decomposed into mean, annual and semiannual harmonic components to obtain seasonal zonal and meridional currents (hereafter called the LJ13 method). The upwelling estimates from the drifters between 2ºN-2ºS are estimated by integrating the zonal and meridional currents on each face of the equatorial boxes to get the zonal and meridional transports and then by multiplying the horizontal transport divergence by 30m. This follows exactly Eq. 2-1 and the choice of 30m is a logical one since the drifter currents are measured at 15m and therefore should be approximately representative of the 0-30 m depth-averaged currents.

The zonal and meridional transport divergence derived from the drifters is shown in
Figure 2-20, along with their sum which provides an estimate of the upwelling transport at 30m. Immediately one can see that most of the upwelling is related to the meridional mass divergence (Figure 2-20), a result of the prevailing trade winds. The zonal mass divergence only weakly impacts the total divergence leading to a net upwelling transport during the entire year. The LJ13 method at large confirms the seasonality and magnitude of the meridional divergence in both boxes, and although there are some differences in the zonal part of the divergence between the two methods they are not significant enough to give significant differences in the inferred upwelling transport. Errors in the drifter-derived
upwelling (see Appendix B.5) depend on the number of measurements per grid point and their spread around the monthly mean value, and although the grid used aims to incorporate measurements such that there is data at any given climatological month in each grid box, in cases with a lower number of data points the standard errors can be large. In the area of the Tropical Atlantic, the drifter data coverage is relatively sparse [Lumpkin and Garzoli, 2005; Lumpkin and Johnson, 2013], however drifters provide one of the few sources of near surface current measurements and we include this estimate of drifter-derived upwelling for comparison to our other upwelling estimates. There are aspects of the drifter cycles that do not seem physically reasonable (e.g. the minimum in upwelling in the western box in summer when the winds are maximum), which may be attributable to the large error bars, particularly those of the meridional component of the divergence. The error bars on the mass balance derived upwelling are about a factor of two smaller in comparison, and for this reason the mass balance results are taken to be more representative of the actual annual cycle of the upwelling in both the western and eastern equatorial boxes.

The annually averaged upwelling estimates from the drifters compare more favorably with those from the subsurface mass balance method (Table 2-2 and Table 2-5). The near surface mean annual upwelling in the western box from the bin-average method is 4.3±1.4 Sv, compared to 3.6±0.9 Sv from the mass balance method. In the eastern box, the respective upwelling transports are 3.7±1.1 Sv (drifters) and 3.0±1.1 Sv (mass balance). It is also noteworthy that the contributions to the annual mean upwelling by zonal and meridional convergences are considerably different for the surface layer and the subsurface layer. The drifter derived upwelling is almost entirely supplied by meridional divergence, which accounts for nearly 100% of the annual mean upwelling in the western box and more
than 80% in the eastern box, (Table 2-5). Conversely, in the subsurface layer it is mostly supplied by zonal mass convergence, mainly through the EUC flow and especially in the eastern box. Essentially, the near surface meridional mass divergence (also the upper limb of TCs) is the “cause” of the near surface upwelling, but once the surface waters have moved poleward, the balancing convergence in the subsurface layers is not supplied proportionally by local meridional convergence (the lower limb of TCs) but by zonal
convergence in the EUC. This is one of the reasons why there is such a strong signal of the subtropical water masses in the equatorial Atlantic upwelling zone.

Finally, we combine the results from the two equatorial boxes to show the net upwelling cycles across the ETA spanning from 0ºE to 23ºW and their associated uncertainties, for both the drifter and mass balance derived estimates (Figure 2-21). In the drifter-derived upwelling estimates (both bin-average and LJ13) the seasonal cycle has an annual character, and the peak transport occurs during May approximately 2 months earlier than in the mass balance estimate. The mean annual estimates of the drifter derived upwelling compared to the mean annual estimates of mass balance upwelling are larger by 0.7-1.7 Sv, depending on the method (LJ13 or bin-average) and the box (west-east). Using the model to compare the mean upwelling transport at 30m with the upwelling transport at 50m the upwelling at 30m is greater in the eastern box, where the maximum upwelling is located between 30-50m. Despite the differences in the seasonality and the expected differences in magnitude of the mean upwelling between the mass balance and drifter upwelling transports, as shown very clearly in the bottom panel of Figure 2-21, the annual mean of all transports compares favorably within their error bars.

Table 2-6 summarizes all the mean upwelling estimates both from the observations and the model in the eastern tropical Atlantic (ETA) from 23ºW-0ºE. It is very interesting how the mean annual upwelling using a variety of methods and limits of integration shows values, both in observations and the model that range from 6.6 Sv to 9.6 Sv. Given the errors in each estimate that we have analyzed in detail in this work, these numbers are relatively close to each other and they provide a very robust estimate of the upwelling in the ETA and are one of the key points of this study.
Table 2-6: Mean annual upwelling transport estimates (in Sv) in the eastern tropical Atlantic (23ºW-0ºE).

<table>
<thead>
<tr>
<th>Method</th>
<th>Transport (Sv)</th>
</tr>
</thead>
<tbody>
<tr>
<td>OBS</td>
<td></td>
</tr>
<tr>
<td>Drifters (0-30m)</td>
<td>9.3±0.7</td>
</tr>
<tr>
<td>Bin average</td>
<td>8.0±1.8</td>
</tr>
<tr>
<td>MODEL</td>
<td></td>
</tr>
<tr>
<td>0-30m</td>
<td>7.7±0.4</td>
</tr>
<tr>
<td>0-50m</td>
<td>8.0±0.4</td>
</tr>
<tr>
<td>SUBSURFACE (50-300m)</td>
<td></td>
</tr>
<tr>
<td>OBS</td>
<td></td>
</tr>
<tr>
<td>Local Divergence at 10ºW</td>
<td>9.6±3.8</td>
</tr>
<tr>
<td>Mass Balance 2ºN-2ºS</td>
<td>6.6±1.1</td>
</tr>
<tr>
<td>MODEL</td>
<td></td>
</tr>
<tr>
<td>2ºN-2ºS</td>
<td>7.0±0.6</td>
</tr>
</tbody>
</table>

2.6 Discussion

The traditional view of the equatorial upwelling is that the near surface meridional mass divergence moves surface waters poleward resulting in upwelling that is compensated at depth by subsurface meridional convergence. This work uses observations to quantify the important role that the zonal convergence plays in supplying the upwelling especially in the eastern part of the tropical Atlantic, previously mainly shown in modelling studies [Philander and Pacanowski, 1986b; Verstraete, 1992; Hazeleger and de Vries, 2003; Hormann and Brandt, 2007]. The zonal convergence (Figure 2-10) in the western box is strong during October-February throughout the whole water column below ~70 m, where strong zonal convergence below the EUC core reflects the slight phase offset of the deep EUC extensions between 23ºW and 10ºW. It is interesting that while the strongest zonal convergence at the EUC depth occurs in July, the zonal divergence at greater depth at that time acts to cancel out the net zonal convergence resulting in the peak of zonal transport convergence occurring later in the year, in November, approximately at same time as the second upwelling maximum. In total, in the western box, the zonal transport convergence
contributes to the upwelling continuously for 7 months of the year, from mid-August to March. In the east, there is a similar pattern of upwelling supplied through zonal convergence; it also lasts almost 7 months but it is phase shifted by 2 months, starting in June. This phase shift is again the result of the deep extension of the EUC, more precisely the lack of the deep extension in the EUC during boreal summer at 0º.

To first order the EUC is driven by the zonal pressure gradient, which is the dominant forcing term below the surface wind-driven layer. In Figure 2-22(a), we show the ZPG along the equator calculated from Argo data at the 3 longitudes, as well as the mean ZPG over the limits of the two equatorial boxes (23°W-10°W and 10°W-0°E). The ZPG in Figure 2-22(a) is estimated by calculating the slope \( \frac{\partial DH}{\partial x} \) of a linear fit that is produced using dynamic heights that are centered within ±5º of the longitude of interest. The ZPG forcing is positive (eastward) in the upper water column throughout the year at 23°W but changes to mostly negative (westward) at 0°E. In between, at 10°W, it remains positive for most of the year but exhibits a brief reversal in October. Therefore, the ZPG is acting to drive the upper part of the EUC in the western part of the study domain and generally retarding it in the east. At all locations, except perhaps 10°W, there is clearly a different behavior of the ZPG near the surface and at depth. At 23°W the ZPG below ~100m is positive from April to September and negative from October to February, and has a similar cycle at 10°W except that the negative ZPG phase starts earlier (in September) and lasts only until about December. These periods of positive and negative ZPG correspond closely
to the build-up and decay phases of the deep EUC observed at both longitudes (Figure 2-10). That is, when the ZPG is positive, the deep flow below the EUC core is accelerating from westward to eastward, and vice-versa when the ZPG becomes negative. Thus, it appears that the lower part of the EUC is responding directly to the ZPG in a quasi-inviscid
manner. As noted above, it is the slight difference in timing between the deep EUC extensions at these longitudes - and the associated ZPG changes - that lead to the large subsurface zonal convergence between 23°W and 10°W in boreal fall. The deep ZPG at 0°E shows a similar though somewhat weaker annual pattern but with yet an earlier occurrence of the negative ZPG phase, in July to October. A corresponding deep westward flow develops, that peaks in October, and although no obvious deep extension of the EUC occurs at 0°E, the westward flow reaches a minimum and the EUC appears to deepen in late spring following the positive ZPG phase.

While the near surface ZPG is almost in phase with the seasonal cycle of the surface wind stress, the seasonal local acceleration and deceleration of the upper EUC is not explained by the ZPG alone. This indicates a more complicated zonal momentum balance where other terms such as vertical turbulent stress and non-linear accelerations may be important in the balance. In particular, the weakening of the EUC that is observed in boreal summer is inconsistent with the occurrence of a maximum eastward zonal pressure gradient at that time, at both 23°W and 10°W. The surface westward wind stress induces vertical dissipation of momentum that generally should act to decelerate the zonal flow while the presence of non-linear advection terms can modulate the seasonal acceleration/deceleration patterns as well ([Wacongne, 1989]). Therefore, while the weakening of the EUC in summer can be thought of as a natural consequence of the maximum upwelling in summer - and corresponding loss of EUC waters to the mixed layer - this is from a mass balance perspective. To understand the actual causes of the seasonal acceleration and deceleration of the EUC, it is necessary to understand the full momentum balance of the EUC, which is the goal of the 2nd part of this work.
The meridional transport is accurately represented by the meridional geostrophic transports below 50m and in this study, we show robust evidence of that using the model’s framework (Figure 2-18). By definition, off the equator where f-plane geostrophy holds, the zonal pressure gradient and the meridional geostrophic currents should have a very similar structure while closer to the equator the β-geostrophic meridional current will have a similar structure to the meridional gradient of the zonal pressure gradient when applying equatorial geostrophy (Eq. 2-4 and Eq. 2-5). Comparing Figure 2-11 with Figure 2-22 (a), there is a very close correspondence between the average 23ºW-10ºW ZPG and the meridional transport in the western box. This means that the beta contribution does not fundamentally affect the meridional flow, except for a small reversal of the ZPG to westward in September and the meridional divergence to convergence a little earlier in August. In the eastern box, there is also a good qualitative comparison between ZPG and meridional geostrophic flow, although similar phase shift of one month between the fall reversal of the ZPG and meridional flow reversal from convergence to divergence in August especially below 100m.

A common feature in both western and eastern boxes is the downstream zonal mass loss maximum in fall. This strong fall zonal convergence is mainly related to the EUC transport maximum that occurs in the western and central ETA basin in fall and to a lesser degree to the westward transport maximum approximately at the same time of the year (Figure 2-15 and Figure 2-23). A number of observational studies provide robust evidence of the existence of a fall maximum EUC transport in the western ETA, west of ~20ºW [Katz et al., 1981; Hisard and Hénin, 1987; Brandt et al., 2014; Johns et al., 2014], and models also accurately represent this western ETA fall EUC maximum [Schott and Böning,
1991; Hazeleger et al., 2003; Arhan et al., 2006; Hormann and Brandt, 2007]. In the central (~10ºW) ETA the fall EUC transport maximum is also present in both observations and models [Arhan et al., 2006; Kolodziejczyk et al., 2009, 2014; Johns et al., 2014], however in the eastern ETA (east of 0ºE) there is weak to no evidence of a fall EUC transport maximum [Arhan et al., 2006; Hormann and Brandt, 2007; Johns et al., 2014; Kolodziejczyk et al., 2014]. This fall EUC transport weakening shown in the above-mentioned studies, is confirmed quantitatively in this work. In both model and observations, the fall EUC mass losses contribute ~67% of the total zonal mass convergence in the western (Figure 2-23(a) and (c)) and ~75% in the eastern boxes (Figure 2-23(b) and (d)), while the rest (33% and 23% respectively) is supplied by westward
transport convergence. From a mass balance perspective, the EUC transport losses of ~5 Sv (out of the total ~7 Sv of zonal mass convergence) in November in the western box in the observations are attributed to ~4.5 Sv of upwelling transport and weaker meridional geostrophic outflow of 2.5 Sv, predicted by the weakly negative ZPG in November (Figure 2-22(a)). In December in the western box the 5 Sv of the EUC transport losses are attributed to the very strong upwelling of ~8Sv, since the upwelling transport is also supplied by weaker poleward meridional geostrophic transport (~1 Sv) as a result of the reversed weakly positive ZPG in that month (Figure 2-15). The model in the western box shows a similar behavior to the observations, where the EUC mass losses (~ 5 Sv out of the 8 Sv of total zonal mass loss) are attributed almost exclusively to the upwelling in October-November (~ 7 Sv) and weak meridional outflow (~1Sv) (Figure 2-15). However, a very different pattern is responsible for the downstream EUC mass losses in the east: maximum meridional poleward transport in September (total meridional transport of 6.5 Sv in the model and 5 Sv of meridional geostrophic transport) plays an equal role to the upwelling of ~ 5 Sv in the observations and an exclusive role in the model (~0.5 Sv of upwelling) in causing downstream EUC mass losses (Figure 2-23 (b) and (d) and Figure 2-15). This strong fall meridional outflow in the 50-300m depth is directly related to maximum of the negative ZPG in the eastern box (Figure 2-22(a)) which in turn is related to the westward wind stress reversal to eastward at 0°E. While previous studies used the ZPG to interpret the EUC fall maximum, as a 1-2 month lagged response to westward winds intensification in June-July (e.g. [Philander, S. G. H., Pacanowski, 1981; Hisard and Hénin, 1987; Arhan et al., 2006]), the present work shows the relation of the EUC downstream weakening in the eastern part of the ETA to the off-equatorial meridional flow that is predicted by the
vertical structure of the ZPG during fall (September) in both observations and model.

In terms of the actual upwelling transport, there are also different seasonal patterns at depth and near the surface. The upwelling is the result of the wind-driven poleward mass divergence due to the prevailing trades and both mass balance derived upwelling and local $w$ at 10ºW peak in July-August, almost at the same time the westward wind stress is maximum (Figure 2-22(b)). However, the eastern box reveals a different behavior where the upwelling transport peaks clearly in August, while the westward wind stress peak 2 months before in June. In the 50-300m EUC layer, the zonal flow lags the ZPG by approximately 90º (quadrature), as it would if the flow was purely inviscid. The deeper (50-300m) ZPG is not set up by the same 1st and 2nd mode wave dynamics that control the upper layer, but instead is controlled by higher than 3rd baroclinic modes ([Katz, 1997; Brandt and Eden, 2005; Han et al., 2008; Hormann and Brandt, 2009; Brandt et al., 2016a]

The lagged response in the eastern part has been previously reported in studies that support that the upwelling in the Gulf of Guinea is not forced by the local wind system but it is rather forced by the winds further to the west, that they indeed peak later in the year (e.g. [Adamec and O’Brien, 1978]; [Servain et al., 1982]; [Busalacchi and Picaut, 1983]; [Picaut, 1983]).

2.7 Summary and conclusions

In this work, we use a large number of observations obtained over the last 15 years to study the seasonal upwelling in the eastern tropical Atlantic. We use the concept of conservation of volume to derive the mass balance in two boxes, a western box (23ºW-10ºW) and eastern box (10ºW-0ºE), bounded by the longitudes where TACE moorings were deployed, and between 2ºN-2ºS latitude. The zonal and meridional transports are
derived along the faces of each box using the TACE mooring data and equatorial geostrophy applied to Argo data in lieu of mooring or other current measurements at the north-south limits of the boxes. We then apply the continuity equation to derive upwelling as sum of total zonal and meridional geostrophic convergences. Another estimate of the upwelling (at 10ºW only) comes from local horizontal velocity divergence and provides inferred vertical velocities at the mooring locations that we then use to estimate the upwelling transport. We also include in our analysis a near surface upwelling transport estimated using mass divergence of the near surface currents obtained from a merged drifter and YoMaHa float data set.

Principal conclusions from the study are:

- Using the subsurface mass balance method, we find significant differences between the western and the eastern boxes in terms of the roles of zonal and meridional convergence in supplying the observed upwelling. Both zonal and meridional geostrophic mass flows in the western 50-300m box are important in supplying the inferred upwelling (the zonal flow contributes the most), which has a mean annual value of 3.6±3.3 Sv. In the eastern box, the annual mean upwelling is estimated as 3.0±2.2 Sv, approximately all of which (~90%) is supplied by zonal transport convergence (3/4 from the EUC and 1/4 from westward flow), and the remaining ~10% of it is attributed to meridional geostrophic inflow.

- The model is used to test the assumption that was made in the observational mass balance estimate, that at 300m, the upwelling transport is negligible. Although within the model’s framework at 300m the upwelling transport is non-zero at both western and eastern boxes, it is found that the vertical transport across the
subsurface box is relatively weak and the assumption would not yield significantly biased estimates. The seasonality of the total meridional transport is well represented by its geostrophic component and in turn, the seasonality of the geostrophic transport in the model resembles closely the seasonality of the geostrophic transport in the observations.

- The near surface meridional mass divergence drives the near surface upwelling but once the surface waters have moved poleward, the upwelling waters from deeper layers are not supplied as strongly by the meridional convergence but by the EUC, explaining the strong signal of the subtropical water masses in the equatorial Atlantic upwelling zone.

- The seasonality of the upwelling can be partially explained by the seasonality of ZPG, but the lagged response of the zonal flow to the ZPG can only be explained for the deeper part of the flow (below 100m) with the present observational analysis leaving room for future studies.
Chapter 3 Seasonal momentum balance in the eastern Tropical Atlantic

3.1 Introductory remarks on the seasonal momentum balance

The area of the Atlantic Equatorial Undercurrent (EUC) shows multiple scales of temporal variability, from intra-seasonal manifested as Tropical Instability Waves (TIWs), to interannual and decadal manifested as the Atlantic Nino mode. In between these scales, the seasonal variability of the Atlantic EUC is the dominant time scale of variability [Brandt et al., 2016b]. The seasonally varying EUC flows along the equator as a result of a zonal pressure gradient set up by the prevailing easterly winds; at the depth where the frictional effects due to surface wind stress are smaller than the zonal pressure gradient, the eastward flowing EUC appears ([Johnson and Luther, 1994; Qiao and Weisberg, 1997]).

In the Eastern Tropical Atlantic (ETA), the EUC exhibits a semi-annual cycle. The semi-annual cycle is stronger in the western part of the basin than in the east, following the pattern of the stronger westward zonal wind stress there ([Bourlès et al., 2002; Schott et al., 2003; Brandt et al., 2006, 2014; Hormann and Brandt, 2009; Kolodziejczyk et al., 2009, 2014; Johns et al., 2014]). The months of seasonal extrema in the EUC strength differ depending on the location and the strength of the local winds, but are not necessarily in-phase with the local winds ([Johns et al., 2014; Brandt et al., 2016b]). Typically, the EUC is characterized by seasonal vertical migration of its core with shallower depths observed during spring and late fall and maximum depths during summer to early fall. ([Kolodziejczyk et al., 2009, 2014; Brandt et al., 2014; Johns et al., 2014]). The seasonal
characteristics of the EUC are well documented within the above mentioned observational studies, but a complete understanding of what is driving the seasonal changes in strength and vertical structure of the EUC at various locations in the ETA is not yet available.

To better understand the processes driving the EUC seasonality, the zonal momentum balance (ZMB) can be used as a diagnostic tool. On the equator, where the Coriolis force vanishes, the seasonal acceleration and deceleration of the EUC can be investigated within the context of the ZMB equation:

\[
\frac{\partial u}{\partial t} = -\left( u \frac{\partial u}{\partial x} + v \frac{\partial u}{\partial y} + w \frac{\partial u}{\partial z} \right) + \frac{1}{\rho_o} \frac{\partial p}{\partial x} + \frac{\partial}{\partial z} \left( k_{m,V} \frac{\partial u}{\partial z} \right) + \frac{k_{m,H}}{LDF} \nabla_H^2 u .
\] (Eq. 3-1)

In equation (3-1), \( u, v \) and \( w \) are the zonal, meridional and vertical velocities respectively, \( \rho_o \) is the density of seawater, \( p \) is the pressure, and \( k_{m,V} \) and \( k_{m,H} \) are the vertical (V) and horizontal (H) coefficients of eddy diffusivity of momentum respectively. Equation 3-1 refers to the large scale zonal flow and the term “eddy” is used here to express the contributions of smaller scales than the annual mean. The term on the left-hand side (LHS) is the local acceleration of the zonal current, which will result from any imbalance on the right-hand side (RHS) between the advection of zonal momentum (“ADV”), the zonal pressure gradient (“ZPG”) and the vertical and lateral diffusion (“ZDF” and “LDF” respectively).

If equation (3-1) is averaged in time and vertically integrated, and both lateral diffusion and the non-linear advective terms are assumed to be small, the balance reduces to [Sverdrup, 1947; Kessler et al., 2003]:

\[
\bar{P}_x = \bar{\tau}_0^x .
\] (Eq. 3-2)

In equation 3-2, \( P_x \) is the vertically integrated ZPG and \( \tau_0^x \) is the surface zonal wind.
stress, where the surface boundary condition of the ZDF term is \(k_{m,v} \left( \frac{\partial u}{\partial z} \right) = \frac{\tau_0}{\rho_0} \) (3-3). The balance described in 3-2 is otherwise known as the linear “equatorial” Sverdrup momentum balance [Sverdrup, 1947] and has been used in a number of studies to investigate the forcing balance between surface winds and pressure gradient set up on the equator by those winds (i.e. [Mangum and Hayes, 1984; McPhaden and Taft, 1988; Yu and McPhaden, 1999b; Brown et al., 2007]).

The relation between ZPG and the strength of the EUC motivated some of the earliest observational studies mainly in the Pacific Ocean [Wyrtki, 1974a, 1974b; Halpern, 1980; Mangum and Hayes, 1984]. Connections between the seasonality of the trades and ZPG in these early studies revealed an off-phase relation between the winds and pressure gradient. The Sverdrup balance was validated by a number of observational studies in the Pacific ocean [Halpern, 1980; Bryden and Brady, 1985; McPhaden and Taft, 1988; Yu and McPhaden, 1999b]. On an annual mean basis a good agreement of the Sverdrup balance was found by [McPhaden and Taft, 1988] , while on shorter than annual time scales the Sverdrup balance does not hold [Hebert et al., 1991]. [Yu and McPhaden, 1999b], demonstrated that on seasonal time scales the differences between ZPG and wind stress drive the acceleration and deceleration of the vertically integrated zonal currents, while on an interannual time scale the local acceleration/deceleration is very small, compared to the ZPG or the wind stress. The advantage of using the depth integrated ZMB, especially in observations, is that it avoids having to estimate the full vertical profile of the ZDF term, which is very challenging due to uncertainties in the magnitude of the vertical eddy viscosity \(k_{m,v} \). Essentially, the validity of the Sverdrup balance, either in a time mean sense
[McPhaden and Taft, 1988], or on a seasonal time scale accounting for local acceleration [Yu and McPhaden, 1999b], provides an estimate of the possible size of the non-linear (ADV) and lateral diffusion (LDF) terms in the vertically-integrated ZMB. However, the vertically-integrated ZMB cannot provide any information on the specific balance of terms at any depth, and is therefore of limited use itself in understanding the seasonal variations in intensity and vertical structure of the EUC.

Existing observational studies of the depth-dependent ZMB have generally been limited in their ability to calculate all of the terms in Eq. 3-1, and have typically inferred the ZDF profile as a residual of the remaining resolvable terms. The most comprehensive available studies are from the Pacific, in particular those by [Bryden and Brady, 1985, 1989; Qiao and Weisberg, 1997; Yu and McPhaden, 1999b].

Examining the role of the ZPG in modeling studies does not necessarily call for the use of the Sverdrup balance, since the all the terms of the ZMB can be explicitly calculated. Using a model, [Waongne, 1988, 1989] found that for a narrow strip of longitudes in the western part of the equatorial Atlantic basin, the ZPG exceeds the ZDF so that the EUC accelerates eastward, while in the rest of the basin it decelerates. In a high resolution model simulation for the equatorial Atlantic, [Wainer et al., 1999] confirmed [Waongne, 1989] that near the western boundary the ZPG was the most important term for the EUC. [Brown et al., 2007] using an isopycnal model for the equatorial Pacific validated the Sverdrup balance on the equator, as it was seen repeatedly on observations.

The next important term in the ZMB is the ZDF. [Halpern, 1980] estimated ZDF as a function of depth assuming that it should balance a linearly decaying with depth ZPG. [Bryden and Brady, 1985; Johnson and Luther, 1994; Qiao and Weisberg, 1997] calculated
ZDF as a residual of the rest of the terms of the balance. [Bryden and Brady, 1985] showed a profile of the term that suggested turbulent (eddy) mixing at least to a depth of 200m below the EUC core, deeper than the 90m depths suggested by [Johnson and Luther, 1994]. [Qiao and Weisberg, 1997], in their Pacific ocean study, obtained a similar ZDF profile to the one from the modeling study of [Wacongne, 1989] and [Bryden and Brady, 1985]. The directly estimated ZDF term from microstructure measurements [Peters et al., 1988] was investigated by [Dillon et al., 1989] and [Hebert et al., 1991], both in the equatorial Pacific and both of which report large differences between direct and residual ZDF estimates.

The interpretation of the ZDF term that is estimated in the models could be challenging, due to the sensitivity of the models to different eddy viscosity coefficients [Schott and Böning, 1991; Wainer et al., 1999]. [Wainer et al., 1999] used two cases of $k_{m,v}$ (high and low) and showed different dynamical regimes in the ZMB for different parts of the basin. Models provide a great test bed for investigating the sensitivity of $k_{m,v}$ in the ZMB, however without appropriate observational studies to validate the physical relevance of the ZDF for different $k_{m,v}$, the results would be inconclusive.

An imbalance between the ZPG and ZDF indirectly suggests that non-linear advection of momentum should be considered. Investigating the role of non-linear terms in the ZMB, [Bryden and Brady, 1985] reported that the sum of zonal and vertical advection profiles is responsible for a deceleration at all depths below 300m and that 20% of the surface wind stress is balanced by the depth integrated sum of zonal and vertical advection. In [McPhaden and Taft, 1988] the non-linear advection was found to play a role in increasing the eastward transport on the equator. [Qiao and Weisberg, 1997] used two different formulations to estimate the non-linear advection (flux divergence and advection
formulation) in both of which the terms were separated in a mean and eddy part. In [Qiao and Weisberg, 1997], the total (mean plus eddy) of both non-linear advection formulations show very similar results within their errors: the ADV accelerates the EUC at the core and decelerates the EUC below the core. In [Yu and McPhaden, 1999b], the importance of the non-linear advection terms in the Pacific EUC was found significant on all time scales but not significant when vertically integrated. This weak contribution of the non-linear advection terms to the vertically integrated momentum balance is common in both [Yu and McPhaden, 1999b] and [Qiao and Weisberg, 1997] studies. The assumption of an equatorially symmetric meridional structure of the EUC has probably led most observational studies to neglect the meridional advection term [Bryden and Brady, 1985; Dillon et al., 1989; Hebert et al., 1991].

The role of non-linear advection terms in the momentum balance has been evaluated within model studies [Wacongne, 1989; Wainer et al., 1999; Harrison et al., 2001; Brown et al., 2007; Nagura and McPhaden, 2014]. In [Wacongne, 1989]'s equatorial Atlantic study, different contributions of the advection terms were found above and below the thermocline: above the thermocline the seasonally varying momentum balance was between the sum of meridional and vertical advection and the sum of ZPG and ZDF, while in and below the thermocline the local acceleration was balanced by the sum of zonal and vertical advection, ZPG and LDF. In the annual mean, the advection terms in the modeling studies of [Wacongne, 1989] in the Atlantic and [Harrison et al., 2001] in the Pacific showed that above the core of the EUC, the non-linear advection terms tend to accelerate the EUC, while below the core of the EUC the non-linear advection terms tend to decelerate the EUC. [Brown et al., 2007] confirmed in their modeling study of the equatorial Pacific
the important contribution of non-linearity to the balance, and by further separating the non-linear advection term into a mean and an eddy component (deviations from the mean), showed the contribution of TIWs. Finally, although the equatorial Indian is dominated by the monsoonal wind-driven circulation and dynamically resembles its Pacific and Atlantic counterparts for only half of the year, [Nagura and McPhaden, 2014] showed in a ZMB study the importance of non-linear advection at all time scales.

Finally, the LDF term in the observational momentum balance studies is usually assumed small and neglected (i.e. [McPhaden and Taft, 1988; Hebert et al., 1991]). [Bryden and Brady, 1989] however pointed out the importance of the LDF term using observations and also suggested that if the $k_{w,H}$ in the modeling study of [Philander and Pacanowski, 1986a] was reduced it would provide more realistic results. Also, [Wacongne, 1989] in her modelling study showed that below about 200m the balance between ZPG and LDF controls the seasonal local accelerations.

Thus far, there are two very significant gaps in the ZMB studies on the equator: (i) there is no observational ZMB study in equatorial Atlantic (to the author’s knowledge) other than the vertically integrated ZMB studies (i.e. [Halpern, 1980; Bryden and Brady, 1985; McPhaden and Taft, 1988; Yu and McPhaden, 1999b]) and (ii) there have been few attempts to study the ZMB on a seasonal time scale. [Johns et al., 2014] discussed the importance of the ZMB, however in a qualitative approach without including any estimate of the ZMB term. This present work gives a first ever estimate of the seasonal momentum balance of the equatorial Atlantic using observations, aiming to fill the gaps in previous ZMB studies. Additionally, the output from a model simulation of the equatorial Atlantic is used to test the methods and assumptions used in the observational analysis, while also
investigating, in parallel, the momentum balance within the model’s framework. It is important to note that the model simulation used in this study realistically captures the seasonal cycle of the EUC, making it a useful tool for such a comparison study.

3.2 Data

3.2.1 Moored observations

The moorings deployed during the Tropical Atlantic Climate Experiment (TACE, 2007-2011) at the equator, 0.75ºN, and 0.75ºS (Figures 2-1) at 3 longitudes (23ºW, 10ºW and 0ºE) were equipped with Acoustic Current Doppler Profilers (ADCPs), providing the

![Figure 3-1: Zonal velocity time series at all the TACE mooring locations (in cm/s). The columns represent different longitudes (from left to right 23ºW, 10ºW and 0ºE respectively) while each row shows different latitudes (from top to bottom 0.75ºN, 0ºN and 0.75ºS respectively). The red shading indicates eastward velocities (like in the EUC) while the blue indicates westward velocities. The colorbar limits are ±100 cm/s highlighting the strong EUC.](image-url)
current time series within the EUC used in this study. A more detailed description of the instrumentation and the deployments can be found in [Brandt et al., 2014; Johns et al., 2014] as well in Chapter 2 (section 2.2.1). Figure 3-1 shows the time series of 40-hour lowpass filtered zonal (u) current velocities from the ADCPs during TACE. The EUC is strongest at all longitudes on the equator and tends to be slightly asymmetric with stronger flow at 0.75°S than at 0.75°N. At both 23°W and 10°W, a summertime deep extension of the EUC is evident, which can be seen most clearly at the equator and 0.75°S. Figure 3-2 shows the meridional (v) current velocities from the ADCPs, where the fluctuations occur on generally faster (non-seasonal) time scales and include the effects of TIWs during boreal

![Figure 3-2: Meridional velocity time series at all the TACE mooring locations (in cm/s), same as Figure 3-1. The red shading indicates northward velocities while the blue shading indicates southward velocities. Note the different from the zonal velocities colorbar limits (±40 cm/s).](image-url)
summer and fall. Note that the zonal currents are almost 3 times larger than the meridional currents, however the temporal variability of the latter is more pronounced. Both zonal and meridional velocity time series are used for the calculation of the advection terms in the momentum balance (Eq. 3-1).

Higher temporal resolution (1hr) moored currents are also used to estimate vertical shear and obtain a direct estimate of the vertical dissipation of momentum (turbulent vertical stress term). For the vertical stress term, the vertical eddy viscosity coefficient ($k_{m,v}$, hereafter $k_m$) is estimated using gradient Richardson number ($Ri$) based parameterizations. By definition, $Ri$ includes the calculation of vertical shear and stratification, both of which are estimated here at the highest possible temporal resolution (hourly). Details on the calculation of shear and $Ri$ are given in the “Methods” section of

Figure 3-3: Time series of temperature, salinity and density profiles at 10°W to a depth of 200m using the merged PIRATA-SIO Argo T/S profiles.
this chapter (3.4).

Monthly and hourly temperature and salinity data from the PIRATA equatorial moorings (http://www.pmel.noaa.gov/tao/) are obtained to calculate both density and stratification frequency. The PIRATA T and S profile time series are combined with a monthly SIO Argo analysis for the equatorial Atlantic region [Roemmich and Gilson, 2009] to determine the most accurate T, S, and density profiles at the mooring locations, as described in [Johns et al., 2014]. In Figure 3-3, temperature, salinity and density data at 10°W show clearly the evolution of the cold tongue (high salinity, low temperature) associated with the summer upwelling. This monthly T/S profile analysis is used to establish a functional dependence between temperature and density that can then be used in the calculation of high resolution (1hr) density profiles from PIRATA temperature profile data. This step is necessary since the hourly PIRATA salinity data have a very coarse vertical resolution with 5 depths at most and with numerous gaps in the records.

3.2.2 Gridded Argo

As mentioned in the previous section, the SIO-Argo T/S profiles are used on the equator at the mooring locations but the full gridded four-dimensional (longitude, latitude, depth, month) Argo data from 40°W to 10°E and 5°N to 5°S is used to compute the densities and the dynamic heights in the greater EUC area. Recall that Section 2.2.2 describes the detailed methodology and tests for the choice of reference level by also combining altimetry products provided by AVISO (Archiving, Validation and Interpretation of Satellite Oceanographic data, (http://www.aviso.altimetry.fr/duacs/). While in Chapter 2 the dynamic topography is used in the calculation of geostrophic currents and ZPG, in this chapter the dynamic topography is used only for the calculation of the ZPG (section 3.4).
3.2.3 Surface winds

The Cross Calibrated Multi-Platform Product (CCMP) wind stress [Atlas et al., 2011] is used to provide surface wind stress estimates. The horizontal resolution of the CCMP winds is 0.25°x0.25° and the temporal resolution is 6-hourly. The zonal and meridional winds at 10m from January 2007 to December 2011 cover the TACE period and are used to investigate the relation between seasonality of the winds and the seasonality of the forcing terms in the momentum balance. Figure 3-4 presents the seasonal cycle of the zonal wind stress and highlights the reversal of the southeasterly trade winds in the eastern part.
of the equatorial Atlantic basin in the Gulf of Guinea. The lower panel focuses only on 10°W and reveals significant seasonal and interannual variability of the zonal wind stress with short lived reversals to eastward wind. However, the eastward reversals are not strong enough to show as a reversal of the local winds in the seasonal cycle at 10°W (upper panel of Figure 3-4).

The CCMP surface winds are also used as a boundary condition in the direct estimate of the vertical stress term in the observational momentum balance, following Eq. 3-3. By using the surface zonal wind stress as a boundary condition in the ZDF, the wind induced forcing can be considered directly in the balance yielding a more accurate estimate of this term. Furthermore, this surface boundary condition is the same one used in the model for the ZDF term and it allows for a more direct comparison of the term between the model and the observations.

3.3 Model

Consistent with the previous chapter (section 2.3), the Nucleus for European Modelling of the Ocean – Océan PAralléléisé (NEMO-OPA) model output is utilized in this chapter. The model has a ¼ degree horizontal resolution, 75 vertical levels and the output used here is for the time period between 2008 and 2012, covering most of the TACE time period. For further details on the particular model simulation the reader is referred to [Jouanno et al., 2011a, 2011b, 2013]. The momentum balance terms in the model are compared to those derived from the observations to test the validity of different methods and assumptions and to complement the seasonal momentum balance analysis. Daily outputs of ZMB terms are computed online during the model integration. However, the fact that the primitive zonal momentum equation in the model is in flux form [Madec, 2008], implies that terms such
as the meridional and vertical advection of the zonal momentum are not explicitly calculated at each time step from the prognostic equation. Instead, the model’s $u$, $v$ and $w$ fields are used to estimate those terms in the balance. This has no significant implication in the closure of the momentum balance of the model, since the tests performed prior to any further analysis (not shown) show a closed balance. From the same simulation and at the equatorial mooring locations, hourly velocities, temperature, salinity output are also used.

3.4 Methods of estimating the observational zonal momentum balance terms

An estimate of the seasonal ZMB in the Atlantic EUC is derived purely from observations in this section. The section also describes in detail the methodology used, the use of equation 3-1, to derive the estimate. Due to data limitations, we are only able to estimate the seasonal ZMB at 10°W, the central longitude of the TACE observing array. Since this study is investigating the seasonality of the momentum balance, hereafter the term seasonal will be used to refer to the climatological seasonal cycle of each term or variable, estimated by considering all i.e. March data in the time series of each variable of interest.

3.4.1 Local acceleration and advection

3.4.1.1 Local acceleration

The term on the LHS of Eq. 3-1 ($\frac{\partial u}{\partial t}$) is the local acceleration term and is estimated using the equatorial mooring zonal velocity ($u$) time series shown in Figure 3-1. The
climatological seasonal cycle of $\frac{\partial u}{\partial t}$ is derived from the 4 years of mooring data available during TACE, after the 12-hour $u$ profile time series is averaged into monthly time series (Figure 3-5 (upper panels)). Similarly, for the model’s case the daily $u$ profile time series are also averaged into monthly before estimating the $\frac{\partial u}{\partial t}$ gradient. This monthly averaging of $u$, prior to estimating the local acceleration, aims to reduce any noise that can be induced in the seasonal cycle by using such a small time step (1 day for the model, $\frac{1}{2}$ a day for the moorings) in the calculation of the $\frac{\partial u}{\partial t}$ gradient (see Appendix C for details).

Figure 3-5: Upper panels show the local acceleration term at (a) $23^\circ$W, (b) $10^\circ$W and (c) $0^\circ$E while the bottom three panels the ZPG at the same locations. The green line is the location of the EUC core at each month, defined as the depth that the EUC is maximum.
3.4.1.2 Zonal Advection

The first term on the RHS of Eq. 3-1 is the zonal advection \( \frac{\partial u}{\partial x} \), which requires an estimate of the \( \frac{\partial u}{\partial x} \) gradient at the longitude of interest. The \( \frac{\partial u}{\partial x} \) gradient at 10°W is estimated by linear regression using the equatorial \( u \) velocity profiles at all 3 mooring locations (see section 2.5.3.1). However, due to the ~10° separation between the mooring longitudes, the term is potentially sensitive to biases. The 1/4°x1/4° horizontal resolution of the model is much higher than the resolution of the observations and allowed for the testing of this method. As it is further discussed in the results section (3.5.3.1), the spacing of 10° degrees of longitude leads to an observational zonal advection estimate that is more representative of a zonally averaged zonal advection, rather than a localized estimate of the term. The \( u \frac{\partial u}{\partial x} \) product is calculated at each time step (12 hours), and then the monthly climatological mean value for each month is estimated by averaging all of the individual values for each month.

3.4.1.3 Meridional Advection

Estimating the meridional advection term \( v \frac{\partial u}{\partial y} \) requires, besides the meridional velocity profile, an estimate of the \( \frac{\partial u}{\partial y} \) gradient on the equator. The two off-equatorial moorings at 10°W, one at 0.75°N and one at 0.75°S, allow for the \( \frac{\partial u}{\partial y} \) calculation using central differences. However, this method could also introduce biases to the estimate of
\frac{\partial u}{\partial y}$ on the equator. The EUC has a Gaussian-like meridional structure that peaks almost on the equator, suggesting that a local estimate of the \( \frac{\partial u}{\partial y} \) gradient on the equator could contain significant differences from one estimated between 0.75°S and 0.75°N. Again, using the model to test to sensitivity of the estimate of the meridional advection term, this work assesses the validity of this method and possible introduction of biases. In section 3.5.3.1, the meridional advection derived within the model’s framework but using the same meridional spacing of ±0.75° as in the observations is compared to a zonally averaged meridional advection estimate to reveal qualitative and quantitative agreement between the too. The localized at 10°W meridional advection from the model when compared to the observation estimate however shows only qualitative agreement, concluding that as for the zonal advection, the meridional advection derived from the observations is a better representation of a zonally averaged than local estimate of the term. More details on this comparison are provided in section 3.5.3.1. As for the \( \frac{u \partial u}{\partial x} \) term, the \( \frac{v \partial u}{\partial y} \) product is calculated at each 12-hour time step and averaged to obtain monthly climatological estimates.

### 3.4.1.4 Vertical Advection

The vertical advection \( \frac{\partial u}{\partial z} \) is a challenging term to estimate. Although the \( \frac{\partial u}{\partial z} \) gradient can be estimated at all mooring locations from direct \( u \) measurements, vertical velocities cannot be directly measured and can only be inferred using, for example, the continuity equation (see section 2.5.3.1). Our previous analysis reveals that a vertical velocity
estimate at 10°W, derived using the continuity equation locally, could lead to a reasonable annual mean estimate. However, each individual month in the seasonal cycle, especially during July, shows very high values of \( w \), not confirmed by either the mass balance inferred upwelling (derived through the conservation of volume transport through the faces of conceptual box described in 2.4.1), or the model’s \( w \) fields.
Figure 3-7 illustrates the clear difference in magnitude between the mass balance inferred $w$ (left panel) and $w$ derived from continuity (right panel). [Bryden and Brady, 1989] show a similar difference between an upwelling velocity profile on the equatorial Pacific derived from continuity equation and mooring currents that they call “measured vertical velocity” and an equivalent $w$ derived from a method similar to the mass balance upwelling in the previous chapter, that they call “steady, geostrophic vertical velocity”. The difference between the two annual mean $w$ profiles, calculated from the seasonal cycles shown in Figure 3-7, is $0.25-0.5 \cdot 10^{-5} \text{ms}^{-1}$, comparable to the differences of about $0.5 \cdot 10^{-5} \text{m.s}^{-1}$ [Bryden and Brady, 1989]. When the same comparison is done within the model’s framework (Figure 3-6 middle and lower panels), it is found that the continuity-derived and mass-balance-derived $w$ estimates also differ in their seasonal cycles, and that a much better agreement is found between model mass balance derived $w$ and the actual vertical velocity on the equator in the model. Based on this result, and since the estimates of $w$ derived from the seasonal mass balance are more reasonable, the mass balance $w$ estimates are used in calculate the seasonal climatology of the $w \frac{\partial u}{\partial z}$ term in the observational analysis. The accuracy of the derived $w \frac{\partial u}{\partial z}$ estimates is later tested, again in section 3.5.3.1, within the model’s framework and shows some qualitative agreement below 50m between model and observational estimate. Unlike the estimates of $u \frac{\partial u}{\partial x}$ and $v \frac{\partial u}{\partial y}$, the $w \frac{\partial u}{\partial z}$ term in the observational analysis is only estimated from monthly mean values, since the mass-balance derived $w$ estimates are available only as a seasonal
climatology.

3.4.2 Zonal pressure gradient

The ZPG on the equator is estimated from absolute dynamic topography fields, derived from Argo T/S profiles. To estimate the ZPG at each longitude, the slope of the dynamic height is estimated by performing a linear fit using the dynamic height data within 10° (5° to either side) of the longitude of interest. The resulting climatological seasonal cycle of the ZPG using 7 years of Argo data (2004-2011) is shown for each longitude in Figure 3-5 (lower panels). A positive value of the ZPG term $\left( - \frac{1}{\rho_o} \frac{\partial \rho}{\partial x} \right)$ acts to accelerate the EUC while a negative value acts to decelerate the EUC. As shown in Figure 3-5, there are considerable differences in the seasonal climatology of the ZPG at 23°W, 10°W, and 0°E. The ZPG changes from mostly positive at 23°W to mostly negative at 0°E, particularly within the upper 100m of the water column. The ZPG at 10°W is intermediate between these two regimes and shows mostly positive ZPG values except for a ZPG reversal below about 50 m during boreal fall. Comparing the upper (local accelerations) to the lower (ZPG) panels in Figure 3-5, changes in the ZPG alone cannot explain the seasonal strength of the EUC, illustrating the need to study the dynamics of EUC including all the terms in the momentum balance (non-linear advection, dissipation and zonal pressure gradient).
3.4.3 Lateral diffusion

The lateral diffusion term in Eq. 3-1 is assumed to be small and is neglected in the observational momentum balance. It is also found to be very small in the model, at least an order of magnitude smaller than the other terms. Figure 3-8 shows the lateral diffusion term in the model (LDF) where the values range between ±0.1·10⁻⁸ m/s², two orders of magnitude smaller than the ZPG or the $u_t$ terms shown in Figure 3-5. It should be noted that within the framework of our seasonal analysis, any contribution of eddies to the lateral diffusion of momentum is explicitly accounted for in the non-linear advective terms $\frac{\partial u}{\partial x}$ and $\frac{\partial u}{\partial y}$. This differs from most previous studies (i.e. [Bryden and Brady, 1989; Qiao and Weisberg, 1997]), in which the total flow is broken into a time mean ($\bar{u}$) and deviations from that time mean ($u'$), and the contributions to the lateral momentum diffusion due to eddies are then explicitly calculated by:

$$k_{m, H} \nabla_H \nabla_H \bar{u} = \nu \nabla_H \nabla_H \bar{u} - \left[ \frac{\partial (\bar{u}' u')}{\partial x} + \frac{\partial (\bar{u}' v')}{\partial y} \right] \text{ (Eq. 3-4),}$$

where $\nu$ is the molecular viscosity. In these formulations, the non-linear terms are then
given by \(-\frac{\partial u}{\partial z}, \frac{\partial u}{\partial y}\) and represent only the contributions to non-linear advection by the

time mean flow. In the calculations of this work, the non-linear terms are calculated at each
time step, therefore including contributions by the (monthly) mean flow as well as by any
resolved eddies (which here includes time scales of \(\geq 1\) day). The lateral diffusion term in
the observational analysis therefore contains, in principle, only the effects of molecular
viscosity, which is negligible in the momentum balance.

3.4.4 Turbulent vertical stress

The vertical stress term (ZDF), cannot be directly estimated using only the available
mooring data. By definition, the vertical stress is \(ZDF = \frac{\partial}{\partial z} \left( k_m \frac{\partial u}{\partial z} \right)\) (Eq. 3-5), with \(k_m\)
being the vertical eddy viscosity and \(\left( \frac{\partial u}{\partial z} \right)\) the vertical shear, which can be calculated at
the mooring locations. For a given the vertical stress term could be directly estimated, but
obtaining representative estimates of \(k_m\) is challenging.

Due to the lack of direct \(k_m\) measurements, the first estimate of vertical stress that is
evaluated for the momentum balance at 10ºW is the “residual vertical stress”, defined as
the difference between local acceleration and the sum of advection and ZPG, given by:

\[
ZDF(\text{residual}) = \frac{\partial u}{\partial t} + \left( u \frac{\partial u}{\partial x} + v \frac{\partial u}{\partial y} + w \frac{\partial u}{\partial z} \right) + \frac{1}{\rho_o} \frac{\partial p}{\partial x} \quad \text{(Eq. 3-6)}
\]

We also attempt to make a direct estimate of the vertical stress term using available
Richardson number (\(Ri\))-based parameterizations for the vertical eddy viscosity \(k_m\). The
gradient Richardson number is defined by \(Ri = \frac{N^2(\text{stratification})}{S^2(\text{shear})}\) (Eq. 3-7). In this study,
\( R_i \) is estimated using (3-7) both for the model and the observations and used to test a range of parameterizations (for a complete list see Appendix D). The shear, defined as

\[
S^2 = \left( \frac{\partial u}{\partial z} \right)^2 + \left( \frac{\partial v}{\partial z} \right)^2
\]

(Eq. 3-8) is estimated using the hourly mooring \( u \) and \( v \) ADCP data in the observations and the hourly \( u, v \) velocities in the models’ output. The stratification, defined as

\[
N^2 = -\frac{1}{\rho_0} \frac{\partial \rho}{\partial z}
\]

(Eq. 3-7), is calculated using density profiles (\( \rho \)) derived from the hourly PIRATA temperature (T) profiles using a cubic fit of the form:

\[
\rho = c_1 T^3 + c_2 T^2 + c_3 T + c_4
\]

(Eq. 3-8). The process of obtaining the coefficients of the fit is explained in detail in Appendix D. In the model’s case, where salinity is available, the hourly T/S profiles are used to provide density profiles and stratification. [Hummels, 2012] extensively tested different types of \( R_i \) parameterizations available in the literature (i.e. [Pacanowski and Philander, 1981; Peters et al., 1988; Large et al., 1994; Zaron and Moum, 2009]) to obtain eddy diffusivity of heat coefficients (\( k_h \)) that were then compared to microstructure derived \( k_h \) collected in the Atlantic Cold Tongue (ACT) region. This study follows closely the work of [Hummels, 2012] in testing the above mentioned parameterizations but within the model’s framework (model’s \( N^2, S^2 \) and \( R_i \)), while it additionally tests the parameterizations suggested by [Hummels, 2012] and [Kunze et al., 1990] (see Appendix E for details). In evaluating these parameterizations, the model is used to help assess which parameterizations produce realistic profiles of \( k_m \). In the model simulation used here, \( k_m \) is evaluated within the Generic Length Scale (GLS) vertical diffusion scheme [Umlauf and Burchard, 2003, 2005]. The GLS scheme uses a prognostic equation for the turbulent kinetic energy \( \bar{e} \) and another one for the generic length scale \( \psi \),
defined as $\psi = C_{0_{\mu}} \rho \bar{\varepsilon}^n l^n$ where $C_{0_{\mu}}$ is a constant dependent of the stability function and $l$ is the turbulence length scale. In the simulation used in this work, $(p, m, n)$ are set to the values given by the known k-$\varepsilon$ closure scheme [Rodi, 1987], whereas the stability function used is the Canuto-A [Canuto et al., 2001]. Details on the exact formulas of the GLS prognostic equations and the time stepping scheme used can be found in [Madec, 2008] and references therein. It is worth mentioning however that both GLS equations have a functional dependence on shear and stratification, but not $Ri$. Although the model's turbulence parameterization is not based on $Ri$ explicitly, the fact that the model reproduces the observed vertical structure and seasonal cycle of the EUC quite accurately suggests that the ZDF term in the model is also reasonably accurate, since it is a leading term in the overall ZMB.

3.5 Results

3.5.1 Observational momentum balance at 10ºW

The first observational momentum balance described in this section is the one using the residual vertical stress (Eq. 3-6), where the climatological (seasonal) cycles for each of the terms in the ZMB are shown in Figure 3-9. It is important to note that some of the terms in the balance (particularly the ADV term) contain significant errors due to both methodological biases and statistical uncertainty, and that these errors will accumulate in the residual ZDF term. As noted previously in Section 3.4.1, the model is used to estimate the uncertainties or biases introduced by the methodology as applied to the observations, and within these uncertainties, there is some qualitative agreement between model and observations.
Overall, the ZPG (Figure 3-9 (d)) appears as the most important forcing term of the balance. It is positive near the surface throughout the year and positive through the deeper water column from January to June, but shows a marked reversal at depth during fall, from approximately August to November. Interestingly, the maximum ZPG in July does not

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**Figure 3-9**: Observational momentum balance at 10°W at the equator. From top to bottom: (a) the seasonal zonal velocity, (b) the local acceleration where red shows acceleration and blue deceleration, (c) non-linear advection (ADV), (d) zonal pressure gradient (ZPG), (e) vertical stress (ZDF) inferred as a residual. Note that except for the ZPG, the rest of the terms are estimated using mooring currents and therefore the profiles are limited to depth below 50m. The units on the colorbars are ±3·10^-7 m/s² for (b) through (e) and 1 m/s for (a).
coincide with any of the maxima in strength of the EUC: instead, the maximum ZPG at 10W occurs during summer, when the EUC is at minimum strength. However, the local acceleration clearly shows a slightly offset seasonal cycle, with a deep extension of accelerating zonal current during summer that cannot be explained throughout its full duration by the ZPG alone. The ADV term (Figure 3-9 (c)) is generally weaker than the ZPG and of comparable magnitude to the local acceleration term. The residual ZDF (Figure 3-9(e)) is of similar magnitude to the ZPG and is the second most important term in the balance.

The month by month variations of the primary momentum balance between ZPG, ZDF and ADV are best investigated in terms of depth layers. Four depth ranges are used: (i) 0-50m (surface layer), (ii) 50-100m (upper EUC layer), (iii) 100-200m (lower EUC layer) and (iv) 200-300m (deep layer). Due to the limitations imposed by the mass-balance derived w profiles that start at 50m, this depth is chosen as the limit of the near-surface layer aiming to isolate this part of the flow where vertical advection, and therefore ADV (and consequently the residual ZDF) cannot be resolved. The 50-100m layer is chosen such that it always includes the EUC core, found at depths between 52m (March) and 71m (September) (Figure 3-9 (a)). The 100-200m layer includes the lower part of the EUC and within this layer the deep extensions during spring-summer can be examined thoroughly since all the terms of the balance can be resolved. In the 200-300m layer the meridional advection term can only be resolved between 200-230m, due to the shorter mooring record of equatorial meridional velocities (Figure 3-2) and therefore the ADV and residual ZDF can only be estimated between 200-230 m as well.
From 0-50m, where the ADV and residual ZDF are not estimated, the ZPG is almost in phase with the wind stress (Figure 3-4) and exhibits 2 maxima: one between June-August and one in January following the maxima of the westward wind stress in June and November respectively. The ZPG winter maximum is well collocated with the local acceleration maximum centered around 30-40m that occurs during December to March, when the EUC core is moving upward toward its seasonal minimum depth in April.

Throughout the rest of the year the surface layer shows a generally decelerating flow that reflects two processes: (a) the deepening of the EUC core toward its seasonal maximum in September-October, and (b) the actual weakening of the EUC core intensity, in July-August. The ZPG, although it weakens in late fall (October-November), is still very strong through the late spring and summer months and therefore cannot account for this deceleration - in fact it should tend to cause a maximum acceleration of the near-surface layer in June-July-August. Therefore, this deceleration must be caused by a combination of negative (decelerating) ZDF and/or non-linear advection that is sufficient to offset the positive ZPG forcing at this time. For ZDF, this should be reasonable since the westward surface wind stress is at a maximum around that time of the year (May-July), but the estimate of residual ZDF is not available at this depth range to confirm it. At 50 m, the shallowest depth where the ZDF and ADV terms can be estimated, the inferred ZDF shows a negative maximum in August, and the ADV term is generally negative from spring through fall (except for a weakly positive value in June).

In the 50-100m layer (the upper EUC layer), the similar magnitude of the ADV, ZPG and ZDF terms shows that they all have the potential to play an important role in the balance. The ZPG is mostly positive in this layer from January to August (except for a
weak reversal near 100 m during March-April), and then becomes strongly negative from September to December, with a peak negative value in October. The ADV term is generally negative in this layer from April to October and weakly positive from December to March.

At the EUC core depth (green line in all panels of Figure 3-9) there is a deceleration (\(\frac{\partial u}{\partial t} < 0\)) between April and June, leading into the summer minimum in EUC intensity, followed by a shorter accelerating period of almost two months (August-September). The ZPG at the EUC core depth is eastward and almost constant at a value of \(1 \times 10^{-7}\text{m/s}^2\) from December to August, when it reverses for the next three months, while the advection term at the EUC core stays negative all year. Although the eastward ZPG can explain the acceleration from December through early March at the EUC core, the EUC deceleration from April to June results from ADV and ZDF overwhelming the acceleration due to the eastward ZPG. The inferred ZDF is generally negative above the EUC core and positive below it, consistent with expectations, but it takes on positive values above the EUC core during fall (September to November). This is an unrealistic result and probably results from inaccuracies in the other terms during September to November. The brief “secondary” accelerating period of the EUC core in September occurs when the ZPG at core depth is almost zero and the ADV is negative, requiring a positive inferred ZDF to explain it. Further, the positive ZDF during October to December at the EUC core results from both negative ZPG and ADV when there is little net acceleration or deceleration of the EUC core. Another questionable feature in Figure 3-9 is the very large (negative) magnitude of the ADV term at \(~100\text{m}\) during May-June, in excess of \(-1 \times 10^{-7}\text{m/s}^2\), which results in a comparably large positive ZDF to balance it.
Below the EUC core, between 75-100m but still within the strong flow of the upper EUC regime, the deceleration that occurs in October to December is qualitatively consistent with the reversal of the ZPG that occurs at the same time. However, the magnitude of the deceleration is considerably smaller than the sum of the negative ZPG and ADV terms and again results in a large - and likely unrealistic - positive inferred ZDF in this depth range, as well as below it to depths of ~150 m. These large values of inferred (positive) ZDF below the EUC core in both spring and fall, of order $0.5-1 \times 10^{-7} \text{m/s}^2$, would require unrealistically large values of $k_m$ and are also inconsistent with the results of the model (to be presented later), which show typical ZDF values of $O(0.1 \times 10^{-7} \text{m/s}^2)$ below the EUC core.

In the 100-200m layer (lower EUC layer), the acceleration of the EUC from April to August results in the deep extension shown in Figure 3-9 (a). The maximum acceleration occurs between mid-May and mid-July (Figure 3-9 (b)) and the EUC responds with a deep extension maximum in July-August. There is a close correspondence between the acceleration and the ZPG term in this layer, with maxima in both occurring in June-July and minima in October. During the first part of the acceleration phase, the positive ZPG forcing appears to be partially offset by negative ADV, while during the transition to deceleration (in August) and extending until the time of maximum deceleration (in October), the ADV term acts to accelerate the flow. The overall deceleration in fall results in a gradual disappearing of the deep extension from October through early March and in a reversal of the flow to westward, therefore in the disappearance of the eastward EUC.

The inferred ZDF becomes mostly small below 100 m, except during fall (October-December), when it has relatively large positive values. Similar to the overlying layer, this
results from the fact that the ZPG magnitude is about twice that of the local acceleration, requiring a positive inferred ZDF to close the balance. Again, this feature is probably unrealistic, and it is suspected to be due to an overestimate of the ZPG term at these depths from the Argo data.

In the 200-300m layer (deep layer), the ZPG stays negative from August to November, driving a local deceleration which leads to the reversal of the zonal flow, similar to the lower EUC layer. The rest of the year (December to May) the ZPG is positive and so is the local acceleration, although weak. The ADV is weak all year between 200-230m, yet positive from July to October contributing to the deep extension and negative the rest of the year (November through June). The residual ZDF is weakly negative from November through July, and positive from August to November, mainly acting to balance the ZPG term since both the local acceleration and non-linear terms are small at these depths. This result is again suspicious, as the vertical stress is expected to be very small at depths well below the EUC where vertical shears are weak, and it is also inconsistent with the model results as shown later. It seems probable that the magnitude of the ZPG term, while qualitatively correct and consistent with the phasing of the local acceleration in the deep layers, is overestimated in the SIO Argo climatology. Within the uncertainty of the observations, the overall balance in the deep layers appears to reflect mainly a linear balance between local acceleration and the ZPG.

The main inferences that can be drawn from the observational analysis are:

1. In the surface layer (0-50m), above the EUC core, the ZPG forcing can explain the acceleration of the uppermost part of the EUC in winter (December-March), but
cannot account for its deceleration in late spring and summer, when the ZPG is strongest and the EUC is weakest.

(2) In the upper EUC layer (50-100m), all forcing terms appear to play an important role in the balance. At the core depth (59-83m) the EUC shows a semiannual cycle with acceleration in winter and late summer, and deceleration in late spring and fall. The deceleration of the EUC in fall appears to be mainly driven by the ZPG, which reverses at this time, while the deceleration in late spring appears to be a result of the positive ZPG being exceeded by the sum of (negative) ADV and ZDF.

(3) In the lower EUC layer (100-200m) and deep layer (200-300m), the acceleration of the zonal flow closely follows the seasonal cycle of the ZPG forcing, resulting in acceleration of the deep EUC in early summer (June-July) and deceleration in fall (Sept-Nov). This results in the "deep extension" of the EUC in mid-summer, which is replaced by westward flow from November to March at depths below ~150 m. Non-linear advection is weaker at these depths but appears to partially offset the deep EUC acceleration in early summer and sustain it in late summer.

3.5.2 Model momentum balance

The observations-only momentum balance shown in Figure 3-9 is a forced balance because of the calculation of ZDF as a residual. In this section, the model is used to get a better picture of the fully-closed momentum balance, including the ZDF term. In the model the zonal momentum balance terms are calculated exactly at 10ºW-0ºN, therefore providing a more localized estimate of the terms that include horizontal gradients, namely the ADV and ZPG. For the advection terms, the $\frac{\partial u}{\partial x}$ and $\frac{\partial v}{\partial y}$ gradients are estimated using
central differences, therefore the spatial scale of the gradients is 0.5° since the horizontal resolution of the model’s grid is $\frac{1}{4}° \times \frac{1}{4}°$. This 0.5° spatial scale for the horizontal gradients differs from the observations, where ~20° and 1.5° were used for the zonal and meridional spatial scales respectively. Similarly, the spatial scale of the ZPG is at 0.5° and it is calculated from pressure fields that are estimated using the hydrostatic equation and the free surface elevation. The model ZPG uses absolute pressure fields as opposed to dynamic height profiles referenced at a 500db level in the observations. However, this methodological difference should not yield significantly different results. The reason for this is that regardless of the referencing level (tested extensively in Chapter 2 and Appendix B2) the surface topography from referenced dynamic height is recovered to within a few cm (~1.6-1.7cm, Appendix B2). Therefore, the most important source of any differences in the comparison of ZPG between model and observations should be the spatial scale over which the gradient is calculated.

The seasonal cycles of the same terms of the momentum balance from the model are shown in Figure 3-10 and can be compared to the observational estimate (Figure 3-9). Note that the model-derived balance extends all the way from the surface to 300m, while the observations are limited to depths greater than 50m and shallower than 230m (ADV and residual ZDF).

Qualitatively, there is good agreement between the model momentum balance terms and the observations, apart from the ZDF term. The seasonal cycles of zonal velocities in observation and model (Figure 3-9 (a) and Figure 3-10(a)) are very similar, showing a
A semiannual cycle for the EUC that exhibits two maxima (spring and fall) in both. The deep EUC extension is also present in the model but starts a little earlier in March, lasts through September (approximately same duration) and appears slightly weaker. As a result, the local acceleration term is very similar between the observations and model. The ZPG also shows a very similar annual cycle (Figure 3-10(d)), with maxima in winter and mid-
summer as in the observations. The near-surface reversal of the ZPG seen in the observations in fall, however, is not a feature of the model; the model shows a weakening of the ZPG but no reversal. At greater depth, the model ZPG does show a reversal in fall, consistent with the observations, and the reversed ZPG during August-November is collocated with deceleration of the lower EUC layers. The non-linear ADV term (Figure 3-10(c)), shows a qualitative agreement with the observations, although certain features appear to be offset in either time or depth relative to the observations, and there are significant quantitative differences between the two. The ZDF in the model reveals a more persistent and simpler picture than in the observations over the depth range of 50-230m where they can be compared (Figure 3-10(e)). The ZDF is westward in approximately the top 100m, reaching its seasonal maximum near the surface and in the upper EUC during summer, and then reverses to weakly eastward with maximum values below the EUC core in fall and winter.

In the surface layer the model shows the westward flowing SEC extending down to approximately 30m all year long, resulting from the westward wind stress. Exactly at the surface Figure 3-11(b), upper) the order one balance is all year long between the ADV and the ZDF, with the ZPG contributing only about a third of the eastward momentum and the seasonality between ADV and ZDF showing a 180° phase. Although the ZPG and ADV are positive all year long in the 0-30m depth range, their sum is not large enough to overcome the ZDF (Figure 3-11, upper left). During summer, the weakening of the EUC can be attributed to the spring deceleration maximum in May (Figure 3-10(a), (b)). This spring through summer deceleration starts in March, where weaker eastward ZPG and ADV are not strong enough to compensate for the deceleration imposed by the ZDF (Figure
3-11(a)). Even when the ZPG starts to strengthen in May, the ZDF also becomes stronger and more westward sustaining the deceleration well through July when the ZPG is almost at its peak. Below 30m, eastward ADV weakens and reverses sign between July-October, contributing to a deceleration of the upper part of the eastward flow (~30m to core depth) at that time. In the surface layer, the model shows the ADV term changing sign with season and depth over the relatively narrow layer of 50m (Figure 3-11c, upper). In terms of magnitude, the ZDF clearly dominates the balance in the first 30m with the sum of ZPG and ADV compensating by contributing almost equal amounts of eastward momentum from 10-30m during spring and summer. It is remarkable that in the 10-30m depth range the ZPG and ADV have an almost mirror-like structure. The vertical structure of the ADV is a result of the meridional and vertical advection that play the most important role in the 0-50m depth range (Figure 3-11 (c), top row).

In the upper EUC layer (50-100m), the vertical migration of the EUC core with season has a clear seasonal cycle with a maximum core depth of 81m during July-August (Figure 3-11(a)). The deceleration at the core of the EUC occurs between March and August (Figure 3-11(b), middle panel): as the core of the EUC deepens it also weakens. During the time of the core deceleration (March to August), the ZPG remains strong at about \(1 \cdot 10^{-7} \text{ m/s}^2\) and peaks during August (\(1.2 \cdot 10^{-7} \text{ m/s}^2\)). During the ‘spring deceleration’ (March to May), the ZDF is responsible, since it remains slightly larger (\(>1 \cdot 10^{-7} \text{ m/s}^2\)) than the sum of the strong ZPG and weaker ADV. In May, the ZPG equals the westward ZDF (\(>10^{-7} \text{ m/s}^2\)) that has started to weaken, and in mid-May the ADV becomes negative and is mainly responsible for decelerating the flow. From mid-May to August, the ‘summer deceleration’ of the core is attributed to the sum of ADV+ZDF, where the westward ZDF continues to
become weaker and ADV becomes more westward. In July-August, the deceleration starts to weaken since the already strong ZPG slightly strengthens as it approaches its seasonal
maximum, while at the same time the westward ZDF continues to weaken and reaches a seasonal minimum, and the ADV remains strong (westward). At the core depth, the vertical advection is zero \( \frac{\partial u}{\partial z} = 0 \) and therefore the negative advection from May to September is a result of meridional advection exceeding the zonal advection, since as shown in Figure 3-11(c) the meridional advection is negative and the zonal advection is positive at that depth. This negative ADV contributes effectively to the deceleration of the EUC at core depth when the ZPG is at a maximum and the westward ZDF at a minimum. Similarly, the positive ADV from September to October contributes effectively to the acceleration of the EUC core when the ZPG is at a seasonal minimum and the westward ZDF strengthens towards its secondary maximum in November. This behavior highlights the importance of
the ADV at the core depth: if ADV did not play a significant role then the flow would decelerate in spring and fall, the times that the EUC is maximum.

In the lower layer (100-200m) the model shows much weaker magnitudes of all the terms in the balance (Figure 3-11(a) and (c) lower panels). The range of values of the terms here are less than \( \pm 1 \cdot 10^{-7} \text{m/s}^2 \), compared to approximately \( \pm 5 \cdot 10^{-7} \text{m/s}^2 \) and \( \pm 2 \cdot 10^{-7} \text{m/s}^2 \) for the surface and upper EUC layers, respectively. In the lower layer from 100-150 the local acceleration is mainly determined by ZPG and ADV (Figure 3-11), as the ZDF is weak at these depths (Figure 3-11(a), lower). At 150 m (Figure 3-11 (b), lower panel), the acceleration closely follows the ZPG annual cycle with a maximum in summer (July) and a minimum in fall (October). The ADV tends to oppose the ZPG, being negative in spring and summer and positive in late fall. The secondary acceleration maximum in December actually results from the positive ADV at that time. Below 150m and all the way to 300m, the ZPG is weaker and shows a continuous reversal layer between 150 and 200m from August through April (Figure 3-10(d) and Figure 3-11 (a), lower panel). The sum of ZDF and ADV below 150m is almost zero and only the ZPG plays a role in accelerating/decelerating the flow: i.e., the momentum balance between 150-300m is essentially linear. Below 200m, [Wacongne, 1989] also found a linear balance to hold on a seasonal time scale with the ZPG in phase with the local acceleration. Elements of deep layer linear balance as in [Wacongne, 1989] are also seen in the narrow 200-230m layer that is resolved in the observational momentum balance. The deep extension of the EUC from March to September is a result of the ZPG that is mainly positive and is slightly modified by a weak contribution of the ADV.
Summarizing the results of the zonal momentum balance derived from the model:

(1) In the surface layer, although the ZDF and ZPG dominate the balance and tend to compensate in that layer, the ADV plays a significant role in modulating the near surface flow.

(2) Negative ADV from May to August contributes effectively in decelerating the EUC at core depth when the ZPG is at a maximum and the westward ZDF at a minimum, while positive ADV from September to October contributes to the acceleration of the EUC core when the ZPG is at a seasonal minimum and the westward ZDF tends to be maximum.

(3) In the lower layer (100-150m), the local acceleration is balanced by the sum of ZPG and ADV and has a semiannual cycle with a summer maximum set by the ZPG and a winter maximum set by the ADV. From 150m to 300m, the momentum balance is linear with the ZPG being in phase with the local acceleration.

(4) The ZPG is primarily responsible for the existence of the deep extensions of the EUC from March to September.

3.5.3 Error analysis

3.5.3.1 Methodological errors: non-linear advection

The observational momentum balance and the model momentum balance both show the importance of non-linear advection between the surface and 300m, also shown in a number of previous studies [Bryden and Brady, 1985; McPhaden and Taft, 1988; Wacongne, 1989; Hebert et al., 1991; Qiao and Weisberg, 1997; Yu and McPhaden, 1999b; Brown et al., 2007; Nagura and McPhaden, 2014]. An obvious question in the observational momentum balance derived here is how realistic is the representation of the
ADV term given the data limitations. As discussed in the methods section, the zonal and meridional advection terms are estimated between mooring locations that are almost 10° of longitude and 1° of latitude apart, raising a question of representativeness of those terms relative to a more localized estimate. Furthermore, the vertical advection term uses a $w$ profile that is estimated through mass divergence, also a non-local estimate. Consequently, the ADV term can potentially introduce uncertainties that would impact the residual ZDF estimate.

Figure 3-12 shows a comparison of the individual advection terms from the observations and from the model, and their sum (the total ADV). There are qualitative similarities in all the individual terms, however the zonal and meridional advection in the

Figure 3-12: Advection of zonal momentum using observations (left) and using the model (right). From top to bottom: the zonal advection ($u_1u_1$), the meridional advection ($v_1v_1$), the vertical advection ($w_1w_1$) and the bottom panel is the sum of the upper three, the total advection of zonal momentum, same as in Figure 3-9(d) (left, observations) and Figure 3-10(d) (right, model). Note that the terms are plotted with the sign they appear on Eq. 3-1, in order to show the contribution to accelerating/decelerating the flow. The green line in all panels is the EUC core depth.
model (Figure 3-12, right) show larger magnitudes than in the observations (Figure 3-12, left). The zonal advection in both model and observations is positive below the core showing the downstream weakening of the EUC \( \frac{\partial u}{\partial x} < 0 \), although the model shows a more pronounced seasonality than the observations. The meridional advection shows relatively large negative values above the core in both cases (model and observations) and weaker negative values below the core for most of the year. The vertical advection is similar in magnitude below the core in both cases and, as expected, follows the core depth seasonality, as the \( \frac{\partial u}{\partial z} \) gradient is zero at the EUC core depth. In both the observational estimates and in the model the zonal and vertical advection terms tend to partially cancel each other, being of opposite sign above and below the EUC core. The large negative values of meridional advection above the EUC core also tend to compensate for the large positive values of the vertical advection term there. As a result, the sum of the advection terms (the total ADV, bottom panels of Figure 3-12) tends to be weaker in magnitude than the maximum values of any of the individual terms, which is especially evident in the model.

As noted in the previous sections, the total ADV in the model tends to accelerate the upper part of the EUC from winter through late spring (November to May) and then decelerate it from July to September. Below the EUC core, to depths of ~150 m, it mostly decelerates the EUC throughout the year. In the observations, the total ADV can only be resolved in the lower part of the EUC and it shows also dominantly negative values there, although its maximum occurs somewhat earlier in the year (May-June) than in the model (July-Sept). Also, weak positive values of total ADV occur there in the observations during
winter (Dec-Feb), which are not present in the model. Below 150 m, the seasonality of the total ADV in the observations is mainly dominated by the meridional advection, while in the model it is controlled more by the zonal advection and to a lesser degree by the meridional advection.

Using the model’s velocities at the same locations (latitudes, longitudes) of the moorings, the same methods (presented in section 3.4.1) used to estimate the advection...
terms in the observations are tested within the model’s framework. The difference between the model’s advection terms and the ones calculated using the model’s fields with the observational methods gives a quantitative estimate of the methodological error in the observational momentum balance.

For all advection components, the diagnosed methodological error (Figure 3-13) does not exceed $\pm 2 \cdot 10^{-7}$ m/s$^2$. The seasonal cycles of the methodological error in the zonal and meridional advection are very similar to the actual model’s zonal and meridional advection (Figure 3-13(a), (b)) but weaker, suggesting that the observational methods are underestimating these terms. The errors in the vertical advection term are generally smaller, but the estimates derived using the mass balance $w$ also tend to underestimate the magnitude of the (negative) vertical advection below the EUC core. Above the EUC core,
where the vertical advection term is strongly positive, these tests suggest that the observational methods would actually tend to overestimate this term (i.e., negative values in Figure 3-13c from ~30-50 m). The total methodological bias in the ADV term, summing the contributions from all terms (Figure 3-13d), suggests that the observational methods lead to a negative bias above the EUC core, in depths from ~20-50 m, and a positive bias at and below the EUC core, in depths from ~50-150 m, as well as near the surface.

The methodological error derived within the model’s framework shows that the methods used in the observations introduce uncertainties as large as the range of values of the terms themselves. The systematic underestimate of the zonal and meridional terms using the observational method therefore means that \( u \frac{\partial u}{\partial x} \) and \( v \frac{\partial u}{\partial y} \) at 10°W (calculated over zonal and meridional spacing of approximately 20° and 1.5° respectively) are systematically underestimated relative to the same terms calculated locally at 10°W and at the model’s native grid resolution (0.5° in both latitude and longitude). The differences in \( w \) are due to differences in the mass balance derived \( w \) versus the actual model \( w \) at the equator. This clearly indicates that the observational methods cannot provide an accurate assessment of the localized momentum balance at 10°W, but instead may provide a more representative measure of the zonally-averaged momentum balance over a range of longitudes centered on 10°W. To examine this further we show in Figure 3-14 (left) the zonally-averaged non-linear terms from the model over a +/- 5° span centered at 10°W (5°W-15°W), compared to the model terms derived using the observational methods, as above. The zonally averaged zonal advection term (Figure 3-14, right), is similar to not only to the observational method’s equivalent (Figure 3-14, left) but also to the observations
themselves (Figure 3-12(a)). Quantitatively, the zonally-averaged zonal advection is very similar to the observational method’s estimate for the 0-150m depth range; below 150m to 300m the zonally averaged zonal advection is slightly more positive. A very good qualitative and quantitative agreement is found between the vertical advection terms in Figure 3-14 from the surface to core depth (0 contour). However, the zonally averaged estimate of the vertical advection below 200m appears less pronounced than in the observational methods estimate (Figure 3-14, left). The zonally averaged meridional advection estimate also provides a good qualitative comparison for the first ~80m; below that depth the seasonality of the two estimates is very different. Overall the comparison between the zonally-averaged non-linear terms and the non-linear terms derived using the observational methods, gives a more reasonable comparison between model and observations advection terms, at least for the upper part of the EUC. This confirms partially that the observational methods are not representative of a localized momentum balance at 10ºW, but rather of a zonally averaged momentum balance for the depth range from the surface to 100m.

3.5.3.2 Measurement and statistical error

In addition to the above methodological errors, there are statistical uncertainties in the seasonal cycle of each of the climatological ZMB terms due to interannual variability within each month, over the period that the seasonal climatologies are constructed. This applies to both the observations and the model. These are quantified in terms of standard errors (ste) for each month, using the standard deviation of each set of monthly mean values and dividing by $\sqrt{N}$, where $N$ is number of independent months in each climatological monthly estimate. For the observations, these statistical errors should also include any
random measurement errors in the terms (e.g. for the ZPG term due to uncertainties in the Argo-derived dynamic height). A simple “noise-to-signal” (NRS) type of metric is shown on Table 3-1 and is used to compare the amplitude of the seasonal cycle to the maximum standard error (ste) of the seasonal cycle of each term. Values of \( NSR \geq 1 \) indicate that ste is as large (or bigger) than the amplitude of the seasonal cycle signal and therefore, the seasonal cycle of the term is not accurately resolved. Given the complicated nature of the estimation of the observational momentum balance terms, any \( NSR < 0.5 \) values are considered to be resolving the seasonality of the term.

Table 3-1: Range of values of the momentum balance terms and maximum error at 10ºW in the observations and model.

<table>
<thead>
<tr>
<th>Metric</th>
<th>OBS</th>
<th>MODEL</th>
</tr>
</thead>
<tbody>
<tr>
<td>( \frac{\partial u}{\partial t} )</td>
<td>[-1.5, 0.8]</td>
<td>0.7</td>
</tr>
<tr>
<td>( ADV )</td>
<td>[-1.2, 0.8]</td>
<td>1.1</td>
</tr>
<tr>
<td>( ZPG )</td>
<td>[-1.2, 2.7]</td>
<td>0.5</td>
</tr>
<tr>
<td>( ZDF )</td>
<td>[-2.1, 1.1]*</td>
<td>1.2</td>
</tr>
</tbody>
</table>

The local acceleration standard error (ste) (Figure 3-15(b)), has a maximum value of \( \pm 0.7 \cdot 10^{-7} \text{m/s}^2 \) near the surface in the observations and a lower maximum value of \( \pm 0.2 \cdot 10^{-7} \text{m/s}^2 \), again observed near the surface (Table 3-1). The near surface ste structure and seasonality of the local acceleration is similar to the ste seasonal cycle of \( u \) (Figure 3-15(a)) as expected. The \( NSR \) for the local acceleration is 3 times larger in the observations (0.3) than in the model (0.1) (Table 3-1), indicating than even when using monthly averaged \( u \)
velocity profiles for the estimate of $\frac{\partial u}{\partial t}$, the observations are noisier than the model.

The standard error (ste) seasonal cycle of the observational advection (Figure 3-15(c), left) shows maximum values up to $\pm 1.1 \cdot 10^{-7} \text{m/s}^2$ near 50m in August (Table 3-1), almost equal to the size of the model’s maximum ste of $\pm 1.2 \cdot 10^{-7} \text{m/s}^2$ (Table 3-1), occurring between March to June and very prominent at shallower depths (30m) (Figure 3-15(c)), right). Below 150m the observational ADV ste and the model equivalent become almost same in magnitude ($< 0.25 \cdot 10^{-7} \text{m/s}^2$), with an exception in August-September at 150m in the observations ($\sim 0.75 \cdot 10^{-7} \text{m/s}^2$). The $\text{NSR}=0.6$ for observational ADV, is not only 3 times larger compared to the $\text{NSR}=0.2$ value of the model’s ADV, but also larger than the

![Error Obs vs Model](image-url)

Figure 3-15: Standard error seasonal cycles in Figure 3-9 in the left (observational momentum balance derived using residual stress) and standard error seasonal cycles in Figure 3-10 in the right (model momentum balance). (a) zonal current $u$, (b) local acceleration $u_t$, (c) ADV (advection), (d) ZPG (zonal pressure gradient) and (e) ZDF (vertical diffusion of zonal momentum, vertical stress)).
NSR < 0.5 value that is used as a threshold. This elevated NSR value for the observational ADV is the largest one among all the terms (model and observations combined) and further highlights challenge in accurately representing this term in the momentum balance.

The ZPG shows the smallest ste and the smallest ste seasonal variability of all the ZMB terms in the observations (Figure 3-15(d) left) and smallest maximum error (±0.5·10⁻⁷ m/s², Table 3-1). However, in the model the ZPG ste (Figure 3-15(d), right) shows larger values with a maximum ste of ±1.2·10⁻⁷ m/s², same as the ADV maximum ste in the model. Interestingly, the NSR of the ZPG is in the model is 3 times larger than the observations, this time highlighting the fact that the observational seasonal ste ZPG estimate results from monthly time series as opposed to the model seasonal ste ZPG estimate that is a result of daily time series.

The ZDF error (residual) in the observations is the largest among the terms of the momentum balance (±1.2·10⁻⁷ m/s², Table 3-1). The NSR < 0.4 for the observational residual ZDF, is larger than the model’s ZDF and has a very different and much smaller range of seasonal values.

The contribution of measurement errors to the observational ZPG estimate can also be assessed. The SIO Argo T/S is the product of Optimal Interpolation (OI) [Roemmich and Gilson, 2009] and the Argo float coverage within ±1° of the equator during the TACE period was approximately one float per 2° of longitude, between 25°W and the Gulf of Guinea [Johns et al., 2014]. In order to assess quantitatively the measurement error in the Argo dynamic height profiles, the differences between the dynamic heights from the merged PIRATA-Argo T/S profiles (section 3.2.2) and the dynamic heights derived from the Argo T/S profiles are considered. The differences between them (Figure 3-16) provides
a measure of the accuracy of the ZPG term derived purely from the SIO Argo analysis (Appendix B2). As shown in Figure 3-16, the differences δDH in the dynamic height are largest close to the surface and grow smaller with depth, showing near surface maximum value for 10°W of 1.2cm. Assuming errors of this magnitude distributed randomly over 10° of longitude, the expected uncertainty in the ZPG term at the surface for any individual month would be approximately $0.8 \times 10^{-7} \text{ m/s}^2$, based on a Monte-Carlo simulation with 1000 iterations. The corresponding measurement error in the climatological monthly estimates using 7 years of Argo data would then be about 2.5 times smaller (divided by $\sqrt{7}$), or $\sim 0.3 \times 10^{-7} \text{ m/s}^2$, and become smaller with depth. This is very similar to the statistical ste for the ZPG term shown in Figure 3-15, and suggests that most of the uncertainty in the climatological ZPG term can be explained by measurement error.

Figure 3-16: Mean dynamic height error profile. Top panel shows the DH differences (referenced at 1000 db), between the DH derived from the merged PIRATA-Argo T/S profiles and the Argo only T/S profile at the three mooring locations on the equator. The bottom panels show the mean error profiles of temperature (left) and salinity (right). Solid lines depict the mean profiles, while dashed lines are the standard deviations (blue for 23°W, red for 10°W and green for 0°E).
3.5.4 Direct estimate of vertical stress in the observations

The main goal of the section is to attempt and estimate a direct ZDF term (Eq. 3-5), one that is not inferred as a residual (Eq. 3-6). If ADV is assumed to be accurately represented in the momentum balance derived from the observations, ignoring any errors that have been discussed thus far, the residual ZDF would be a reasonable estimate of the ZDF term. The strong eastward momentum induced by the residual ZDF during September-October, contributes more than the ZPG in accelerating the EUC in the upper layer. Qualitatively, this feature is questionable: although the local westward wind stress is at minimum during September allowing for the ZPG to relax considerably, there is no evidence of wind stress reversal that is necessary to induce such a pronounced eastward momentum. Additionally, the error analysis of the observation-based momentum balance presented in the previous section (3.5.3) shows that the magnitude of the ADV errors (ste) can be as large as the magnitude of the ADV term itself, indicating that the uncertainties that are included in the residual ZDF can be large and, therefore, the latter is not accurately represented at all times.

Calculating a direct estimate of the ZDF terms from observations is a challenging task. Prior to apply any parameterizations to obtain $k_m$ profiles as described in section 3.4.4, the seasonal cycles of shear, stratification and $Ri$ number from the model are compared to those estimated from observations (Figure 3-17). There is a good qualitative agreement between the seasonality of the model’s and observations’ $N^2$ (Figure 3-17(a), (b)) and $S^2$ (Figure 3-17 (c)-(d)). In the observations, smaller $N^2$ values ($<10^{-4}s^{-2}$) occur at depths greater than about 100m relative to the model, however the larger $N^2$ values ($>10^{-3.5}s^{-2}$) appear at similar depths in both cases (Figure 3-17(a), (b)). The $S^2$ values in the observations and in the
model show pronounced and similar seasonality in the surface to 50m depth range. Below 50m in the observations (Figure 3-17(c)), \( S^2 \) falls below \( 10^{-4}\text{s}^{-2} \) at \(~100\text{m}\) from October to May, while between June and September, shear values \(< 10^{-4}\text{s}^{-2}\) are observed at shallower depths, indicative of reduced shear during the weaker EUC period. The deepening of the higher shear values during September-October follows the EUC core that reaches maximum depth during September (black dashed line in Figure 3-17 (e)). In the model, the core depth (black dashed line in Figure 3-17(f)) is associated with smaller shear than in the observations \(< 10^{-4}\text{s}^{-2}\) and there is not a clear deepening of high shear values during July when the core reaches its maximum depth. Just above 100m, the model’s shear exhibits a semiannual cycle with two maxima (one in February and one in late October), associated

![Figure 3-17](image)

Figure 3-17: (a) Seasonal stratification \( N^2 \) from observations, (b) seasonal stratification \( N^2 \) from the model, (c) seasonal shear \( S^2 \) from observations, (d) seasonal shear \( S^2 \) from the model, (e) seasonal \( Ri \) from observations, (f) seasonal \( Ri \) from the model, (g) percentage of \( Ri < 0.5 \) per month for each depth from observations, (h) percentage of \( Ri < 0.5 \) per month for each depth from the model. From (a)-(f) the \( N^2, S^2 \) and \( Ri \) are in logarithmic scale. The dash black line in (e) and (f) is the EUC core depth.
with the EUC maxima at approximately the same time (Figure 3-10(a)). [Hummels et al., 2013] used the data from shipboard ADCPs and CTDs to estimate seasonal cycles for the shear and stratification at 10°W between 2°N and 2°S that are very comparable to this analysis (the reader is referred to in Fig. 7(a) and (d) in [Hummels et al., 2013]). In [Hummels et al., 2013] both shear and stratification show a very comparable seasonality to the ones in Figure 3-17(a) and (c) for the first 100m: the deepening of the higher shear and stratification during boreal fall (September-October) is evident in both estimates (this work and [Hummels et al., 2013]).

The estimates of $Ri$ from the observations and model are presented in Figure 3-17 in two ways: (a) the simple monthly mean climatology of $Ri$ (Figure 3-17e and f), and (b) the percentage of hourly $Ri$ estimates in each month that fall below a threshold of $Ri<0.5$ (Figure 3-17g and h). This threshold of $Ri<0.5$ is chosen following the work of [Polzin, 1996] that provides evidence of elevated dissipation rates $\varepsilon$ even at the range $0.25<Ri<1$ (for more detail on that choice the reader is referred to Appendix E). Since vertical mixing is expected to be episodic, the number of occurrences of relatively low $Ri$ is likely to be a better indicator of the overall intensity of vertical mixing than the simple monthly mean of $Ri$. There is some qualitative agreement between the observational and model monthly-mean $Ri$ (Figure 3-17(e)-(f)), especially above the core of the EUC where the increased vertical shear between the overlying SEC and the EUC creates favorable conditions for mixing. Values of $\log(Ri)<0$ ($Ri<1$) occur in both model and observations approximately at the same time (December to April), concurrent with intensification of the EUC and core shoaling that intensify the shear between SEC-EUC. As expected, the core of the EUC is characterized by large $Ri$ with stratification exceeding shear by a factor of 10 in the
observations (log($Ri_{obs}$)−1) and by a factor of 100 in the model (log($Ri_{mod}$)−2). The difference in the $Ri$ values between model and observations is attributed to the differences in the values of both shear and stratification between the two estimates. Therefore, at the EUC core depth where the model shows larger stratification and smaller shear than the observations, the $Ri$ values are larger. Conversely, the larger values of $Ri$ at depths below 100 m in the observations could be due in part to measurement noise in the ADCP velocity profiles, where even small errors in $S^2$ can lead to low $Ri$ values in the presence of very weak stratification.

The percentage of $Ri<0.5$ in both model and observations has a semiannual character just above the core of the EUC with two maxima occurring in March and in December. An important feature in Figure 3-17(g) and (h) is another very clear maximum of the percentages of $Ri<0.5$ in July, evidence of mixing, when EUC gets weaker. In fact, high percentages of $Ri<0.5$ persist from March through July in the upper EUC part in both model and observations, when at the same time the westward wind stress is strongly increasing, allowing a continuous transport of westward momentum to be transported downward and decelerate the EUC. A similar qualitative explanation of the EUC weakening during summer was also suggested by [Johns et al., 2014]. Figure 3-17(g) and (h) differ significantly below the core of the EUC, where the much weaker stratification below 100m in the observations results in significant percentages of $Ri<0.5$ from January through April. On the contrary, the model below the core shows no evidence of small $Ri$. The fact that the comparison between model and observations’ $Ri$ is poor below the EUC does not pose significant problems for the rest of the analysis, since as shown in Figure 3-10(e) the ZDF term is most important at the upper part of the EUC where the comparison is qualitatively
meaningful.

The model’s $k_m$ displayed in Figure 3-18 (thick black profile) results from the ZDF term by integrating Eq. 3-5 using the surface wind stress $\tau_o$ as a boundary condition (Appendix E). After applying the different parameterizations described in Appendix E, only the ones that better captured the seasonal vertical structure of the model’s $k_m$ profiles were further tested within the observational framework. [Hummels, 2012; Hummels et al., 2014] (hereafter called H12) and [Pacanowski and Philander, 1981] (hereafter called PP81) (Figure 3-18). It is evident that one parameterization alone is not adequate to describe the model’s $k_m$ vertical structure: above the core of the EUC the structure of $k_m$ is better described by the H12 parameterization, while below the core, the $k_m$ structure of PP81

![Figure 3-18: Vertical diffusivity of momentum coefficient $k_m$ in the model (solid black line) and the vertical diffusivities for the most plausible $R_i$ based parameterizations to be tested in the observations. PP81 is shown in blue and H2012 is shown in grey. The core of the EUC is the horizontal red line.](image-url)
parameterization is closer to the model’s $k_m$. Therefore, a “hybrid” $k_m$ profile time series that is a combination of H12 above and PP81 below the core is used in this work (Appendix E). Both the H12 and PP81 have functional forms for which $k_m = f(R_i^n)$, with $n = -2$ for the PP81 and $n = -1.2$ for the H12 parameterizations and they are expected to have similar

![Diagram](image)

Figure 3-19: (a) Vertical diffusivity of momentum coefficient $k_m$ derived from the observations ($N^2$, $S^2$ and $R_i$) by merging the $k_m$ from H12 parameterization (surface to core depth) and the depth averaged $k_m$ from the PP81 parameterization from core depth to 230m. This the “hybrid” $k_m$. (b) Same as (a) but inferred from the model’s ZDF term. More details on the construction of those profiles are found in Appendix E.
trends: $k_m$ is expected to decrease (increase) with increasing (decreasing) $R_i$. The resulting time-mean “hybrid” $k_m$ profile is maximum closer to the surface, where the $R_i$ is small, and rapidly decreases within the first $\sim 50$ m (Figure 3-19(a)). The minimum at the core depth is associated with significantly small shear levels where the EUC is maximum that lead in turn to very high $R_i$ (Figure 3-17(c), (e)). Below the core of the EUC the increased shear between the undercurrent and the slower westward flows are sources of smaller $R_i$ and the hybrid $k_m$ increases to $2.4 \cdot 10^{-4}$ m$^2$/s, a value that is constant below approximately 75 m. The comparison between the observational $k_m$ (Figure 3-19(a)) to the model’s $k_m$ (Figure 3-19(b)) shows a good qualitative and quantitative agreement above the core, while below the core the model shows smaller values than the observational hybrid $k_m$. From September to December both model and observations show elevated $k_m$ values above $\sim 50$ m, while at the same time elevated $R_i$ values and low percentage of $R_i < 0.5$ cannot account for the high $k_m$s at those depths. However, according to H12 the functional dependence of $k_m$ in the observations and model above the core is (after using the equations in Appendix E) is

$$k_m = f\left(S^2, N^2\right)^{1.2}$$ (Eq. 3-9). Equation 3.9 reveals a stronger dependence of $k_m$ to the inverse of stratification than to shear. From September to December, stratification shows a pronounced near surface minimum in both model observations and according to Eq. 3-9, the smaller stratification is the more $k_m$ increases, providing an explanation for the seasonal $k_m$ maximum at the time. Since $k_m$ above the core in Figure 3-19(a) and (b) depends strongly on stratification, a deepening of the high $k_m$ follows the deepening of the lower stratification values.

The seasonal direct ZDF estimate (Figure 3-20(a)) derived using the hybrid $k_m$ profiles from 30-300 m and the CCMP surface wind stress as a surface boundary condition, has
significant qualitative similarities to the model’s ZDF (Figure 3-10(e)): (i) at the surface the ZDF seasonal cycle is the seasonal cycle of the wind stress (boundary condition), which is almost the same between model and observations, therefore they both show maximum westward near surface stress during spring, at the same time and location where the critical (<0.5) $R_i$ show their highest percentages, (ii) they both show reduced although still significant westward momentum at and below the EUC core. The latter is explained by expanding the vertical gradient of the product of km and shear of the zonal current in Eq. 3-5 using the chain rule: \[
\frac{\partial}{\partial z} \left( k_m \cdot \frac{\partial u}{\partial z} \right) = \frac{\partial k_m}{\partial z} \cdot \frac{\partial u}{\partial z} + k_m \cdot \frac{\partial^2 u}{\partial z^2} \ (\text{Eq. 3-10}).
\]
While at the core depth, where $u$ is maximum the first term on the right-hand side by definition becomes 0, the second term should be negative (westward) because it is the product of a positive number ($k_m$) and of the second derivative of a local maximum (inflection point) that also by definition should be negative. Below ~ 100m both model and observations show positive ZDF with no significant seasonality, the observational estimate though appears larger than the model and has a stronger tendency to accelerate the flow.

The seasonal direct observational ZDF yields a reasonable qualitative estimate, however that does not necessarily mean that quantitatively it would close the momentum balance. The residual in Figure 3-20(b) should not be directly compared to the residual ZDF in Figure 3-9(e). Figure 3-20(b) is the sum of all observational zonal momentum balance terms after using the direct estimate of the ZDF (hybrid) and provides a check on the overall momentum balance closure, whereas the 3-9(e) is an estimate of the ZDF as a residual. Two points are highlighted on Figure 3-20(b): (1) the observational ADV is not a localized estimate therefore can be considered a source of momentum imbalance and (2)
the direct ZDF although qualitatively sound, clearly is responsible for overestimating the momentum budget along the core of the EUC, where the residual takes the maximum value of $+3 \cdot 10^{-7}$ m/s$^2$. The overall conclusion of the direct estimate of ZDF using observations is that it can only yield a qualitative estimate.

Figure 3-20: (a): Vertical dissipation of momentum using a “hybrid” vertical stress $k_{os}$ (combination of H12 and PP81, see text for details). (b) Residual of the observational momentum balance estimated using equation 3-4.
3.5.5 The role of eddies in the zonal momentum balance

A question that could be addressed in view of the momentum balance is that of the potential role that eddies have on a seasonal time scale. Although the effect of eddies is accounted for in the climatological mean of the advection terms (see also section 3.4.3), the role of eddies in the momentum balance can still be assessed. This estimate of the lateral eddy component can be inferred by differencing $\frac{\partial \bar{u}}{\partial x}$ with $\bar{u} \frac{\partial u}{\partial x}$ (and similarly for $v \frac{\partial v}{\partial y}$) that are in turn estimated from the monthly climatological analysis:

$$eddy_{\text{H}} = \left( u' \frac{\partial u'}{\partial x} + v' \frac{\partial u'}{\partial y} \right) = \left[ \left( \bar{u} \frac{\partial u}{\partial x} - u \frac{\partial \bar{u}}{\partial x} \right) + \left( \bar{v} \frac{\partial v}{\partial y} - v \frac{\partial \bar{v}}{\partial y} \right) \right] \text{(Eq. 3-11)}.$$

The results from Eq. 3-11 for both model and observations are shown in Figure 3-21(a, b) and while both estimates have similar magnitudes, they show some differences in seasonality and vertical structure. There is some similarity above the EUC core, mostly negative values and with a
minimum in both model and observations during summer, although the observations become positive in fall and at the same time the model remains weakly negative. There is a surprising similarity between the model’s lateral eddy contribution and the zonal advection term, however when the contributions of each term in Eq. 3-11 are considered.
separately (Figure 3-23), the meridional terms ((d)-(f)) are the ones that play a dominant role over the zonal terms. In the observations, the meridional advection terms (mean and eddy components) also play a more dominant role in the lateral eddy momentum flux divergence (Figure 3-22). For both model and observations, the inferred zonal and meridional eddy momentum flux divergences are much smaller than the mean zonal and meridional advections, but when added together the lateral eddy momentum flux has a non-negligible contribution to the balance.

The mean annual lateral eddy contribution to the momentum flux divergence (Figure 3-24(a, b)), shows very clearly a tendency to decelerate the EUC above the core, almost exclusively due to the meridional eddy momentum flux divergence in both model and observations. Below the core, the model shows a clear tendency to accelerate the EUC down to ~200m, but in the observations, it only weakly tends to accelerate the core for a few meters and then shows a weak tendency to decelerate the EUC. Below the core in the model, the zonal and meridional eddy momentum flux divergences contribute equally to the lateral eddy momentum flux divergence that tends to accelerate the EUC below the core. In the observations, the meridional eddy momentum flux divergence dominates above and below the core. The vertical structure of the lateral eddy momentum flux divergence in the model, is similar to what [Bryden and Brady, 1989] showed in an observational study of the Pacific EUC, that the eddies tend to decelerate/accelerate the EUC above/below the core. As shown here in both model and observations (Figure 3-24(a, b)) the meridional eddy flux divergence in [Bryden and Brady, 1989] supplies almost exclusively the eddy momentum flux divergence.
The magnitude of the seasonal and annual mean ADV is larger than the seasonal annual mean inferred lateral eddy momentum flux divergence for both model and observations. In the model’s case that the ADV is better resolved, the annual mean ADV tends to accelerate/decelerate the EUC above/below the core (Figure 3-25) which is opposite to the effect of the eddies. In an analogous manner, the seasonal eddy momentum flux divergence (Figure 3-21(b)) in the model mainly opposes the net accelerating/decelerating role of the seasonal ADV (Figure 3-10(c)) from January to June, while the rest of the year the eddies

![Obs mean eddy contribution in the ZMB](image)

![Model mean eddy contribution in the ZMB](image)

Figure 3-24: Annual mean zonal and meridional contributions to the inferred lateral eddy momentum flux in the (a) observations and (b) model.
contribute weakly to the ADV. Therefore, the role of the eddy momentum flux divergence in the model is to weakly constrain—due to its small magnitude—the effect of the ADV to the overall momentum balance almost all year long.

3.5.6 Mean annual zonal momentum balance

In the previous sections of this work while an estimate of the seasonal momentum balance has been attempted, it has been proven to be quite challenging in the observations because of: (i) the methodologies used for the calculation of the non-linear advection terms, (ii) the difficulty in calculating directly the vertical diffusion terms and (iii) the large uncertainties that result due to (i) and (ii). It is therefore useful to also consider the annual mean zonal momentum balance, which may help to average out some of the errors and uncertainties (Section 3.5.3) and also provide a more direct comparison with previous studies (i.e. [Wacongne, 1989; Qiao and Weisberg, 1997]). The annual mean momentum balance on the equator has been the focus of most momentum balance studies thus far (i.e. [Bryden and Brady, 1985, 1989; McPhaden and Taft, 1988; Qiao and Weisberg, 1997]).

On Figure 3-25, the mean profiles are estimated from the seasonal cycles of each term of the momentum balance for the observations and the model (Figure 3-9, Figure 3-10 and Figure 3-20(a)). The annual ste of each term in Figure 3-25 is similarly estimated from the ste seasonal cycles (Figure 3-15) following $\sigma_{\text{mean}} = \sqrt{\frac{\sum_{i=1}^{12} \sigma_i}{12}}$, where $\sigma_i$ is the monthly ste for each term.

In the annual mean momentum ZMB, where the local acceleration is by definition 0, the balance between ADV, ZPG and ZDF is examined to reveal different regimes above and below the core of the EUC, following in part the analysis performed by [Wacongne,
1989] and [Qiao and Weisberg, 1997]. There are obvious similarities between model and observations: (i) the ZPG is positive above the EUC core, (ii) the hybrid ZDF is strong and westward down to ~100m, (iii) the ADV is negative induces maximum westward momentum decelerating the lower EUC at about 100m, (iv) above the EUC core the ADV is mainly set by partial compensation between the vertical and meridional advection, while below the core the ADV is set by the compensation between the sum of meridional and vertical advection that is negative and the zonal advection. The observational annual mean ZMB (Figure 3-25(a)) shows both the ZDF estimates (residual and hybrid), but since the hybrid ZDF gives a large residual (Figure 3-20(b)) and does not close the momentum balance only the residual ZDF will be considered in the discussion of the annual mean momentum balance.

From the surface to 50m layer that is only resolved in the model, the sum of strong positive ADV and ZPG is balanced by the ZDF (Figure 3-25(c)). The ADV in the model is in the near surface depth of 0-20m almost exclusively attributed to positive meridional advection and to a lesser extent to vertical advection (Figure 3-25(d)). Exactly at the surface the ADV (2.8·10^{-7} m/s^2) is almost exclusively meridional (3·10^{-7} m/s^2) (Table 3-3). Below 20m down to ~50m the ADV although stays positive and tends to accelerate the EUC, it decreases as the vertical advection is offset at depth by strong westward meridional advection. At 50m the ADV \rightarrow 0 and meridional advection becomes from 2.8·10^{-7} m/s^2 at the surface -2.4·10^{-7} m/s^2 at 50m and the vertical advection from 0 at the surface increases
to $2.2 \cdot 10^{-7}$ m/s$^2$ at 50m (Table 3-3). [Wacongne, 1989] at 10°W in the Atlantic EUC (Figure 5 in [Wacongne, 1989]), showed the same annual mean positive ADV at the 0-50m, that gradually reduces with depth and together with positive ZPG balances the strong westward

![Figure 3-25](image)

Figure 3-25: (a) Annual mean zonal momentum balance for the observational estimate, (b) annual mean advection of zonal momentum for the observations, (c) same as (a) for the model end (d) same as (b) for the model.
ZDF and characterized that balance as “inertio-frictional” regime. \cite{Qiao and Weisberg, 1997}, showed in the Pacific EUC the same regime of positive ADV that added to the positive ZPG balances the ZDF at 50m (Figure 14 in \cite{Qiao and Weisberg, 1997}).

From \(~50\)m to the core of the EUC, in the observations the positive ZPG that tends to accelerate the EUC is balanced by negative and westward ZDF and ADV (Figure 3-25(a)), while in the model where the ADV is almost zero, positive ZPG balances the ZDF (Table 3-2). It is remarkable how the annual mean net ADV is almost exactly zero over the upper part of the EUC even though the individual contributions are all large there. The negative ADV in the observations is the result of meridional advection \((-0.7\cdot10^{-7}\) m/s² at the core) exceeding the vertical advection \((0.2\cdot10^{-7}\) m/s² at the core) while the zonal advection is negligible (Figure 3-25(b)) at the core depth (Table 3-2). \cite{Qiao and Weisberg, 1997} also showed ADV that tends to decelerate the EUC at the core positive ZPG is balanced by the sum of negative ADV and ZDF (their “weakly non-linear” regime). In the model the ADV is almost zero from 50m to the core depth due to the balance between the strongly negative meridional advection and the sum of positive vertical advection (reducing with depth) and positive zonal advection (increasing with depth (Figure 3-25(d)). At the core depth, the observations show that the ADV balances the ZPG and ZDF tends to 0 in the case of the residual ZDF and consequently the \(-2.5\cdot10^{-7}\) m/s² value of the hybrid ZDF equals the residual at core depth (Table 3-2). The model however shows at the core that the positive ZPG is 89% compensated by the ZDF while ADV accounts for only 11% of the needed westward momentum to close the balance (Table 3-3).

Below the core of the EUC to \(~80\)m, the positive ZPG is balanced by negative ADV and ZDF in the observations, then the ZPG becomes weakly negative down to \(~140\)m. This
weakly negative ZPG and negative ADV below the core have to be balanced by positive eastward ZDF in the observations, that is interestingly also confirmed by the hybrid stress with a small residual at 135m (Figure 3-25(a), Table 3-2). The weak negative ZPG below ~80m is a significant difference between model and observations, since this is the main reason why the residual ZDF has to be positive, which is unrealistic, suggesting that the Argo derived ZPG is underestimating the ZPG in the lower part of the EUC. The same regime of westward ADV and ZDF balanced by eastward ZPG is shown in the model below the core to ~120m. At 135m the model shows a balance between negative ADV and positive ZPG, while below that level the ADV becomes zero and the signs of ZPG and ZDF reverse to weakly negative at 200m (-0.1·10^{-7} m/s^2 at 200m) and weakly positive (0.1·10^{-7} m/s^2 at 200m).

Adapting the definitions given in [Qiao and Weisberg, 1997] for the momentum balance regimes, there are 4 different regimes that can describe the annual mean momentum balance in the observations and the model: (i) a linear regime where the ZDF is balanced by the ZPG, (ii) a strongly non-linear regime where all terms (ADV, ZPG and ZDF) are non-negligible and ADV plays an important role, (iii) a weakly non-linear regime where again all terms are non-negligible but ADV plays a weaker role compared to the rest of the terms in the balance and (iv) a non-linear regime where ADV balances either the ZDF or the ZPG. Following the above definitions, the balance at various depths (Tables 2 and 3) reveals zonal flow regimes that are different between observations and model at all depths. In the observations (residual ZDF) the flow regimes are: (1) weakly non-linear at 50m, (2) non-linear at the core depth (3) linear below the core at 135m with a negative ZPG and positive ZDF and (4) linear at 200m with positive ZPG and negative ZDF. In the
model, the flow regimes are different: (1) strongly non-linear at the surface, (2) linear at 50m with positive ZPG and negative ZDF, (3) weakly non-linear at the core, (4) non-linear below the core at 135m and (5) linear with negative ZPG and positive ZDF at 200m.

Table 3-2: Annual mean momentum balance at 10ºW using observations.

<table>
<thead>
<tr>
<th>DEPTH</th>
<th>(\frac{\partial u}{\partial t})</th>
<th>-u (\frac{\partial u}{\partial x})</th>
<th>-v (\frac{\partial u}{\partial y})</th>
<th>-w (\frac{\partial u}{\partial z})</th>
<th>ADV</th>
<th>ZPG</th>
<th>ZDF</th>
<th>residual</th>
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<td>-1.3 ± 0.1</td>
<td>1.3 ± 0.1</td>
<td>-0.1 ± 0.2</td>
<td>1.0 ± 0.1</td>
<td>-0.9 ± 0.2</td>
<td>-1.8 ± 0.1</td>
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<tr>
<td>CORE</td>
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<td>0.2 ± 0.0</td>
<td>-0.7 ± 0.1</td>
<td>0.0 ± 0.1</td>
<td>-0.5 ± 0.1</td>
<td>0.5 ± 0.1</td>
<td>0.0 ± 0.2</td>
<td>-2.5 ± 0.1</td>
</tr>
<tr>
<td>135m</td>
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<td>0.1 ± 0.0</td>
<td>0.0 ± 0.1</td>
<td>-0.1 ± 0.0</td>
<td>0.0 ± 0.1</td>
<td>-0.2 ± 0.1</td>
<td>0.2 ± 0.1</td>
<td>0.3 ± 0.1</td>
</tr>
<tr>
<td>200m</td>
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<td>0.0 ± 0.0</td>
<td>0.0 ± 0.0</td>
<td>0.0 ± 0.0</td>
<td>0.0 ± 0.0</td>
<td>0.0 ± 0.0</td>
<td>0.0 ± 0.0</td>
<td>-0.1 ± 0.1</td>
</tr>
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Table 3-3: Annual mean momentum balance at 10ºW using the model.

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<th>-u (\frac{\partial u}{\partial x})</th>
<th>-v (\frac{\partial u}{\partial y})</th>
<th>-w (\frac{\partial u}{\partial z})</th>
<th>ADV</th>
<th>ZPG</th>
<th>ZDF</th>
<th>residual</th>
</tr>
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</tr>
<tr>
<td>50m</td>
<td>0.0 ± 0.0</td>
<td>0.2 ± 0.2</td>
<td>-2.4 ± 0.2</td>
<td>2.2 ± 0.1</td>
<td>0.0 ± 0.2</td>
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<td>-1.7 ± 0.1</td>
<td>0.0 ± 0.0</td>
</tr>
<tr>
<td>CORE</td>
<td>0.0 ± 0.0</td>
<td>1.3 ± 0.2</td>
<td>-1.2 ± 0.2</td>
<td>-0.2 ± 0.1</td>
<td>-0.1 ± 0.1</td>
<td>0.9 ± 0.1</td>
<td>-0.8 ± 0.1</td>
<td>0.0 ± 0.0</td>
</tr>
<tr>
<td>135m</td>
<td>0.0 ± 0.0</td>
<td>0.7 ± 0.1</td>
<td>-0.4 ± 0.1</td>
<td>-0.4 ± 0.1</td>
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<td>0.1 ± 0.0</td>
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<tr>
<td>200m</td>
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<td>-0.1 ± 0.0</td>
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3.6 Discussion

This study provides a first ever assessment of the zonal momentum balance along the equator in the Atlantic based on observations, focusing on the processes that control the seasonal cycle of the EUC. The Tropical Atlantic Climate Experiment (TACE) mooring array provided an adequate number of measurements to resolve zonal circulation in the ETA between 23ºW and 0ºE and are combined with other observations (Argo, PIRATA and altimetry) to examine the ZMB near 10ºW. The output from a model simulation of the equatorial Atlantic, which realistically captures the seasonal cycle of the EUC, was used to test the methods and assumptions in the observational estimate, while also investigating, in a parallel analysis, the momentum balance within its framework. The seasonal and
annual mean observational and model zonal momentum balances at 10ºW resulting from the present work are examined and evaluated in a detailed error analysis, while the role of eddies in the balance was also discussed. This section is synthesizing all the information provided by the zonal momentum balance analysis to answer fundamental questions about what is driving the seasonal changes in strength and vertical structure of the EUC, using both the observational and model results.

The core velocity of EUC at 10ºW has a semiannual cycle in the observations with maximum velocities in mid-March (primary) and mid-September (secondary) and minimum in September (Figure 3-9(a) and Figure 2-3), while in the model, the EUC maximum velocity has a very weak semiannual, closer to annual, seasonal cycle with a maximum in March and much weaker one in October and a minimum in July-August (Figure 3-10(a) and Figure 2-3). [Kolodziejczyk et al., 2014] showed similar EUC seasonal cycle from a model simulation that showed very weak, almost no secondary EUC core maximum in fall. At the same time, the observational seasonal EUC in [Kolodziejczyk et al., 2014] showed also a weak semiannual signal, although the time period used to calculate the seasonality from mooring measurements was different (2003-2007) than the TACE period (2007-2011). Therefore, the presence of the model’s secondary EUC core maximum in October shows local acceleration that tends to, but does not cross 0 (Figure 3-11(b), is questionable. In order to answer questions like the presence/absence of the semiannual EUC core velocity variability in the observations/model at 10ºW, it is important to understand what is the forcing balance that controls the EUC core minima and maxima.

The presence of the primary spring EUC maximum \( \left( \frac{\partial u_{core}}{\partial t} = 0 \right) \) in both model and observations is attributed to the following flow regimes: (i) to a linear regime in the model
in March where the eastward ZPG balances the westward ZDF (ADV=0), and (ii) to a strongly non-linear regime where westward strong ADV and ZDF (|ADV|>|ZDF|) balance the ZPG in the observations. The linear regime in (i) for the model is the result of a weakly non-linear regime from December through March, where the ADV stays small and almost zero and the eastward ZPG slightly exceeds the westward ZDF providing continuous decreasing acceleration to EUC until March (EUC maximum in the model). The strongly non-linear regime described in (ii) for the observations occurs one month after the maximum acceleration in January-February, that is in turn caused by a non-linear regime where the acceleration results from strong eastward ZPG that exceeds the sum of almost

Figure 3-26: Seasonal observational momentum balance using residual stress at various depths (like in Figure 3-11(B)). “Surface” in the observations is equivalent to the 50m depth level.
equal amounts of westward ADV and ZDF (Figure 3-26). Although the ZPG becomes weaker at core depth from February to April, the constant sum of ZDF and ADV still doesn’t exceed the ZPG and consequently the EUC is supplied by continuously smaller amounts of acceleration until it reaches its maximum in mid-March-April (observations). The spring EUC core velocity maximum is well reproduced in the study of [Brandt et al., 2016a] at 23ºW by using a linear dynamics, reduced gravity model. In their study [Brandt et al., 2016a] found that the superposition of two baroclinic modes (second and fourth) can be used to describe the seasonal cycle of the EUC at the core depth. Although in [Brandt et al., 2016a] 23ºW is the only longitude where the EUC’s core velocity is presented, the realistic reproduction of the EUC transport seasonal cycle at 10ºW, compared to [Johns et al., 2014] (Figure 11(b) in [Brandt et al., 2016a]), may indicate that linear dynamics provides a plausible explanation of the EUC’s core velocity seasonal variability. In that case, the linear balance suggested in the model to explain the EUC core maximum in March, is also confirmed by [Brandt et al., 2016a]. Therefore, the spring EUC maximum at core depth in the model can be dynamically explained by the superposition of the second and fourth baroclinic modes (representing the semiannual and annual cycles respectively). Additionally, the weakly non-linear momentum balance regime in the model from December to about May supports the [Brandt et al., 2016a] explanation for the EUC core velocity seasonal variability for half of the year. However, the spring maximum in the observations that results from a non-linear regime where ADV plays a significant role, suggests that linear dynamic probably cannot adequately represent the mid-basin ETA EUC core velocity seasonal variability at 10ºW.
In the model during October when the questionable, weak secondary EUC core maximum occurs, the sum of ZPG and ADV (both positive, eastward and of the same magnitude), tends to but not exactly balances the westward ZDF. The EUC fall maximum in the model results from a strongly non-linear flow regime that starts in mid-July and lasts through November and during which the ADV tends to strongly decelerate (July-September) and then strongly accelerate the EUC core (September-November). In the observations, the secondary fall maximum occurs in late September (Figure 3-26, “core”). However, the forced closure of the observational momentum balance when using the residual ZDF, calls for a very strong eastward ZDF to balance the ZPG and ADV (both negative and westward). This very strong ZDF is a feature that is highly questionable (sections 3.5.1 and 3.5.4) and a balance that is exactly opposite to the one suggested by the model. However, the negative ADV in the observations during the secondary fall EUC maximum, attributed to the meridional advection term (Figure 3-12), is a result of the methods used in the calculation of the observational advection that introduce biases. During fall and at core depth, the possible methodological error estimated within the model’s framework (Figure 3-13) shows a seasonal peak. Additionally, the model’s ADV estimated using the observational methods shows negative ADV (Figure 3-14) like in the observations, concluding that the model’s strongly non-linear flow regime explains the fall EUC maximum while the fall ADV in the observations is most likely a biased estimate. The fall EUC core maximum occurs in both cases (model and observations) after a strong contribution of the non-linear advection term, suggesting that linear dynamics and simple superposition of basin modes at in [Brandt et al., 2016a] would most likely not accurately reproduce the EUC core velocity seasonal cycle at 10ºW at that time.
Regarding the EUC summer minimum, occurring in July in the observations and mid-July in the model (Figure 3-9, Figure 3-10 and Figure 2-3), the momentum balance reveals slightly different flow regimes being responsible for the EUC core deceleration maximum in the observations and in the model. In the observations, the residual momentum balance (Figure 3-26) shows zero local deceleration in July as result of a weakly non-linear flow regime where westward ADV and ZDF ($|ADV|<|ZDF|$) balance the eastward ZPG, a month after the ZPG’s seasonal maximum. In the model, local deceleration crosses zero in mid-July-August and the flow begins to accelerate, at a time when the ZPG is again balanced by westward ZPG and ADV. The flow regime in the model in July-August is close to strongly non-linear since $|ADV|>|ZDF|$. Consequently, the momentum balance in both model and observations shows an EUC core minimum under non-linear (weakly or strongly) flow regimes and EUC core maxima under weakly or strongly non-linear flow regimes as well, with an exception being the linear balance in the model in March maximum. An important conclusion of this present work is that, from a momentum balance perspective, mid-basin in the ETA at 10°W non-linear advection terms are very important for the EUC core seasonality. This confirms what [Qiao and Weisberg, 1997] showed in their observational study of the Pacific EUC about the importance of including non-linear ADV in the momentum balance.

The EUC core migrates vertically on a seasonal time scale. The core of the EUC at 10°W exhibits an annual cycle with shallowest depth appearing in April and deepest core location appearing in September, not in phase with the seasonal cycle of EUC core velocities (Figure 3-9) that is semiannual. As in the observations the core depth in the model exhibits a different seasonal cycle than the core velocities with maximum depths
reached in April (and a weaker maximum in October) and a minimum in July. This off-phase vertical migration of the core with the maximum velocities of the EUC, also shown by [Brandt et al., 2014] at 23ºW and [Kolodziejczyk et al., 2009] at 10ºW and seen in the model, reflects different forcing balances, different flow regimes (Section 3.5.6). The shallowest depth of the EUC is in April in both model and observations, when in the observations a strong non-linear regime is accelerating the EUC with eastward ZPG exceeding the sum of westward ADV and ZDF, while in the model in April, the EUC is weakly decelerated due to strong westward ZDF that exceeds the sum of strong eastward ZPG and weak ADV (weakly non-linear regime) (Figure 3-11 and Figure 3-26). In [Brandt et al., 2016a] the vertical displacement of the EUC core at 23ºW is explained by linear dynamics and has an annual character, like in the observations at 10ºW, that from a linear dynamics perspective is dominated by the fourth baroclinic mode. In the study of [Brandt et al., 2016a] the spring EUC core at 23ºW is its shallowest depth in April and is dynamically explained by the vertical structure of the fourth baroclinic mode that crosses zero approximately at the EUC core depth, therefore “strengthening the EUC above the core and weakening the EUC below”, resulting in the shoaling of the core. The model shows in March-April (Figure 3-11(a)) a shoaling of the ZPG maximum values and also shoaling of the ZDF maximum values. Under the weakly non-linear regime at core depth in spring in the model, were the ADV plays a subtle role in the flow, the larger ZPG values at shallower depths can increase the zonal momentum of the upper part of the EUC causing the shoaling of the core, consistent with the linear dynamics explanation provided by [Brandt et al., 2016a]. In the observations, during April the residual ZDF is weakly negative (due to the strong decelerating ADV) and although the ZPG also shows a
minimum at that time (Figure 3-26, “core”), the core effectively moves upwards.

During the July EUC core deepening in the model, the eastward ZPG is deepening in July-August showing elevated values at greater depths. At the same time, ZDF has started to extend deeper 1-2 months prior to the ZPG deepening, causing the most enhanced and effective loss of momentum in the upper EUC in May-June (Figure 3-11 (a)), that along with the strong westward ADV pushes the core deeper. A similar explanation for the summer core deepening but at 25ºW is given in [Wacongne, 1989], where she suggests that the ZPG effect penetrates deeper than the opposing ZDF effect, causing deceleration of the upper EUC (due to ZDF) and acceleration of the lower EUC (due to ZPG) that effectively pushes the core downward. In the observations, the maximum core depth in September occurs under a rather unrealistic balance, where strong eastward ZDF is balanced by strong ADV while ZPG is almost zero, and therefore from a momentum balance perspective, the EUC core deepening in September in the observations cannot be explained. It is rather [Brandt et al., 2016a] that provide an explanation for the fall core deepening (but at 23ºW) in view of linear dynamics: the fourth baroclinic mode is responsible for weakening the EUC above the core and strengthening the EUC below the core causing downward displacement of the core. The [Brandt et al., 2016a] mechanism that explains the fall EUC core deepening is very similar to the one suggested by [Wacongne, 1989] probably because their studies concern similar longitudes (23ºW and 25ºW) respectively, although they show maximal depths at slightly different times of the year.

The eastward velocities that extend well below the core of the EUC, start to appear as early as March (Figure 3-9 (a)) in the observations and February in the model (Figure 3-10 (a)) and last through September in both model and observations [Kolodziejczyk et al., 2009;
Brandt et al., 2014; Johns et al., 2014]. The observational studies of [Kolodziejczyk et al., 2009] and [Johns et al., 2014] provide evidence, from shipboard and mooring ADCPs that the presence of the EUC deep extensions can be attributed to transient (not occurring every spring-summer), subsurface eastward cores at about 200-400m that appear to be separate from the EUC. From a momentum balance perspective, the presence of the deep extensions is closely linked to the seasonality of the ZPG below the core of the undercurrent. In the observations, below ~100m the local acceleration is almost in phase with the ZPG, producing the deep extensions (Figure 3-9). The ADV modifies effectively what would be an almost linear balance between local acceleration and ZPG in the observations especially in May to June, when westward ADV acts to offset the EUC acceleration and deep extension strength, and also during August when eastward ADV tends to sustain the EUC deep extension (Figure 3-26, “150m”). In the model, the momentum balance at 150m shows the ZPG to be almost in phase with the local acceleration all year long, especially from March to October when acceleration below the core is positive causing the presence of the deep extensions (Figure 3-11 (b)). The ADV in the model at the time of the EUC deep extensions plays a very small role, therefore the flow regime is an almost linear balance.

Dynamically, the EUC deep extensions can be adequately explained in view of linear dynamics [Brandt and Eden, 2005]. Equatorial Kelvin waves that are excited by seasonally variable wind stress at the western part of the tropical Atlantic basin can be reflected as equatorial Rossby waves at the eastern boundary that are dominated by higher than 3rd baroclinic modes, as shown by [Brandt and Eden, 2005]. The Rossby waves are related to downward energy propagation of energy, enhancing the deep eastward flow at depths
below the EUC core. While [Brandt et al., 2016a] describe reasonably the EUC variability using the superposition of two vertical modes (2nd and 4th) at 23ºW, the reconstruction of the EUC transport at both 23ºW and 10ºW using those modes does compare to the observations as favorably. [Brandt et al., 2016a] attribute the differences between the simulated from basin mode reconstruction and observational transports, to the fact that in their simulations the deep EUC extensions are not accurately represented.

A large part of this work was dedicated to the estimate of a direct, instead of residual ZDF term in the observations. While quantitatively this attempt failed, qualitatively gave reasonable results confirming the importance of vertical dissipation especially during summer. The weakening of the EUC core during summer at 10ºW, while ZPG shows a seasonal maximum can be explained by strong and deep penetrating westward ZDF, as has been suggested by [Johns et al., 2014]. Here the model clearly shows a weakening of the upper EUC due to westward ZDF that is maximum in early summer (Figure 3-10(e)), consistent with the westward wind stress maximum (Figure 2-22(b)). In the observations, the residual ZDF in summer is negative near 50m but quickly reverses to strongly positive, which is highly unrealistic. However, the hybrid vertical stress term accurately represents a maximum westward ZDF in June that can weaken the upper EUC even though the ZPG is still strong (Figure 3-20). This direct hybrid ZDF results from a vertical diffusion coefficient $k_m$ that is combination of different $Ri$ based $k_m$ parameterizations. The investigation of the appropriate $k_m=f(Ri)$ is done within the model’s framework, although from an oceanographic modeling perspective this may seem a drawback. [Blanke and Delecluse, 1993] in a pioneering work, were among the first ones to show how the use of $k_m$ as a function of turbulent kinetic energy (TKE), using a prognostic TKE closure scheme
that is explicitly solved for each model time step, significantly improves the simulation of
the EUC in ocean General Circulation Models (GCMs). In the test simulation of [Blanke
and Delecluse, 1993] with \( k_m = f(TKE) \) the EUC was stronger and penetrated deeper in the
eastern part of the basin, favorably comparing with observations. The model used in this
work uses also a vertical diffusion scheme in which the \( k_m \) is a function of TKE. However,
the schemes used in the models cannot be applied in the observations since they include
prognostic equations that are solved numerically, not analytically, and are part of the
model’s system of primitive equations. It is therefore because of their simplicity that the
\( k_m = f(Ri) \) parameterizations are considered in this study (Appendix E) and the fact that they
can be estimated easily from observations (T/S profiles and currents). Ideally, the
microstructure measurements that have recently been added to the PIRATA mooring array
will provide new insights on the ZDF term, especially since the \( \chi \)-pods (or chi-pods) that
have been successfully deployed in the TAO/TRITON array [Moum et al., 2013], provided
robust evidence of seasonal mixing in the equatorial Pacific.

The results of this study show that it is indeed very challenging to construct a closed
momentum budget for the EUC based only on observations, even with the addition of the
extensive data set provided by the TACE program (Figure 2-2). The fact that the
observational estimates are a mixture of terms that are effectively calculated locally (e.g.
\( \frac{\partial u}{\partial y} \) ) or represent zonally averaged values (e.g. ZPG) or approximations to zonal averages
(e.g. \( u \frac{\partial u}{\partial x} \)), makes it problematic to compare the result directly to either a purely local or a
zonally-averaged momentum balance from the model. However, the qualitative agreement
that is found between the observationally-derived and model momentum balances provides
a measure of confidence in the models results, which can therefore be used to investigate the momentum balance in a fully closed form. This, as well as the model's ability to reproduce quite accurately the seasonal cycle of the EUC established from the TACE observations, has provided new insights on the EUC's momentum balance and an updated understanding of the main processes controlling the seasonal cycle of the EUC in the ETA.

3.7 Conclusions

This chapter presents the estimate and analysis of the zonal momentum balance of the Atlantic EUC as derived from observations in the central Eastern Tropical Atlantic basin, at 10°W. To complete the picture, the output of a model simulation has been invaluable to this study. While the EUC seasonal cycle at 10°W compares well between model and observations, the individual momentum balance terms show also significant differences between the model and observational estimates. These differences, are not always attributable to the methods used for the observational estimates but also reveal different flow regimes between model and observations. A detailed analysis of both the methods used and the errors for each term in the balance, provide a synthesized attempt to estimate the seasonal momentum balance in the Atlantic EUC.

Principal conclusions from the study are:

- From a momentum balance perspective, mid-basin in the ETA at 10°W the EUC core seasonality can only be accurately reproduced at 10°W when non-linear terms are taken into account. During the fall EUC core maximum that occurs in both model and observations, the strong contribution of the non-linear advection term suggests that linear dynamics and simple superposition of basin modes would not
accurately reproduce the EUC core velocity seasonal cycle at 10°W. Additionally, the momentum balance in both model and observations shows an EUC core minimum under non-linear (weakly or strongly) flow regimes and EUC core maxima generally under weakly or strongly non-linear flow regimes as well, further highlighting the importance of non-linear ADV in the Atlantic EUC momentum balance.

- In the model, the shallowest core depth in spring occurs under a weakly non-linear regime. The larger ZPG values are found at shallower depth, increasing the zonal momentum of the upper part of the EUC above the core and causing the shoaling of the latter. This is consistent with the linear dynamics explanation [Brandt et al., 2016a]. In the observations, during April the residual ZDF appears to be weakly negative (due to the strong westward ADV) and although the ZPG also shows a minimum at that time (Figure 3-26, “core”), it is still strong enough at shallower depths to effectively move the core upwards by strengthening the upper EUC. During maximum EUC core deepening in the model (July) the eastward ZPG is deepening showing elevated values at greater depths but the westward ZDF has started to extend deeper 1-2 months prior to the ZPG deepening, causing the most enhanced and effective loss of momentum in the upper EUC that continues through July and along with the strong westward ADV pushes the core deeper.

- The observational seasonal momentum balance below ~150m is described by a weakly non-linear flow regime, with the eastward ZPG following the local acceleration and ADV partially offsetting the EUC acceleration and therefore the deep extension in early summer and sustaining the EUC acceleration in late
summer. Below ~150m the seasonal zonal momentum balance in the model is described by a linear flow regime with the eastward ZPG being almost in phase with the local acceleration and being primarily responsible for the existence of the EUC deep extensions.

- The observational ADV is not a localized estimate therefore can be considered a source of “momentum imbalance”. This is confirmed by the comparison of zonally-averaged (within ±5° of longitude centered at 10°W) non-linear terms from the model and the non-linear terms derived from model fields using the observational methods, that provides a more reasonable comparison between model and observations advection terms, at least for the upper part of the EUC.

- The seasonal direct ZDF has significant qualitative similarities to the model’s ZDF. However, although the observational direct ZDF is qualitatively reasonable, quantitatively it is not able to close the momentum balance and large momentum biases along the core of the EUC remain.
Chapter 4 Overall conclusions and future work

The present dissertation is a contribution to the Atlantic EUC studies by addressing two major questions: (i) what is the seasonal upwelling transport in the Eastern Tropical Atlantic (ETA) that is related to the EUC transport and (ii) what is the forcing that controls the seasonality of the EUC. The first question is addressed in view of the seasonal mass balance, while the second question is addressed by the seasonal momentum mass balance, using a unique data set of observations collected in the area of the ETA over the last 15 years and in particular during the Tropical Atlantic Climate Experiment (TACE) (Figure 2-2). The 5-year daily output of a regional high resolution (1/4°x1/4°) model simulation of the Tropical Atlantic was a valuable complement to the observational estimates of mass and momentum balances, providing not only a test-bed for various methodological sensitivity tests but also a great comprehensive tool when observations were not enough to resolve certain aspects of the balances. This work is one of the few observational studies that quantifies the seasonal cycle of upwelling in the ETA, one of the few attempts to study the zonal momentum balance on a seasonal time scale, and the first observationally based zonal momentum balance study in the equatorial Atlantic EUC.

The mass balance method in the 50-300m depth range (subsurface box), reveals significant differences between the western and the eastern boxes mainly in the role that zonal and meridional convergences play in supplying the observed upwelling. Both zonal mass flow and meridional geostrophic mass flow in the western subsurface box are
important in supplying the inferred upwelling, although the zonal convergence contributes more. The mean annual value upwelling in the western subsurface box has a value of 3.6±3.3 Sv. In the eastern subsurface box, the annual mean upwelling is estimated as 3.0±2.2 Sv, approximately all of which (~90%) is supplied by zonal transport convergence (3/4 from the EUC and 1/4 from westward flow), and the remaining ~10% of it is attributed to meridional geostrophic inflow.

The model provides a testbed for the sensitivity of the methods used in deriving upwelling estimates inferred from mass balance. After the methodologies are tested in the model, it is shown that the assumption of “no vertical flow” at the bottom of the subsurface box, although not true at all times of the year, introduces a relatively small uncertainty of about 1 Sv in the upwelling estimates. Examining the differences between the meridional transport divergence and the meridional geostrophic transport divergence, is another methodological test performed within the model’s framework. The seasonality of the total meridional transport is well represented by its geostrophic component and in turn the seasonality of the geostrophic transport in the model resembles closely the seasonality of the geostrophic transport in the observations.

The near surface meridional mass divergence drives the near surface upwelling, according to the traditional view of the equatorial upwelling, but once the surface waters have moved poleward, the upwelling waters from deeper layers are not supplied as strongly by the meridional convergence but by the EUC, explaining the strong signal of the subtropical water masses in the equatorial Atlantic upwelling zone. The seasonality of the upwelling can be partially explained by the seasonality of ZPG, but the lagged response of
the zonal flow to the ZPG can only be explained for the deeper part of the flow (below 100m) with the present observational analysis leaving room for future studies.

Strong meridional outflow in the eastern subsurface box accounts for a large fraction of the EUC mass losses between 10ºW and 0ºE during fall, approximately 50% in the observations and 75% in the model with the remainder going to supply the upwelling. This strong meridional outflow from the equator is directly related to maximum of the negative ZPG in the eastern box (related to the westward wind stress reversal to eastward in the eastern ETA) at the time. The present work shows quantitatively the relation of the EUC downstream weakening in the eastern part of the ETA to the off-equatorial meridional flow that is predicted by the vertical structure of the ZPG during fall in both observations and model.

From a momentum balance perspective, mid-basin in the ETA at 10ºW, non-linear advection terms are very important and the EUC core seasonality can only be accurately reproduced at 10ºW when non-linear terms are taken into account. During the fall EUC core maximum that occurs on both model and observations, the strong contribution of the non-linear advection term suggests that linear dynamics and simple superposition of basin modes would not accurately reproduce the EUC core velocity seasonal cycle at 10ºW. Additionally, the momentum balance in both model and observations shows an EUC core minimum under non-linear (weakly or strongly) flow regimes and EUC core maxima generally under weakly or strongly non-linear flow regimes as well, further highlighting the importance of non-linear ADV in the Atlantic EUC momentum balance. The importance of the non-linear ADV terms in the momentum balance is assessed quantitatively with observations for the first time in the Atlantic EUC.
Regarding what controls the seasonal cycle of the EUC core depth, the momentum balance shows some dependence to non-linear ADV terms. In the model, the shallowest core depth in spring occurs under a weakly non-linear regime, where the larger ZPG values at shallower depths can increase the zonal momentum of the upper part of the EUC causing the shoaling of the core. This is consistent with the linear dynamics explanation in [Brandt et al., 2016a], where the fourth baroclinic mode can be used to reconstruct the seasonal cycle of the EUC’s core depth. In the observations, during April the residual ZDF appears to be weakly negative (due to the strong westward ADV) and although the ZPG also shows a minimum at that time it is still strong enough at shallower depths to effectively move the core upwards by strengthening the upper EUC. During maximum EUC core deepening in the model (July) the eastward ZPG is deepening showing elevated values at greater depths but the westward ZDF has started to extend deeper 1-2 months prior to the ZPG deepening, causing the most enhanced and effective loss of momentum in the upper EUC that continues through July and along with the strong westward ADV pushes the core deeper (as per [Wacongne, 1989])

The EUC deep extensions at the central ETA (10°W) can be explained by the ZPG seasonal cycle. The observational seasonal momentum balance below the depth of 150m is described by a weakly non-linear flow regime, with the eastward ZPG following the local acceleration and ADV partially offsetting the EUC acceleration and therefore the deep extension in early summer and sustaining the EUC acceleration in late summer. Below again the depth of 150m, the seasonal zonal momentum balance in the model is described by a linear flow regime with the eastward ZPG being almost in phase with the
local acceleration and being primarily responsible for the existence of the EUC deep extensions.

The observational ADV is not a localized estimate and it has proven to be a source of “momentum imbalance”, confirmed by the comparison of zonally-averaged non-linear terms from the model and the non-linear terms derived from model fields using the observational methods. This comparison has proven to be more reasonable for the upper part of the EUC (surface to 100m). Additionally, the observational ADV has shown the largest errors from all the momentum balance tests.

The seasonal direct ZDF that has been attempted in this study has significant qualitative similarities to the model’s ZDF and is qualitatively reasonable, however it significantly overestimated the momentum budget, especially at EUC core depth. Microstructure measurements on PIRATA equatorial mooring array analogous to the ones in the equatorial Pacific TAO array ([Moum et al., 2013]), have been recently added to the array and are expected to provide a better representation mixing on the equatorial Atlantic.

Despite the challenges in estimating upwelling from indirect methods and attempting to close the observational momentum balance, this work provides an updated understanding of the main processes controlling the seasonal cycle of the EUC and upwelling in the ETA. Through the course of this work, three areas where significant improvements should be made have been detected: (i) the zonal pressure gradient estimate, (ii) direct vertical diffusion estimate and (iii) the need for longer duration mooring u, v velocity time series to better determine the climatological seasonal cycle.

The zonal pressure gradient estimate is based on monthly dynamic heights that are derived from the monthly SIO Argo T/S profiles for the period between 2004-2011. The
use of a “level of no motion” at a certain depth to reference the dynamic heights, introduces uncertainties that are extensively discussed in this work. Although the uncertainties related to the choice of the level of no motion are small, the knowledge of a true reference surface would help in eliminating them. Furthermore, the T/S profiles that are measured from floats used to produce the gridded monthly climatology, can include noise or seasonal biases introduced from the interpolation scheme used to obtain the Argo product. Increasing the Argo float density in the near-equatorial region would obviously help to improve the accuracy of the ZPG estimates, as will the accumulation of Argo data going forward. Using T/S profiles from shipboard CTD sections along the equator could provide some "snapshot" estimates of the ZPG, but would not necessarily be representative of seasonal averages or be able to resolve the seasonality of the term. Regardless, the fact that there are ongoing cruises in the equatorial Atlantic area (i.e. to serve the PIRATA mooring array), a shipboard based zonal pressure gradient “data base” can constantly be updated to try to improve the climatological estimates of the ZPG term.

The addition of microstructure instrumentation (χ-pods) to the PIRATA array will eventually yield time series of turbulent kinetic energy dissipation rates ($\varepsilon$). Those measurements, could act as test-bed to derive plausible $k_m$ parameterizations as a function of shear, stratification and $Ri$ and in turn better resolved vertical diffusion estimates. But an even more important contribution could be made in the vertical diffusion schemes used in ocean models. As in the model used in the present study, most ocean models use vertical diffusion schemes in which the $k_m$ is calculated using prognostic equations for $\varepsilon$ and a series of known turbulence closure schemes. To this day, the comparison of ($\varepsilon$) used the models and from direct observations is not common, mainly due to the lack of measurements. In
an area as sensitive to vertical diffusion parameterization schemes as the equator [Blanke and Delecluse, 1993], any improvement in the those schemes by comparing them with observations will result in improvement in the reproduction of the equatorial circulation. In turn, improved circulation can be translated to better representation of the EUC eastward penetration, the EUC core intensity and depth, the strength and seasonality of the upwelling.

During TACE, the almost 4-year time series provided a first ever estimate of the seasonal cycle of the EUC in the central and eastern equatorial Atlantic. However, the zonal and meridional velocity mooring time series suffered from various gaps in their records. For example, at 10ºW on the equator, only ~2.5 years of continuous mooring current measurements are available. This raises a question of how well the seasonality of the circulation at 10ºW is resolved with such a small record. At 23ºW on the equator, there is an ADCP on the PIRATA mooring, but this is the only location in the array where currents are measured and longer time series can be available. Although the basic seasonal variability of the EUC has been reasonably resolved with the available TACE measurements, accurate estimates of terms in the seasonal momentum balance, in particular the ADV term, would benefit from longer-term moored observation to obtain more statistical reliability.

Within the framework of this study, the results from the comparison between observational model mass and momentum balances can contribute to the future modeling studies. Although in the model used in this work the EUC seasonal cycle and magnitude are comparable to the observations, the fact that the EUC in the Gulf Guinea (at 0ºE) is different and weaker in the model suggests there are improvements to be made. The
downstream EUC weakening in the model suggests that there is a significant eastward momentum decrease. Assuming that the model’s vertical diffusion of momentum is still strong in the Gulf of Guinea, while the ZPG weakens downstream, then the net result would be a decrease in EUC strength in the model. This in turn would lead to moderate supply from the EUC to the upwelling transport in the eastern part of the basin. As mentioned before, the improvement of the vertical diffusion of momentum in the ocean general circulation models is responsible for the better representation of the EUC. However, with this study we show that there is still room for improvement, especially in the Gulf of Guinea where the warm SST biases in climate models are of such importance. The vertical diffusion of momentum schemes in the models can be further fine-tuned to reproduce more accurately the downstream EUC weakening. Since, the physics involved in the $k_m$ parameterization schemes are directly linked to adequately resolving turbulence, more simulations using higher resolution regional ocean circulation (i.e. $1/10^\circ$) can be expected yield more realistic EUC in the easternmost part of the basin. This can be also the case for coupled climate models, higher resolution simulation, although computationally expensive, will most likely provide a more realistic EUC.

While the present study utilized as many observational data sets and as possible and a model to answer questions related to seasonal mass and momentum balance, it also showed weaknesses of the various data sets and methodologies. This provides the necessary motivation to keep revisiting the same questions with updated data sets and methods, since there is still a lot to learn.
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A Application of the OW method to estimate westward zonal transports

Widths that range from 0.1° to 1.4° degrees of latitude with a step of 0.1° (14 different widths) are used to estimate transports, using all possible combinations and following equation (2) in [Johns et al., 2014] but for the westward \((u<0)\) velocities. Only the width combinations that show differences between the OW westward transport and section westward transport from 1.2°N-1.2°S that (i) minimize the bias (mean of the differences approaches zero), (ii) minimize the standard deviation, (iii) minimize the root mean square and (iv) have values somewhere between (i), (ii), (iii) (otherwise the mean between (i)-(iii)) are considered as possible candidates. Out of the candidate with combinations the final choice is one that not only reproduces as close as possible the section transport profile, but also shows minimum root mean square error, maximum correlation coefficient and minimizes the absolute difference. The widths chosen in this analysis are summarized in Table 2-2. Fig. A-1(a) shows an example during the Meteor May 2002 cruise along 23°W where the two westward transport estimates agree very well with each other such that their cumulative transport shows almost no difference. For the case of 10°W, for the first part of the TACE experiment the 0.75°N mooring was missing, resulting in the shortest record between the three longitudes and additionally, the 0°N mooring location had a short mooring for the most part of the experiment. The limitations of the mooring record at 10°W and how the OW is applied in the case of the eastward (EUC) transport is discussed in detail in [Johns et al., 2014], and the mooring transports are reconstructed by “merging”
the 2 mooring and 3 mooring records. In Fig. A-1(b) during the CORIOLIS August 2004 cruise), a merged transport profile is used that consists of transport estimated using 3 velocity profiles for the first 100m and 2 velocity profiles for depths > 100m to account for the times that the equatorial mooring was measuring currents only in the upper 100m, and again the OW reproduces quite well the actual section transport. Finally, at 0ºE (Fig. A-1(c), EGEE 1 cruise in June 2005), the transport is reconstructed using 2 velocity profiles that do not capture the westward transport as well, resulting in differences in the cumulative
transport between the section and the estimated from OW as large as 1Sv. As discussed in Section 4(b) 0°E is the case where the reconstruction might introduce biases.

B  Error Analysis

B.1  Errors in the seasonal Optimal Width (OW) derived zonal transports from 50-300m

We have estimated the root mean square (rms) differences for the westward transports in the OW method (Figure 2-6) for all longitudes of interest, while [Johns et al., 2014] and [Brandt et al., 2014] estimated the equivalent EUC rms differences. The measurement error $\sigma_{Zonal, mean}$ in the zonal transports is therefore the square root of the sum of the squares of westward $\sigma_{OW,WESTWARD}$ and EUC $\sigma_{OW,EUC}$ rms differences, expressed by:

$$\sigma_{Zonal, mean} = \sqrt{\sigma_{OW,EUC}^2 + \sigma_{OW,WESTWARD}^2} \quad (B1)$$

The measurement error in the total zonal transports is ±0.6 Sv at 23°W, ±1.0Sv for the merged and ±1.8 Sv for the 2-mooring profile OW reconstruction at 10°W, and ±0.9 Sv at 0°E.

Assuming that the error decorrelation time scale of the OW zonal transports is the same as the decorrelation time scale of the actual transport from 50-300m, we use the integral time scales of the transports at each longitude $\tau_{OW,EUC}$ and $\tau_{OW,WESTWARD}$ to define the degrees of freedom. The integral time scale is estimated as the time step in the time series at which the autocorrelation of the variable of interest is crossing zero. The degrees of freedom $N_{DOF,EUC}$ for the EUC and $N_{DOF,WESTWARD}$ for the westward transports are then given by:

$$N_{DOF,EUC} = \frac{N}{2\tau_{OW,EUC}} \quad (B2a)$$
\[ N_{DOF,WESTWARD} = \frac{N}{2\tau_{OW,WESTWARD}} \] (B2b)

respectively. In equations (B2), \( N \) is the number of days of OW transport values used to estimate the monthly mean in the seasonal cycle of every climatological month. The standard error (ste) \( \sigma_{Zonal}^{STE} \) of the total zonal transport between 30-300m at each longitude is then given as:

\[
\sigma_{Zonal}^{STE} = \sqrt{\left( \frac{\sigma_{STD,EUC}}{\sqrt{N_{DOF,EUC}}} \right)^2 + \left( \frac{\sigma_{STD,WESTWARD}}{\sqrt{N_{DOF,WESTWARD}}} \right)^2} \] (B3)

In (B3) \( \sigma_{STD,EUC} \) and \( \sigma_{STD,WESTWARD} \) are the standard deviations of all the OW EUC and westward transports in each climatological month. The zonal transport convergence is then given by

\[
\sigma_{Zonal,conv}^{STE} = \sqrt{\left( \sigma_{Zonal}^{STE} \right)^2 + \left( \sigma_{Zonal}^{STE} \right)_W^2} \] (B4)

where subscripts W and E are for the western and eastern sides of each box. The blue shaded areas Figure 2-12 and Figure 2-13 show the uncertainties in the zonal transports.

B.2 Errors in the meridional geostrophic transports: reference depth, measurement error, and errors for each climatological month

The deepest mooring depth is 300m and the mooring data available provide no information on what the depth of the “level of no-motion” might be. We test as reference depth the 2000, 1000 and 500 db levels. We then compare monthly DH anomalies (DHA) relative to the time mean between 2007 and 2011, roughly during the whole TACE period to the monthly absolute dynamic topography anomalies (ADTA) from AVISO relative to
the same time period for each reference depth. The mean root mean square (rms) of the monthly ADTA minus the DHA in our area of interest is 1.6 cm for DHAs estimated using 2000 and 1000db and 1.7 cm using 500db. We conclude that when estimating DH near the equator from SIO Argo T/S, reference depths of 500db and below provide a good approximation for the level of no motion and the resulting DHs capture well the sea surface level variability when compared to concurrent AVISO products. In our study, we choose to use 1000db because: (1) it provides very reasonable near surface estimates of the sea level topography when compared to AVISO altimetry products (2) it is the reference depth used in the L99 approach for equatorial geostrophy, the same approach we choose to use in our study.

Following [Johns et al., 2014], we merge the PIRATA with the Argo T/S profiles on the equator at 23ºW, 10ºW and 0ºE and then use the merged profiles to estimate dynamic heights. The differences between the dynamic heights from the merged T/S profiles and the dynamic heights derived from the Argo T/S profiles are considered to represent accuracy estimates of the vertical structure. The differences between the two dynamic height profiles are attributed to the differences between the PIRATA T/S profiles and the Argo T/S profiles. The mean temperature errors are <1ºC and mean salinity errors <0.2psu can result in maximum dynamic height differences of 1.5 cm, value comparable to 1.6 cm we get comparing the surface monthly DHA time series referenced at 1000db to the ADTA AVISO.

The uncertainties in the meridional geostrophic transports can be calculated as monthly profiles, using the differences between the DH from the merged Argo-PIRATA corrected T/S profiles \((DH_{cor}(z,t))\) and the DH derived from the Argo T/S \((DH_{uncor}(z,t))\) profiles:
\[ \sigma_{DH}(z,t) = DH_{cor}(z,t) - DH_{uncor}(z,t) \] (B5),

where \( z \) is the depth and \( t \) is the timestamp of the monthly profile, starting from January 2004 to December 2011. The typical standard error profile is then:

\[ \sigma_{DH}^{STE}(z) = \frac{\sigma_{DH}}{\sqrt{N}} \] (B6),

where \( N \) is the number of months to average. For each monthly DH profile, using 4(b) the meridional transport per unit depth integrated from \( x_w \) (23°W or 10°W) to \( x_e \) (10°W or 0°E) is given by:

\[ V_{tr,g}(z,t) = g \frac{x_w}{f} \int_{x_e}^{x_w} \frac{\partial DH(z,t)}{\partial x} = g \frac{x_w}{f} \left[ DH_{x_w}(z,t) - DH_{x_e}(z,t) \right] \] (B7)

Assuming that the standard error at the three equatorial PIRATA locations is representative for the area within our boxes, then according to (B7) the standard error in the transport profile is

\[ \sigma_{V_{tr,geos}}^{STE}(z) = \frac{g}{f} \sqrt\left( \sigma_{DH,x_w}^{STE}(z) \right)^2 + \left( \sigma_{DH,x_e}^{STE}(z) \right)^2 \] (B8)

Then the total estimated measurement error in the vertically integrated transport between two depths \( z_1 \) and \( z_2 \) (e.g. 30 to 300m) is given by

\[ \sigma_{Q_{geos}}^{STE} = \int_{z_1}^{z_2} \sigma_{V_{tr,geos}}^{STE}(z) dz \] (B9)

Applying (B5) to (B9) yields total measurement error in the meridional geostrophic transport between 50-300m of 0.30 Sv western box and 0.28 Sv for the eastern box.

The measurement error does not include any uncertainly in the transport due to statistical temporal variability. In principal, one can assume that any measurement error that exists will contribute to the observed statistical variability as random geophysical
noise. Since the seasonal monthly climatology is the mean for each climatological month, the standard error for each climatological month is given by the standard deviation of the month divided by the square root of the degrees of freedom. In the case of Argo DH profiles, the degrees of freedom $N_{\text{clim}}$ are estimated from the number of measurements per each month in the 2004 to 2011 Argo monthly time series, therefore ste in the meridional transport seasonal climatology is given by

$$\sigma_{\text{STE, GEOS (mon)}}^\text{STD} = \frac{\sigma_{\text{STD, GEOS (mon)}}}{\sqrt{N_{\text{clim}}}} \quad (B10)$$

For the meridional geostrophic convergence, the standard error is then:

$$\sigma_{\text{STE, conv, GEOS (mon)}} = \sqrt{\left(\sigma_{\text{STE, conv, GEOS (mon)}}^\text{STD} \right)^2 + \left(\sigma_{\text{STE, GEOS (mon)}} \right)^2} \quad (B11)$$

B.3 Errors in the inferred upwelling transport for mass balance

Since the upwelling inferred from the mass balance is the sum of zonal and meridional convergence, the error bars associated with it are given by:

$$\sigma_{\text{STE, Upwelling (mon)}} = \sqrt{\left(\sigma_{\text{Zonal, conv (mon)}}^\text{STD} \right)^2 + \left(\sigma_{\text{STE, conv, GEOS (mon)}} \right)^2} \quad (B12)$$

B.4 Errors in deriving the vertical velocity at 10°W using moorings

We use two types of error bars for the vertical velocities, both shown in Figure 2-19. The first error bar is the standard error of $w$ at each depth derived from the $w$ time series ($\sigma_w$). In the second case, the error bars of $w$ at each depth are derived by integrating the
standard errors of zonal and meridional gradient time series \((\partial u/\partial x \text{ and } \partial v/\partial y)\) from the deepest level to 50m. The error in that case at each depth is given by:
where \( \Delta z = 5m \) and \( DOF_z = 2 \). The reasoning behind using two degrees of freedom is that the feature with the most important vertical variability in this part of the water column are the tropical cells. The TCs have roughly a depth of \(< 100m\) and since here we are interested in the first \( \sim 300m \), we can derive DOF in the vertical as \( DOF_z = \frac{H(= 300m)}{2 \cdot L_{TC}(< 100m)} > 1.5 \) which equals roughly to 2 degrees of freedom.

The standard error at each depth of the time series of interest (\( \sigma_w, \sigma_{\partial w}, \sigma_{\partial v} \)) is calculated by first estimating monthly standard deviation for each month of each year in the time series (i.e. std of January of 2008), divided by the square root of the degrees of freedom at each depth. Similar to equations B2(a,b), the degrees of freedom are defined by the number of samples in each month of a certain year divided by twice the integral time scale of the time series of the variable at that depth. Then the standard error of a climatological month is the mean of the standard errors for this particular month (i.e. Jan), divided by the square root of the number of years in the time series.

B.5 Errors in the upwelling transport from drifters

For the “bin-average” method: in each grid box for each month of the year the mean currents and their standard deviation (std) and standard errors (ste) are estimated based on an integral time scale of 5 days and the number of drifter days per box. For the [Lumpkin and Johnson, 2013] method (LJ13), the error bars in the harmonic decomposition for each grid box are estimated using the square root of the sum of the squares of the formal error
bars that are associated with the coefficients of each one the terms in the harmonic
decomposition. For the full mathematical formulation the reader is referred to the [Lumpkin
and Johnson, 2013] paper and references therein. For each method, each grid point has its
own error associated with the value of the zonal and meridional currents, such that at each
grid point \((i,j)\) (\(i\) is a longitude index and \(j\) a latitude index) has a set values for the zonal
currents and their uncertainties (ste) \(u(i,j)\pm \delta u(i,j)\) and meridional currents and ste
\(v(i,j)\pm \delta v(i,j)\). Fig. B-1 shows examples of the seasonal cycles of zonal and meridional
currents using the two different methods. In all cases of the zonal currents both methods
seem to capture the seasonality quite well, however in the case of the meridional currents,
the cluster of points does not show a significant seasonality, the data is far too noisy to
resolve the seasonal variability of the meridional currents. This is reflected in the
differences in the seasonal cycles of the meridional currents extracted using the two
different methods.

The zonal transport uncertainties using the drifters are given by:

\[
\delta U = \Delta y \cdot \sqrt{\left( \sum_j \delta u(x_w,j) \right)^2 + \left( \sum_j \delta u(x_e,j) \right)^2}
\] (B14),

where the indexes \(W\) and \(E\) refer to the longitudes at the west and east sides of the box.
Similarly, the meridional transport uncertainties are given by:

\[
\delta V = \sqrt{\left( \sum_i \delta v(i,y_N) \Delta x_i \right)^2 + \left( \sum_i \delta v(i,y_S) \Delta x_i \right)^2}
\] (B15),

where this time the \(N\) and \(S\) indexes refer to the north and south sides of each box. As
mentioned in section 5.4, the grid has an optimized zonal resolution, therefore each grid
box needs to be multiplied by the width of the respective box with longitude index \(i\). The
uncertainties in the drifter derived upwelling then are given by

\[ \delta W = \Delta z \cdot \sqrt{\delta^2 U^2 + \delta^2 V^2} \quad (B16) \]

C Testing different temporal resolutions for estimating the seasonal cycle of the local acceleration at 10°W

Seasonal cycles of local acceleration are estimated using two different temporal resolutions of \( u \) profiles time series: a high temporal resolution (twice daily for observations or twice daily for the model) and a low temporal resolution (monthly averaged time series of \( u \) for both model and observations). The seasonal cycles of the local

![Fig. C-1: Seasonal cycle of the local acceleration term using the observations (left) and the model (right) and estimated from twice-daily time series (a, left) and daily time series of \( u \) (a, right) and from monthly time series (b).](image)
acceleration terms estimated from high temporal resolution time series are noisier (Fig. C-1 (a)) than those estimated from low temporal resolution time series (Fig. C-1(b)).

There is a larger difference between the observations’ high resolution case to the low resolution case, than between the model’s high/low cases, which has to be related to the fact that in the model’s high resolution case, the $\Delta t \sim 1 \text{ day}$ while in the observations $\Delta t \sim \frac{1}{2} \text{ day}$. Estimating the local acceleration using a $\Delta t$ of half a day could potentially introduce a bias related to the deep cycle turbulence which is a diurnal phenomenon [Gregg et al., 1985; Smyth and Moum, 2013].

Fig. C-2: Seasonal cycle of the standard error (ste) in the local acceleration term using the observations (left) and the model (right) and estimated from twice-daily time series (a, left) and daily time series of $u$ (a, right) and from monthly time series (b).
For the case of 10°W, the standard errors in the seasonal cycle from the high-resolution time series are at least as large, or even exceed the magnitude of the local acceleration term in both observations and model (Fig. C-2(a)), suggesting that the seasonal cycle derived from the high temporal resolution profiles hardly differs from the statistical variability. It is therefore more meaningful to use the monthly averaged profiles of $u$ to estimate the seasonal cycle of local acceleration since the ste associated with the low resolution case show errors that much smaller than the amplitude of the seasonal cycle’s signal (Fig. C-2(b)), again in both model and observations.

D Estimating stratification frequency from density

In order to estimate the stratification frequency (Brunt-Väisälä frequency) $N^2$ from the hourly resolution PIRATA data, we first estimate high resolution potential density profiles $\rho$ and apply $N^2 = -\frac{1}{\rho_0} \frac{\partial \rho}{\partial z}$ (Eq. D-1). The vertical resolution of the PIRATA salinity records at all equatorial mooring locations is far coarser than the temperature. In addition, data are available only for the first 120m of the water column (500m for temperature).

![10°W ARGO–PIRATA Corrected Temperature vs Density and cubic fit curve](image)

Fig. D-1: Potential Density derived from the PIRATA corrected SIO-Argo monthly profiles of T/S versus T. The scatter plot includes all points from all depths and months.
Using only the hourly PIRATA T/S profiles to obtain density profiles would not adequately resolve all of the depth ranges that is of interest in this study. Instead, the hourly temperature data are used to estimate density profiles by fitting to the temperature profiles a cubic polynomial with coefficients that are calculated from the monthly merged PIRATA- SIO Argo T/S profiles. The merging process of these data sets is analyzed in more detail in [Johns et al., 2014], which covers the period between 2004-2011 [Roemmich and Gilson, 2009], providing 96 profiles of T/S and density that fully incorporate the concurrent PIRATA measurements.

The first step to create the hourly density profiles is to assess the type of fit that would be used. Fig. D-1 shows a scatter plot of all monthly \((T, \rho)\) points at all depths for the merged PIRATA-Argo data set. It is very clear from Fig. D-1 that density is a 3\(^{rd}\) degree polynomial function of temperature, and performing the same type of assessment on a month-of-the-
year (climatological) basis further shows that a cubic fit is appropriate (Fig. D-2). The same type of assessment was performed using polynomials up to 6th degree (not shown) but no significant improvement was found, therefore the 3rd degree was chosen for the fit. The next step is to calculate monthly coefficients for the fit such that \( \rho = aT^3 + bT^2 + cT + d \) (Eq. D-2) where \( a, b, c, d, \rho \) and \( T \) are all functions of depth \( (z) \) and month \( (t) \). The time-series of profiles of coefficients that result from that previous step are used in Eq. D-2 with the equivalent \( T \) profiles. The resulting fitted density profiles are plotted against the actual density profiles to show how well the method works (Fig. D-3) and they are separated in months of the year to highlight seasonal changes. Indeed, no evidence is found that this

Fig. D-3: Reconstruction of monthly potential density profiles for each year from temperature and the cubic fit coefficients.
method - obtaining densities from temperature profiles only - is failing at certain times of the year, however small differences were found at the extrema of the profiles, but in all cases the differences are less than 10%.

The final step is to use the monthly profiles of cubic fit coefficients to each fit for the respective month (i.e March of 2008) with the hourly temperature profiles and Eq. D-2 to get the hourly density profiles and then from Eq. D-2 the stratification frequency $N^2$.

E Richardson number ($R_i$) based parameterizations of the vertical coefficient of the eddy diffusivity of momentum ($k_m$)

As shown in Figure 16, there is a qualitative agreement between the model’s and observation's $R_i$, for the upper part of the EUC. This conclusion is mainly based on the similarity between model and observations percentage of $R_i<0.5$ seasonal cycle (Figure 13(g) and (h). In this work, the $R_i<0.5$ was chosen because: while it is common to use the criterion of $R_i < 0.25$ as a threshold for mixing events [Miles, 1961], this critical Richardson number is a “necessary but not sufficient criterion” for the existence of mixing [Polzin, 1996]. [Polzin, 1996] using observations of turbulent kinetic energy dissipation rates ($\varepsilon$), shear and stratification, showed that elevated dissipation rates $\varepsilon$ occur even at $R_i$ larger than 0.25. In his study, [Polzin, 1996] reported that 63% of the total $\varepsilon$ is associated with $R_i < 0.5$ but only 28% of $\varepsilon$ is associated with $R_i < 0.25$, when shear, stratification and $R_i$ are all evaluated using $\Delta z=4m$ for the calculation of the vertical gradients. The percentage of total $\varepsilon$ associated with $R_i < 0.5$ and $R_i<0.25$ changes to 74% and 43% respectively when $\Delta z=2m$, a fact that brings to the attention the sensitivity of the calculation of shear, stratification and therefore $R_i$, to the vertical resolution. Also following the work of [Polzin, 1996],
Dengler and Quadfasel, 2002, used $Ri_c=0.33$ as a critical number in their study of abyssal mixing within the equatorial deep jets in the Indian Ocean.

E.1 Summary of available mixing parameterizations

$Ri$ based parameterizations for the vertical eddy diffusivity coefficient of momentum suggested by [Pacanowski and Philander, 1981], [Peters et al., 1988], [Kunze et al., 1990], [Large et al., 1994], [Zaron and Moum, 2009] and [Hummels, 2012] are all tested first within the model's framework. Ultimately a combination of the [Hummels, 2012; Hummels et al., 2014] and [Pacanowski and Philander, 1981] parameterizations are found to produce the most realistic results.

The parameterization described in [Pacanowski and Philander, 1981] (hereafter PP81) is given by the functional form: $k_{m(PP81)} = \frac{k_{m_0}}{(1 + \alpha Ri)^n} + k_{m_b}$ (Eq. E-1), where the only variable is $Ri$ and the rest of the parameters are constant with values $\alpha=5$, $n=2$, $k_{m_b} = 10^{-4} \text{ m}^2/\text{s}$ and $k_{m_0} = 5 \cdot 10^{-3} \text{ m}^2/\text{s}$. The PP81 model was used in the modeling study of [Wacongne, 1989] for the equatorial Atlantic, one of the few momentum balance studies of the Atlantic EUC.

[Peters et al., 1988] analyzed direct microstructure measurements in the equatorial Pacific and proposed two parameterizations: one relative to high shear levels (hereafter called PGT88-high) and one more applicable to lower shear levels (hereafter called PGT88-low). The functional form of the high shear regime, mainly related to the upper EUC layers, is given by $k_m = \alpha R_i^b$ where the best determined values for the constants according to microstructure observations take values, such as $k_{m(PGT88-high)} = 5.6 \cdot 10^{-8} R_i^{-8.2}$ (Eq. E-2a). For the low shear regime, associated with the lower EUC layers, where $Ri$ is
higher, the PGT88-low parameterizations has a functional form similar to PP81:

\[
k_{m(\text{PGT88-low})} = \frac{5 \cdot 10^{-4}}{(1 + 5R_i)^{1.5}} + 2 \cdot 10^{-5} \quad (\text{Eq. E-2b}).
\]

The parameterization described by [Kunze et al., 1990] (hereafter KWB90), was very thoroughly tested by [Polzin, 1996] and was also applied by [Dengler and Quadfasel, 2002] to diagnose abyssal mixing within the equatorial deep jets in the Indian Ocean. According to KWB90, \( \varepsilon \) (dissipation rate of TKE) is expressed through

\[
\langle \varepsilon_{\text{KWB90}} \rangle = fr \cdot \Delta z^2 \left( \frac{S^2 - 4N^2}{24} \right) \left( \frac{S^2 - 2N^2}{4} \right) \quad (\text{Eq. E-3a}).
\]

KWB90 has the disadvantage of being valid only at times when unstable events can occur. Unstable events are ones that occur at \( Ri < Ri_c \), and for the purposes of this study \( Ri < 0.5 \). In Eq. D3-a \( fr \) is the percentage of the water column that is unstable (contains values of \( Ri < 0.5 \)) and \( \Delta z \) is the vertical scale over which the shear \((S^2)\) and stratification \((N^2)\) are calculated. Here, \( \Delta z \) is a function of depth while \( fr \) has one value for each profile of \( N^2 \), \( S^2 \) and \( Ri \). Despite the limitations of KWB90, the large percentage of \( Ri < 0.5 \) in the upper part of the EUC implies that KWB90 has the potential to provide a valuable parameterization at least for those depths. For the estimate of \( k_m \) from Eq. E-3a, \( k_m = \frac{(\Gamma + 1)\varepsilon}{S^2} \) (Eq. E-3b) [Osborn, 1980; Peters et al., 1988] is used with mixing ratio \( \Gamma = 0.2 \) to yield:

\[
k_{m(\text{KWB90})} = fr \cdot \Delta z^2 \left( \frac{S^2 - 4N^2}{24} \right) \left( \frac{S^2 - 2N^2}{4} \right) \frac{(\Gamma + 1)}{S^2} \quad (\text{Eq. E-3c}).
\]

One of the most well-known parameterizations for \( k_m \) is the one proposed by [Large et al., 1994], widely known as “K-profile” or “KPP” (hereafter KPP94). KPP94 has been used exclusively or as a module in many Ocean General Circulation Model (OGCMs),
including the model used in the present study (NEMO). Below the boundary layer near the
ocean’s surface (called surface boundary layer, or SBL by [Zaron and Moum, 2009]), the
KPP94 can be applied in order to provide a $k_m$ for the ocean’s interior. According to
KPP94, the appropriate parameterization for $k_m$ when mixing is related to shear instability
is given by: $$\frac{k_{m(KPP94)}^s}{k_0} = \begin{cases} 1, & Ri < 0 \\ \left[1 - \left(\frac{Ri}{Ri_0}\right)^2\right]^{3/2}, & 0 \leq Ri \leq Ri_0 = 0.8 \end{cases}$$ (Eq. E-4a), while when the
mixing is related to internal waves the $k_m$ is given by $k_{m(KPP94)}^w = 10^{-4} \text{ m}^2 / \text{s}$. In KPP94,
$k_0 = 40 \cdot 10^{-4} \text{ m}^2 / \text{s}$ after [Peters et al., 1988] however the $Ri_0 = 0.8$, $k_{m(KPP94)}^w$ and $k_0$ are
the same values used in [Large and Gent, 1999; Zaron and Moum, 2009]. To describe the
total $k_m$ in KPP94 due to the different types of mixing the sum of the different $k_m$’s is used:
$$k_{m(KPP94)} = k_{m(KPP94)}^s + k_{m(KPP94)}^w$$ (Eq. E-4b). Note that in the original KPP94 study [Large
et al., 1994], mixing due to double diffusion is also accounted for. For the purposes of this
study (large spatiotemporal scale and open ocean conditions) the effects of double diffusion
(e.g. salt fingers) are not taken into consideration.

[Zaron and Moum, 2009] proposed another Ri-based scheme for mixing in the
Equatorial ocean. In their observational study, they also evaluated KPP94 to highlight the
different implications that this very widely used parameterization could have. The [Zaron
and Moum, 2009] parameterization (hereafter ZM09) has the form of
$$k_{m(ZM09)} = k_0 \phi_m^z(Ri)$$ (Eq. D-5a), where $\kappa_0 = \frac{\sqrt{u^2 + v^2}}{S}$ (Eq. E-5b).
They proposed two functional forms: \( \phi_m^{alt} = a \left( \frac{Ri_1}{Ri - Ri_1} \right) + be^{-\beta Ri} + c \) (Eq. E-5c) and
\[
\phi_m^{rev} = \begin{cases} 
\phi_m^{max}, & Ri \leq Ri_1 \\
\Delta \phi_m e^{-\gamma(Ri - Ri_1)} + \phi_m^w, & otherwise 
\end{cases}
\] (Eq. E-5d). The constants used here are:
\[
a = 8 \cdot 10^{-7}, \quad b = 3 \cdot 10^{-4}, \quad c = 2 \cdot 10^{-7}, \quad \alpha = 5, \quad \beta = -4, \quad Ri_1 = 0.25, \quad \phi_m^{max} = 1.2 \cdot 10^{-3},
\]
\[
\Delta \phi_m = 1.2 \cdot 10^{-4}, \quad \phi_m^w = 2 \cdot 10^{-6}, \quad \gamma = 9.61 \text{ and } Ri_2 = 0.168. \]

The main difference between the ZM09-alt and ZM09-rev is that the first one “parameterizes the Ri dependence of the diffusivity”, while the second one the “parameterizes the Ri dependence of the vertical flux”.

According to [Hummels, 2012; Hummels et al., 2014] (hereafter called H12), based on direct microstructure observations in the ETA, the most plausible estimates of \( k_h \) are produced when the microstructure measurements of turbulent kinetic energy dissipation rate (\( \varepsilon \)) are fit to \( Ri \) using a functional dependence of the form \( \varepsilon = aRi^b \), where \( a = 4 \cdot 10^{-8} \) and \( b = -1.2 \). Since the measured quantity in microstructure measurements is the turbulent kinetic energy dissipation rate (\( \varepsilon \)) [Osborn, 1980], Eq. E-3b) from [Peters et al., 1988] can be used to estimate \( k_m \) from \( \varepsilon \), where again \( \Gamma \) is the mixing efficiency [Osborn, 1980]. In the equatorial Atlantic the value of \( \Gamma = 0.2 \) is commonly used in observational studies as well as in [Hummels, 2012].

E.2 Testing and application of mixing parametrizations to the TACE observations

Applying each of the parameterizations described above, hourly time series profiles of \( k_m \) at 10\(^\circ\)W at the equator are estimated. Those time series are further smoothed by omitting
values that fall outside the range of $10^{-6} \text{ m}^2/\text{s} \leq k_m \leq 10^{-2} \text{ m}^2/\text{s}$: any value outside that range is considered an outlier. If within a profile more than 35% of the values are outliers, the profile is omitted altogether and not taken under further consideration. After discarding the outlier values the remaining $k_m$ profiles are averaged to monthly climatological values and the results are presented in Fig. E-1.

As a first test of these parameterizations, they are systematically compared to the effective $k_m$ values that are produced from the turbulence closure within the model. To estimate the seasonal climatology of the model’s $k_m$ (Fig. E-1(a)), the daily profiles of the
ZDF term and vertical shear $\frac{\partial u}{\partial z}$ are used such that:

$$k_m(z)_{\text{(mod el)}} = \frac{\tau_0 - \int ZDF dz}{z \frac{\partial u}{\partial z}(z)}.$$  

In Figure E-1, there is no parameterization of $k_m$ that captures the best the model’s $k_m$ seasonality and vertical structure, but there are elements in some parameterizations that agree with $k_m(z)_{\text{(mod el)}}$. At a first glance PP81 (Fig. E-1 (b)) and KPP94 (Fig. E-1 (e)) seem to have the most similar structures to the model. However, PP81 has some very small values close to the surface during fall, where the model’s $k_m$ exhibits a clear maximum. Also, KPP94 has stronger $k_m$ values below the EUC core (black dashed line) and shows little seasonality there. The ZM09 (alt and rev) parameterizations show results that are very different from the model, and both are considered implausible and not used further. ZM09-alt (Fig. E-1 (f)) severely over-estimates the upper part of the $k_m$ and underestimates the lower part, while ZM09-rev (Fig. E-1 (g)) generally underestimates $k_m$ at almost all depths. PGT88-low (Fig. E-1 (c)) underestimates $k_m$ at all times, while the PGT88-high is not shown since it produced excessively large values that were considered outliers. KWB90 (Fig. E-1(d)) produces values only above the core where $Ri<0.5$, that are very strong compared to the model, and have a seasonality that is very different from the model. H12 (Fig. E-1(h)) is the most similar to the model’s structure from the surface to the EUC core, but shows weaker $k_m$ values near the surface. Below the core H12 performs poorly.

The analysis of Richardson number ($Ri$) based parameterizations of $k_m$ and their comparisons with the model’s $k_m$ gives a clear message: there is no single parameterization that seems capable of yielding plausible $k_m$ values across all depths and
seasons. Rather, a hybrid $k_m = f(R_i)$ parameterization that combines features of the most reasonable parameterizations above is considered as the most appropriate choice for application to the observations. Among those parameterizations, the H12 parameterization - based on actual observations collected in the ETA, appears to work best in the upper water column, from the surface to the core of the EUC, while at greater depths the PP81 parameterization is overall most reasonable. Therefore, a “hybrid” parameterization using the H12 from the surface to the EUC core and PP81 from the core downward is used in the observations in order to obtain a qualitative direct estimate of the ZDF term.

E.3 Construction of the “hybrid” $k_m$ profiles and direct vertical stress estimate

The hybrid $k_m$ profiles for each time step (hourly profiles from hourly shear and stratification) are constructed following the next steps:

1) Locate the EUC core for each profile. For this step, the core is defined as the depth at which the eastward flowing current below 30m becomes maximum.

2) Apply the H12 from the surface (first resolved depth level) to the core depth.

3) Apply the PP81 from the core depth to ~300m (or the deepest resolved depth).

4) Smooth the part of the resolved profile with a 15m running average. Note that the vertical resolution of the unsmoothed profiles is 5m, therefore the 15m smoothing essentially is a 3-point moving average smoothing.

The direct estimate of the vertical stress is then calculated by first multiplying the $k_m$ profile of each time step with the shear of the zonal current $\frac{\partial U}{\partial x}$, then using the CCMP wind stress at the surface Eq. 3-5 is applied. Note that the CCMP wind stress has 6hr
resolution and to use it as a boundary condition to the hourly stress profiles, it is linearly interpolated to hourly values. Since the seasonal climatology of the time series of vertical stress profiles is of interest, this interpolation to a finer temporal resolution is not introducing biases to the result.