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Annual to Interannual Barotropic Variability in the Atlantic Western Boundary

Alvaro Montengro
THE FLORIDA STATE UNIVERSITY
COLLEGE OF ARTS AND SCIENCES

ANNUAL TO INTERANNUAL BAROTROPIC VARIABILITY IN THE ATLANTIC WESTERN BOUNDARY

By
ÁLVARO MONTENEGRO

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The members of the Committee approve the dissertation of Álvaro Montenegro defended on November 7, 2003.

Georges L. Weatherly
Professor Directing Thesis

Steven L. Blumsack
Outside Committee Member

William M. Landing
Committee Member

Doron Nof
Committee Member

James O’Brien
Committee Member

Kevin Speer
Committee Member

The Office of Graduate Studies has verified and approved the above named committee members.
To Eunice and Érico, meus amores!
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ABSTRACT

A method for estimating deep, depth independent current variability is described. The procedure uses XBT derived dynamic heights to remove the near surface signal from altimetric sea surface height (SSH). The difference between SSH and dynamic height is operationally defined as barotropic height (BRT). Currents are obtained from BRT slopes using the geostrophic balance. The method requires the variability below the deepest XBT measurement to be small. Results are restricted to temporal variability, as geoid uncertainties in the SSH data render absolute current estimations impossible. The technique is originally developed for use in the Atlantic Western Boundary Current (WBC). Comprehensive verification of the methodology requires simultaneous SSH, XBT and current meter measurements. There are no available Atlantic data sets that meet these requisites. The alternative is to use synthetic data from the CLIPPER Atlantic model (1/6° resolution). Correlation ($r$) between estimated and modelled near bottom velocities in WBC areas of interest vary from 0.7 to 0.8. Further tests are conducted with observed data from the Shikoku Basin boundary current in southern Japan, where the method is capable of reproducing the directly measured near bottom current variability ($r = 0.6$). The procedure is also tested using north Pacific ($5° - 55°$N) data from the OCCAM model. Correlations between model derived and BRT estimated velocities are around $r = 0.7$ for the Shikoku Basin northern boundary and for the Pacific WBC. Values reach $r = 0.9$ in large areas of the basin’s interior, specially over smooth topography. The above method is used to generate time series of the barotropic variability in two areas of the Atlantic Western Boundary. One site is located at 38°N, inshore of the Gulf Stream. The other is at 8°S, off the Brazilian coast. Both series are approximately 6 years long. Empirical Orthogonal Function analysis results conducted on SSH and sea surface temperature data are used to confirm the feasibility of applying the method in the chosen South Atlantic area. Currents are compared to scatterometer derived local along-shore wind stress and basin wide wind stress curl. In both areas, current variability is significantly correlated to basin averaged...
wind stress curl and also to local along-shore wind stress. The relationship between currents and wind curl is coherent with the WBC response to interior Sverdrup flow. We propose that local wind stress exerts control over the flow by divergence of the Ekman flow at the coast. In the north, the variability is dominated by interannual oscillations of the wind curl. The effects of the local stress are secondary and have annual frequency. Both wind stress curl and along-shore wind are significantly correlated to the currents on the southern site, but the local effect appears to be the dominant forcing. The main observed results are confirmed by data from a numerical model with 1/6° horizontal resolution.
CHAPTER 1

INTRODUCTION

The swift and conspicuous flows along western boundaries were among the first large scale ocean currents to be thoroughly described. Accurate charts of the path of the Gulf Stream, for example, were available by the end of the 18th century. Due mainly to their importance to shipping, descriptive knowledge of the Western Boundary Currents (WBC) kept growing during the 19th and the first half of the 20th centuries [1].

The first theoretical model of the westward intensification was proposed by Stommel in the late forties [2]. In this elegant wind forced model, the presence of a mean WBC is caused by the latitudinal variation of Coriolis effect. During the 1950’s and 1960’s, further models by Munk, Veronis and others ([2], [3]) added to the basic physics present in Stommel’s solution, refining our knowledge of the steady state WBC.

These time mean models are instrumental for the present understanding of the WBC. Still, by the mid 1970’s, as data from better and more consistent sampling accumulated, it became clear that the ocean system is essentially turbulent, and that the time dependent portion of the circulation is very important [4].

In the WBC, variability occurs in a wide range of time scales, from seconds to periods that most certainly exceeds the span of the few long term observations available (for some examples: [5], [6], [7], [8]). In the annual to interannual range, the frequency interval pertinent to our objectives, several mechanisms have been associated in the literature with WBC change.

Moore and Wilkin [9], analyzing current meter data from the South Pacific, attribute a 150 day current oscillation to reflected Rossby waves. Sturges and Hong [5] use tide station sea surface elevation to estimate the transport variability of the Gulf Stream over a 40 year period. They see a correlation between Rossby waves forced by wind stress curl in the interior and decadal variability in the western boundary. Fischer and Schott [8], based on current
meter observations in the Equatorial Atlantic, propose that annual to semiannual changes in the flow are related equatorial waves. A correlation between WBC and interior Sverdrup transport is present in current meter data analyzed by Vivier et al. [10] (South Atlantic) and Lee et al. [11] (long term array east of the Bahamas in the North Atlantic). An interannual correlation between these last two parameters is observed by Kutsuwada [12] off the south coast of Japan.

The results described in the last paragraph are in accordance with the theoretical framework proposed by Willebrand et al. [6] and later expanded by Pierini [13]. These authors use linear wave theory to present a comprehensive analytical and numerical analysis of the oceanic response to large scale wind and pressure forcing. According to their finds, stochastic wind forcing with periods between 1 and 300 days generate a predominantly barotropic oceanic response. For periods below 30 days, the wind generates rapidly propagating Rossby waves. Between 30 and 300 days, the response is in the form of a wind stress curl forced, Sverdrup like, interior flow compensated by a return in the WBC. Forcing with periods above 300 days permits the establishment large scale, baroclinic Rossby waves.

Another set of observational and modelling studies point out an annual relationship between the local along-shore wind stress and WBC flows. This is the case of the current meter analyses conducted by Lee et al. [14] (North Pacific) and Lee and Williams (North Atlantic). This local stress-WBC correlation is also present in numerical results from North Pacific simulations by Kubota et al. [15] and Greatbatch and Goulding ([16], [17]).

The theoretical framework better suited to explain this local along-shore wind stress influence over WBC oscillations seems to be the shelf dynamic mechanisms proposed by Csanady [18]. In this analytical model, the along-shore wind affects the along-shore currents via convergence and divergence of the Ekman transport at the coastal boundary. Lee et al. and Lee and Williams ([14], [19]) use Csanady’s model to explain WBC variability with periods of a few days. Their conclusion is that the annual correlation between currents and along-shore winds are caused by seasonal modulation of storm activity. This is not the case of the numerical results ([16], [17], [15]), where the models are forced by monthly wind fields.

A third mode of WBC variability in our scale of interest is proposed by Choboter and Swaters [20]. These authors, based on model results, relate an annual signal in the Antarctic
Bottom Water (AABW) component of the Deep Western Boundary Current (DWBC) at the equatorial region to varying rates of this water mass formation. Note that, different from the previous two mechanisms, this process is related to the thermohaline, and not the wind forced circulation.

The basic model of the oceanic thermohaline circulation was proposed by Stommel and Arons [21]. Based on analytical and laboratory results, they conclude that the density driven circulation could be explained by a series of localized deep water sources compensated by space homogeneous upwelling in the thermocline. The sinking water would tend to flow along the bathymetry, forming deep and/or near bottom intensified currents along the western boundaries. A number of observational studies have since confirmed the presence of these DWBCs (for some examples: [22], [23], [24], [25], [26], [27], [11]).

Deep surface convection is primarily confined to relatively small areas in the Labrador Sea and around Greenland as well as in the Antarctic shelf. As would be expected, many of the first DWBC descriptions were based on North Atlantic observations. Also from the North Atlantic came the first indications that these deep flows play an important role in the ocean heat budget ([28], [29]).

More than that, as has been proposed by Gordon [30], the Atlantic DWBC is an important segment of the World Ocean thermohaline circulation. This author puts forward a conceptual model of the ocean’s overturning circulation (the “ocean conveyor belt”) calling attention to its importance in climate and climate change.

The actual global thermohaline flow patterns are probably much more intricate than the simplified version advanced by Gordon ([31], [32]). Nevertheless, there are a many indications of a connection between climate, climate change and the Atlantic DWBC. Some come from recent estimates of the oceanic heat budget, that confirm the important role played by this current in the global heat transport, both for the steady state ([33], [34], [35]) and the seasonal variability [36].

More evidence of the climatic relevance of the Atlantic DWBC comes from paleoceanography studies, which indicate that the characteristics of the Atlantic DWBC varied significantly as Earth’s climate changed ([37], [38], [39], [40]). These inferences must be taken carefully though. LeGrand and Wunsch [41] use inversion modelling techniques to demonstrate that
the distribution of oceanic paleotracers used as proof of past DWBC change could also be obtained by present day circulation patterns.

The possibility of anthropogenic induced global warming and findings indicating large climatological fluctuations with decadal time scales has made climate change one of the central themes in present geophysical research ([42], [43]). The clear connections between oceanic heat transport, climate and WBC attest to the current relevance of WBC variability studies.

Apart from its climatic pertinence, WBC variability has also been associated with changes in nutrient transport and, consequently, in local to regional changes in primary production ([44], [45]). By directly influencing larval transport and recruitment, changes in these currents are also important in controlling the availability marine organisms of commercial significance, like squid off Argentina ([46].

It is clear that WBC and their variability are phenomena that merit scientific attention. Their importance is not restricted to the light they might shed on our knowledge of ocean dynamics. These flows also play a role in a number of processes of direct and immediate societal concern.

While much progress has been achieved in the description and explanation of WBC variability, many questions remain. As very common in oceanography, a important factor hindering present knowledge is the lack of real world data.

The length and number of direct WBC observations is severely limited by technical and financial constraints. Most Eulerian measurements span periods of a few months to no more than 2 years. Such records have no information about the interannual variation of the currents. Most of them, in fact, are too short to capture even the annual signal, limiting their usefulness in climatological studies for example. According to Roemmich et al. [47], long term monitoring of the WBC flows is one of the missing components on the present climate observation system and should be a priority for the post-WOCE (World Ocean Circulation Experiment) sampling efforts.

The dissertation has as objective observe, describe and provide a physical interpretation of the barotropic variability in the Atlantic Western Boundary Current (WBC), especially in the annual to interannual period range.
The choice of barotropic flow stems from three main reasons. First, there are many relevant unanswered scientific questions about the role of the winds in WBC variability. Second, these depth independent currents can be estimated using a method based on the abundantly available satellite altimetric data and Expendable Bathythermographs (XBT) temperature profiles (XBT samples, while not as prevalent as the altimetric information, are much more common than direct velocity measurements). Finally, if WBC flows are poorly sampled, DWBC ones are even more. A good portion of the deep ocean variability is of barotropic nature. By concentrating on this member, the work is in fact adding to the understanding of the most data starved component of the western boundary flows.

The longer time series derived through the methodology will permit the determination of the relative importance of basin wide wind stress curl versus local along-shore wind stress forcing in the annual to interannual DWBC variability. In most (among the few) observational studies able to resolve these periods, only the wind stress curl action is analyzed ([10], [22], [12]). In the work by Lee et al. [11] both effects are associated with DWBC annual variability, but no effort is made to evaluate which is dominant. There are no references about the possible interannual action of the local wind stress.

It is important to note that, while the density driven flow is the western boundary component of greater relevance to steady state heat transport analysis, recent results by Jayne and Marotzke [36] point out that seasonal variability of this parameter is essentially controlled by the wind and is of a strong barotropic nature. This to stress that, within the time scales presently inspected, barotropic flows can play a pertinent role in the oceanic heat budget. How much of a role and by which processes are questions still not fully answered.

While the thermohaline component is not present in the results presented here, these flows are embedded in the barotropic currents analyzed. In that way, by setting amplitudes and phases of the depth independent variability (and by determining which type of wind forcing is generating these changes) our results can aid in the partitioning between density driven and wind driven flows from a given deep current measurement or estimate. For example, how much of the annual signal studied by Choboter and Swaters [20] is, as assumed by the authors, related to seasonal changes in the rate of AABW production and how much of it can, in fact, be attributed to wind forcing (which they neglected)? While this example deals
with yearly oscillations, there is no reason to believe the same type of decomposition could not be applied to longer term variability.

This study is divided in two parts. The first one presents and tests a procedure that estimates barotropic variability in the WBC. In the second part, the methodology is applied to specific areas of the Atlantic.

As can be deduced from the section above, any observational study dealing with WBC variability in periods longer than a few months must first overcome the scarcity and short duration of available records. This is done by application of a method that uses sea surface height (SSH) from satellite altimetric measurements and XBT temperature profiles to estimate barotropic currents.

The basic concept behind what we name the barotropic height method is to use XBT derived dynamic heights to remove the baroclinic, near surface signal from the altimetric data. The assumption is that whatever is left afterwards is related to the barotropic component of the circulation (hence, barotropic height, or BRT).

BRT values are converted to currents by assuming the slope between two BRT estimates are in geostrophic balance. Due to the uncertainties of the geoid, the barotropic height method can provide flow variability but not absolute current values. Given the large number of both XBT and SSH measurements, this approach can generate WBC time series much longer than most data available today. It would also offer a cost efficient alternative for any future monitoring program.

The method is being used for the first time in such analysis, consequently, prior to its application in the Atlantic WBC, it is adapted and tested using data from the Shikoku Basin near Japan and also from two numerical models. Results from this initial phase are reported in chapter 2.

After the method is shown to be capable of capturing the barotropic variability of observed and modelled data, it is applied to two regions of the Atlantic WBC. One centered around 8°S and the other around 38°N. These were selected based on their good XBT temporal and spatial coverage.

Two time series, each approximately 6 years in duration, are generated. Both series have irregular temporal resolution, determined by the XBT availability. The southern series is composed of 30 values, with an averaged of one measurement every 2.5 months. In the
north site, the 82 current estimates are nearly one month apart from each other. The time series obtained are correlated to the wind forcing appointed by the literature as important factors in WBC variability (basin wind stress curl and local along-shore wind stress). The relationship between winds and WBC barotropic flows are also explored using data from the CLIPPER numerical model. This part of the project is dealt with in chapter 3.

Please note that chapters 2 and 3 are actually two independent, thematically linked papers ready for submission to a scientific journal, hence the brevity of the Conclusions (chapter 4). Also, there exists some redundancy. More specifically, section 3.4 describes the method in much the same way done by section 2.2. As section 2.2 is more comprehensive, the reader familiar with it could skip section 3.4.
CHAPTER 2
THE BAROTROPIC HEIGHT METHOD

2.1 Introduction

Western boundary currents (WBC) are relevant to many oceanic and climatological phenomena of interest. These flows play an important role in present day ocean heat, salt and freshwater transports ([33], [48], [34]). Paleoceanography studies indicate that changes in past climate states were associated with changes in WBC characteristics ([37]). They can also influence nutrient and primary production variability on regional and synoptic scales ([49], [45], [44], [46]).

For all their importance, most WBC direct Eulerian measurements span periods of a few months to 2 years. Records are scarcer and tend to be shorter in the case of Deep West Boundary Currents (DWBC) (for example, [11], [50], [25], [51], [22]). This confines their application to the understanding of variability at periods near or shorter than one year, a serious limitation in the study of, for example, many important climate phenomena. According to Roemmich [47], long term monitoring of these flows is one of the missing components on the present climate observation system and should be a priority for the post-WOCE sampling efforts.

Here we describe and test a technique that would greatly simplify observations of the barotropic variability in DWBCs. Based on satellite altimetric measurements and Expendable Bathythermographs (XBT) acquired temperature profiles, this method would permit the establishment of time series much longer than most data available today. It would also offer a cost efficient alternative for any future monitoring program.

Practical and theoretical motivations for developing the procedure are discussed in section 1.1. The method, its limitations, as well as some of its applications are dealt with in section
2. Tests, where results from the proposed technique were compared to observed and synthetic data, are presented in section 3, which is followed by our conclusions in section 4.

2.1.1 Motivation and preliminary results

Here we present the specific question that motivated the development of the method as well as some preliminary results indicating that the use of satellite data for DWBC analysis is viable. These results are not the main focus of the present study and will only be briefly described. More details about them are available in a companion paper by Montenegro and Weatherly (submitted), where the method is applied to data from the Atlantic DWBC.

As noted, direct measurements of deep flows are scarce and rarely exceed 2 years. The present work started as an effort to confirm the indications of an yearly signal observed by Weatherly et al. [22] in 18 months of current meter data in the Brazil Basin.

According to Gill and Niiler [52], annual variability in the mid latitude open ocean should be dominated by temperature induced steric oscillations in the thermocline and by wind induced Sverdrup transport below that layer. Jones et al. [53] concludes that SSH and sea surface temperature (SST) correlate well for large scales in the South Atlantic, especially away from the boundaries. As our area of interest is at the boundary, we assume factors other than thermal expansion might be influencing local SSH behavior and that it would be interesting to check if, in the WBC, there exists a correlation between satellite SSH slopes and directly measured deep currents.

For such, Empirical Orthogonal Function (EOF) and spectral analysis were conducted on (SSH) and sea surface temperature (SST). SSH data were monthly values from January 1993 to December 1999 interpolated to a 1° by 1° grid. The sea surface temperature data were also monthly means for the same period and with the same spatial resolution. This data set is a blend of ship, buoy and bias-corrected satellite temperatures provided by the Integrated Global Ocean Services System (IGOSS). All analysis were restricted to Brazil Basin in the western South Atlantic (10° − 25°S and 15° − 45°W).

The spatial pattern of the first SSH EOF, with annual frequency, agreed qualitatively with the peak in integrated DWBC transport between January and March observed by Weatherly et al. [22]. The same general pattern could be observed in the amplitude of the 12 months signal given by spectral analysis of the SSH. The amplitude of the current annual
oscillation (0.2 m/s) resulting from the SSH slopes in the spectral analysis was similar to the directly measured variability. Both results were in good agreement with modelled seasonal patterns of the global bottom pressure ([54]).

The EOF results also showed significant differences between the SSH and SST annual variability spatial patterns, which could be taken as an indication that regional SSH is not predominantly dictated by thermal steric effects in this frequency range and that some information about deep fluxes could be noted in the altimetry data even before any effort to remove the baroclinic signal.

Based on these results we conclude it would be worthwhile to continue and refine this type of analysis. More specifically, a large fraction of the observed SSH slope is compensated by the density field at relatively shallow depths ([55]). Any method trying to infer deep variability from altimetric data would have to somehow remove the effects of these near surface phenomena.

One option would be to estimate dynamic heights from climatological temperature and salinity data. These values could then be subtracted from the SSH data as a first approximation of the baroclinic signal. One problem with such approach is that, near boundaries, the baroclinic component of the circulation will be heavily influenced by phenomena not well represented in the climatology, such as waves, eddies and migration of surface currents. Also, the seasonal thermal expansion cycle is not a dominant factor in WBC SSH variability, not even in the annual frequencies ([53]). The need for a more precise way of removing the near surface effects is what lead to the development of the procedure presented below.

### 2.2 Method

The method aims to describe temporal changes in deep, depth independent currents. The assumption behind it is that these changes are predominantly related to the variability of “barotropic height” slopes (Fig.2.1). Barotropic height (BRT) is operationally defined as the local difference between sea surface height (SSH) and dynamic height anomaly (DYH). The altimetric SSH is defined by:

\[
SSH(s, t) = \eta_a(s, t) - \Phi(s)
\]  

(2.1)
with $\eta_a$ the position of the sea surface given by the altimeter and $\Phi$ the geoid. As indicated, SSH and $\eta_a$ are functions of time and space, while $\Phi$ is a temporal mean.

The DYH is also a deviation from a time mean state, given by:

$$DYH(s,t) = \int_{RL}^{\eta} \delta(s,t) \, dz - \int_{RL}^{\bar{\eta}} \delta(s) \, dz$$

(2.2)

where $\delta$ is the specific volume anomaly, $\eta$ the sea surface and RL a depth reference level. The over bar denotes time mean. Substituting for the symbols in Fig.2.1:

$$DYH = \eta_d - \varphi$$

(2.3)

We have then:

$$BRT = SSH - (DYH/g)$$

(2.4)

where $g$ is the gravitational acceleration.

Given BRT values in two or more distinct points in space, one can use the geostrophic balance to estimate the mean barotropic velocity between them:

$$M_b = \frac{g}{f} \ast \frac{\delta_{BRT}}{\delta_s}$$

(2.5)

where $f$ is the Coriolis parameter, $g$ gravity, and $\delta_{BRT}/\delta_s$ the BRT spatial gradient.

For all following results (2.4) is solved with SSH measured by the altimeter on board TOPEX/POSEIDON (TP). The SSH product used was the along-track values provided by NASA ([56], [57]). DYH values are calculated using XBT derived temperature profiles from which density is estimated by a temperature salinity or temperature density relationship.

This choice of input data imposes two basic limitations. First, Geoid uncertainties in the SSH data render impossible the use of BRT slopes in absolute current estimations. Consequently, results are restrained to temporal variability. We have then, from (3.5):

$$\frac{\Delta M_b}{\Delta t} = \frac{g}{f} \ast \frac{\Delta_{BRT}}{\Delta_s}$$

(2.6)

where $t$ is time and $\Delta$ indicates use of discrete values. (2.6) makes the procedure not only independent from the geoid ($\Phi$) but also from the mean dynamic height ($\varphi$).

Also, as defined by (2.4), the BRT would only represent the total barotropic signal when the reference level (RL) for the DYH calculation is at the bottom. Otherwise, the
BRT refers to a partial barotropic height (relative, in the present case, to the deepest XBT reading). Velocities estimated from a partial BRT slope contain the desired depth independent component plus the baroclinic velocity at the RL. If the method is to be used in estimates of deep flows, the density field below the RL must be relatively stable.

In the tests presented below, $M_b$ time series are obtained following the same general procedure and by integrating (2.6) in time. The first step is to identify areas with good XBT spatial and temporal coverage (the satellite provides essentially constant coverage, but XBT sampling is much more irregular). Next, so that DYH can be obtained from XBT casts,
a local temperature-salinity (T-S) or temperature density (T-D) relationship is constructed from historical hydrographic data. Interpolation in space and time is then applied to the SSH and DYH in order to generate time series for at least two spatial points. These data are then plugged into (2.4) and (2.6).

2.2.1 Applications and limitations

The method described above is tested for the analysis of DWBC low frequency barotropic variability. It is especially well suited for high-energy boundary regions where strong signals are expected to rise above the errors in the SSH and DYH data.

Another possible use would be in studies related to long term variability of kinetic energy near the bottom and its impact in vertical mixing. Also, it could be applied in long term estimates of Sverdrup transport in the interior.

By only resolving the barotropic component, the method cannot be directly applied to study the climatic relevant thermohaline circulation. It does however, offer insight into the barotropic variability in which the density driven flows are imbedded. It can also be useful for investigating meridional heat transport in the annual and seasonal frequencies, where variability is apparently dominated by the balance between surface Ekman fluxes compensated by below thermocline, depth independent, return flows ([36] [35]).

It is worthwhile to point how the basic aspects of this technique have been applied by other researchers. One of the first references of satellite altimeter derived deep current estimates is Weatherly et al. [17]. This study compares flows obtained from GEOSAT SSH gradients to current meter data measured at 5400m in the interior of the Argentine Basin. There is good agreement between SSH derived and directly observed currents for periods larger than 34 days. From this, the authors conclude that, within these time scales, barotropic flow dominated the whole water column during the 11 month experiment.

Blaha and Lunde[55] compares SSH and DYH during the calibration phase of the TOPEX/POSEIDON mission. Their data consists of airplane launched XBT along a TOPEX/POSEIDON track as the satellite orbited above. The authors center their analysis on the similarities between altimetric SSH and XBT derived DYH and mention that part of the misfit might be caused by a barotropic component.
Other studies use XBT and altimetric data to create models that estimate the subsurface density structure from SSH data alone ([58], [59]). Basin and world ocean scale analysis similar to the ones developed here have also been used by projects trying to separate steric from wind induced effects (Sverdrup transport and forced waves) on satellite derived SSH data ([60], [61], [62]).

The method is limited to areas with good temporal and spatial XBT coverage and where measured BRT slopes are larger than the errors in SSH and DYH values. Large distances between XBT and SSH measurements might compromise the quality of the results. Lastly, the maximum temporal resolution is dictated by the satellite orbit (20 days for TP data).

As already stated, the technique will not give a reliable estimate of the barotropic currents when significant changes in the density field occur below the depth sampled by the XBTs. This inherent weakness cannot be overcome by anything but deeper data. There are, however, relatively simple ways of quantifying the errors associated with it.

Either modal decomposition of the local density profile or EOF analysis of nearby current profiles can be used to estimate the amplitude of the baroclinic current variability below the deepest XBT depth. This amplitude can be interpreted as the baroclinic “contamination” of the desired barotropic signal. In other words, what fraction, on average, of the observed variability is being caused by deep steric changes not removed from the SSH data because they were not sampled by the XBT’s.

2.3 Testing the Method

The procedure was originally developed for the analysis of the Atlantic DWBC. Ideally, the method would be tested at or near the area where it would later be applied. Such validation requires that the region around some current meter array in the Atlantic DWBC also be well sampled by XBTs during the period the moorings were in the water. Unfortunately, there are no available data sets with these characteristics. The alternative is to test the method at another conveniently sampled DWBC site, in this case the northern part of the Shikoku basin. To confirm the good performance obtained by the observed data, tests are then conducted using synthetic data.

In the Atlantic, the method was first tested using the current data described in Weatherly et al. [22]. Their array, however, is not located close to a TP track and there was a
limited number of XBT casts near the moorings. The results were inconclusive. The lack of appropriate XBT information presents itself for all the other Atlantic DWBC meter data at our disposal. The option is then to use a numerical model to verify the methodology in areas previously selected for their good XBT coverage.

On all the tests, be it synthetic or real data, the $M_b$s are compared to near bottom flows. This is based on the assumption that these deep currents are essentially depth independent, which is a simplification. Still, we argue that possible errors caused by this choice would not justify the use of some form of partition of the current data into its barotropic and baroclinic signals. Especially because any procedure of this sort would also generate errors.

### 2.3.1 Tests with observed data

#### 2.3.1.1 Shikoku Basin

The method is tested against current meter (CM) data from mooring 6 of WOCE’s Pacific Mooring Array Number 5 (PCM5, Asuka) in the Shikoku Basin off southern Japan (Fig.2.2). These instruments are placed along a T/P track. There is good XBT coverage, with many samplings along the array during the current meter deployment, from Oct. 1993 to Sep 1995. The XBT data comes from the WOCE Upper Ocean Thermal Data Assembly Center.

Observations of a deep boundary current have been reported in this area of the Shikoku basin ([63], [64]). Still, the data set used here was designed for providing reference level values in a study monitoring Kuroshio transport ([65]) and no analysis of the deeper current meters are available in the literature. There is a need then to verify if, in fact, the method is being tested on a DWBC where barotropic variability is important. This is confirmed by visual inspection of the measured current time series (Fig.2.3). There is a clear near bottom intensification of the flow, with values at 4600m larger than at 2400m. There is also good agreement between these two current meters. Local seafloor depth is 4720m. The deep currents have a mean southwestern direction while flows tend to be to the northeast at mid-depth(600-1500m).

The bottom intensification and inversion of direction with depth are features not observed on the current data acquired by the near bottom (2400m) instrument of PCM5 mooring 5, deployed during the same period on the continental slope 29km inshore of mooring 6. Records
from mooring 5 (not shown) exhibit much weaker near-bottom flows with NE mean. PCM5 mooring 7 is the nearest offshore mooring, located 57 km from mooring 6. Mooring 7’s deepest CM was placed at 1360m, in an area where the local depth is 4500m. It had a mean NW flow during the deployment. We take these as an indication the current data used in the test is the best available representation of the local DWBC.

The mean maximum depth of the XBT data is about 450m. Shallower casts are discarded and the ones that went below it truncated. Density profiles for each cast are estimated using seasonal TS relationships constructed from various realizations of WOCE hydrography lines PR17 and PR27. DYH’s are calculated for individual profiles, separated into 10 day blocks (coinciding with satellite orbits) and then interpolated to two points along the satellite track (Xs in Fig.2.2). The interpolation method is a simple average of all DYH’s within a 1/4° radius of each point. The same spatial interpolation is done for the SSH data.

Temporal changes in barotropic currents are calculated using (2.4) and (2.6). These values are integrated in time and compared to current data acquired by the instrument.
Figure 2.3. Time series of current data used in the Shikoku test (circle on Fig.2). Daily values are filtered with a 15 day Hanning filter. Top is the meridional and bottom the zonal component. Thin solid line, 670m; dotted, 1500m; dash-dotted, 2400m; thick solid, 4600m. A 33 day gap around Apr/95 in the 4600 series was linearly interpolated.

positioned at 4600m (circle in Fig.2.2). The CM data is rotated so that values are normal to the array. To test if the method performs better than currents based on SSH and DYH slopes by themselves, these are also calculated by substituting \( \Delta_{BRT}/\Delta_s \) by \( \Delta_{SSH}/\Delta_s \) and \( \Delta_{DYH}/\Delta_s \) in (2.6).

In Fig.2.4, the sum of estimated changes in velocity, given by:

\[
M_b(n) = \sum_{k=1}^{n} \frac{\Delta M_{b}}{\Delta t}(k)
\]  

(2.7)

is compared to the CM flows. To facilitate analysis, the first CM value is subtracted from the rest of the series. Also as \( M_b=0 \) for \( k=1 \) (or \( t=0 \)), the first \( M_b \) estimate seen on the graphs represent the change in \( M_b \) from \( t=0 \) to \( t=1 \). The \( k \) indices in (2.7) refer to number
of BRT estimates and are not to be confused with the much higher number of CM data observations. The XBT data allowed the generation of 34 BRT slopes, hence 33 $\Delta M_b$'s (24 of them during the 4600m CM deployment).

Error bars (lines to the left) relate to temporal smoothing (10 days) and uncertainties in SSH and salinity (1.3 cm/s). The salinity error (54% of total), is very conservative, assuming all salinity deviations along an XBT cast are positive or negative. Still, errors are smaller than the signal’s amplitude.

There is good agreement between current observations and values estimated by BRT slopes (Fig.2.4, top), with correlation coefficient $r = 0.6$. The correlations for the other cases (Fig.2.4, center and bottom) are below the 95% significant level of $r = 0.44$, indicating that BRT slopes increase the quality of the estimate. For the correlation estimate, the current meter data was filtered with a 30 day Hanning filter. This is based on the mean interval between BRT estimated velocities. For the unfiltered meter data the correlation drops to $r = 0.54$.

As explained in the previous section, given the depth of the XBT casts, the $M_b$ time series in Fig.2.4 is in fact the velocity at 450 m. At first, it would seem more sensible to compare it to current data observed closer to this level and not at 4600 m, as has been done. The goal though, is to test how well the method is capable of capturing the barotropic variability, not the flows at a fixed depth, and the 4600 m data should be a better representation of the depth independent flow.

2.3.2 Tests with model data

2.3.2.1 Pacific DWBC

Differences in spatial and temporal resolution, as well as the low number of realizations, make quantitative comparisons based on the observed data above difficult. In order to perform a more objective validation, the method is tested using north Pacific ($5^\circ - 55^\circ$N) data from the OCCAM model. Beyond simple confirmation of the real data results, the model makes possible some general statements about the method’s applicability.

Differently from the altimetry data, model SSH is not a relative value. Based on such information, the procedure could be used to compute absolute speeds and not only their variability. Still, we opt to apply the method as if the SSH came from TP. That is, using
Figure 2.4. Method comparison. Solid lines are the measured and diamonds the estimated speeds. Top: From SSH-DYH (BRT). Center: From SSH only. Bottom: From DYH only. The lines to the left are velocity and time error bars.
equations (2.4), (2.6) and (3.7). Also, as in the Shikoku Basin tests, the $t = 0$ model deep velocity is subtracted from the rest of the series for each grid point.

The version of OCCAM used here has $1/4^\circ$ resolution. The surface is forced by ECMWF monthly mean winds and relaxation to the Levitus seasonal surface temperature and salinity fields.

Temperature (T), salinity (S), SSH and velocity fields are obtained every 10 days for one year. The method is applied to the modelled T, S and SSH. DYH’s are referenced to 600m. This choice of reference level is based on the most common XBT measurements, that have a maximum depth varying between 450m and 850m. Resulting velocities are then compared to the model’s deep currents. In order to minimize topographic effects, deep currents are defined as the velocities two grid boxes above the deepest layer. Given OCCAM’s vertical resolution, this results in currents that are 300 to 500 meters above the bottom. Only areas deeper than 1000m are analyzed.

Fig.2.5 depicts the correlation between BRT estimated and modelled deep velocities near Japan. Only significant values ($r > 0.4$) are plotted. There are relevant correlations of 0.75 and above (lightest gray) along the continental slope of the northern Shikoku Basin ($30^\circ - 34^\circ$N). These values are similar to the $r \simeq 0.6$ of the observed data tests. Also evident is the good performance of the method on the northeastern flank of Japan. This last result must be considered with some care, as the OCCAM version used here does not simulate the Pacific Deep Western Boundary Current well at these latitudes ([66]).

Significant correlations are found over most of the basin, with higher values in the mid basin and along the Western Boundary (Fig.2.6). The methodology performs poorly along the eastern boundary and on the area between $40 - 45^\circ$N off Japan, where OCCAM reproduces the subtropical confluence ([66]). Values in Fig.2.6 have been filtered and interpolated to a $2^\circ$ X $2^\circ$ grid.

A possible cause for the low correlations over the confluence region is that the 600m DYH reference level might be too shallow, so that the BRT estimated velocities still contain a significant baroclinic component. We do not have an explanation for the bad results in the eastern boundary. In any case, no more attention will be given to these basin-scale results, as they are not within the main focus of the present study.
Figure 2.5. Local OCCAM results. Correlation between BRT estimated and model near bottom velocity around Japan. Top, U component; bottom, V component. Blue, $0.4 \leq r < 0.6$; yellow, $0.6 \leq r < 0.75$; red, $r \geq 0.75$. In the white areas the correlation was not significant (below 0.4). Black areas are land.

2.3.2.2 Atlantic DWBC

For the Atlantic tests the CLIPPER Atlantic Ocean circulation model is used. This $1/6^\circ$ resolution model was developed by the French consortium that includes Ifremer and the universities of Paris, Grenoble and Toulouse ([67]). The results used here come from a run that has been spun up for 8 years, then integrated from 1979 to 2000. Tests are conducted on data from the final two years, with outputs every 5 days. The ocean is forced by European Centre for Medium-Range Weather Forecasts (ECMWF) reanalysis winds and
Figure 2.6. Basin OCCAM results. Correlation between model barotropic currents and BRT estimated velocity in mid-latitude North Pacific. Only significant values are plotted. Top, U component; bottom, V component. Blue, $0.4 \leq r < 0.7$; yellow, $0.7 \leq r < 0.9$; red, $r \geq 0.9$. In the white areas the correlation was not significant (below 0.4). Black areas are land.
fluxes. The switch from OCCAM is motivated by two factors. CLIPPER’s higher spatial resolution produces a more reliable DWBC and OCCAM’s North Atlantic results were not available to the community at the time of the study.

As with the OCCAM case, model near bottom velocities are compared to flows estimated by the BRT method using model SSH, salinity and temperature. The reference level for the DYH calculation is 500m. Differently from the Pacific, no basin wide analysis are performed. Tests are conducted on 3 specific sections of the Atlantic DWBC. The goal is to verify the methodology in areas where it would later be applied to study DWBC variability. The position of the DWBC is determined from the 5 year mean model velocity field.

Based on the WOCE Upper Ocean Thermal data set, the 3 chosen locations present the best XBT coverage as far as the BRT method is concerned. In these areas, casts sample approximately the same cross-DWBC transects for more than 6 years with somewhat constant frequency.

While the numerical model produces results every five days, the XBT sampling limits the temporal resolution of any real application of the method to about 1 month in the North Atlantic and 2.5 months in the South. The correlations are calculated with the original CLIPPER output and also with results subsampled every two months. In all areas, correlations are higher for the low frequency data. For both areas, significant correlations (with 95% confidence) are given by $r > 0.07$ for the high frequency (HF) and $r > 0.4$ for the low frequency (LF) comparison.

In the South Atlantic (Fig.2.7) the only insignificant correlation comes from the U component in Area1, where bathymetry is markedly N-S and zonal flows are small. The other values vary from $r = 0.49$ (HF) to $r = 0.79$ (LF).

The XBT rich area in the North Atlantic (around 38°N, Fig.2.8) lies inshore of the mean position of the Gulf Stream. This region is not well reproduced by CLIPPER, where the Gulf Stream separates too far upstream. To circumvent this problem, the test is conducted not in the same geographical position of the XBTs, but in an area where the model presents similar oceanographic features further to the north. Correlations there are 0.87 for the LF to 0.94 for the HF. We believe these higher northern values are due to CLIPPER’s relative homogeneous density structure in this area.
Figure 2.7. Model test in the South Atlantic. Center: Map of area and position of available XBT casts from 09/92 to 12/00. A: Time series of model (dotted) and BRT estimated (solid) U and V on the DWBC. B: Same as Top Right with velocities subsampled every 60 days. C and D: same as A and B for Area 2. The r’s inside plots are correlation coefficients.
2.4 Conclusions

The method is able to reproduce deep velocity variability (considered here a good estimate of the barotropic component) in a satisfactory manner on all performed tests. While the comparisons are generally more favorable for the synthetic data, it is important to remember the observed values come from an experiment not specifically designed to fit the method’s needs.

Due to various factors, among them costs and instrument technical limitations, XBT and SSH measurements are, and should in the foreseeable future remain, much more abundant than direct deep current measurements. While no substitute for in situ observations, this technique could be useful in projects that require low frequency, long term monitoring of barotropic or deep and near bottom flows. While not tested here, it will perform even better in estimating flows near the deepest available XBT depth, which range today between 400 and 800 m.

In Western Boundaries, relatively inexpensive XBT sampling programs could be established, for example, to fill in the gap between successive mooring deployments. This would have the added advantage of having the relative BRT estimated velocities referenced to the current meter data.

The gravity fields measured by GRACE satellites could be used to estimated changes in deep currents in a more direct way than proposed here. The satellite data spatial resolution though, would impose serious limitations on any studies interested in the narrow DWBC region ([68], [69]).

It is worth repeating that, while developed for the DWBC, the method could be used to study other regions, this as long as the temperature and salinity data are deep enough to remove most of the near surface density variability and the barotropic signal of the deep current to be estimated is large compared to the errors in SSH and DYH values.
Figure 2.8. Model test in North Atlantic. Bottom: Map of area and position of available XBT casts from 09/92 to 12/00. A: Time series of model (dotted) and BRT estimated (solid) U and V on the DWBC. B: Same as A, with velocities subsampled every 60 days. The r’s inside plots are correlation coefficients. Note the test was not performed on area of dense XBT coverage as explained in the text.
CHAPTER 3

BAROTROPIC VARIABILITY IN THE ATLANTIC WBC

3.1 Introduction

Western boundary currents (WBC) are conspicuous features of the ocean circulation. They are important in determining ocean heat, salt and freshwater transports ([33], [48], [34]), and have also been seen to influence biological parameters such as nutrient availability and primary production ([49], [45], [44]).

Better understanding of these phenomena is necessarily linked to good estimates of not only WBC variability, but also the physical processes behind such changes. Still, technical and financial constraints limit the number and span of WBC direct eulerian measurements, which are usually no longer than 2 years. Records of such length are at the limit or below the period needed to describe such fundamental modes of variability as the annual cycle. Observations are rarer and tend to be shorter in the case of Deep West Boundary Currents (DWBC) (for example, [11], [50], [25], [51], [22]).

Here we study the annual to interannual barotropic variability in two sites of the Atlantic WBC based on current time series that span approximately 6 years. The series are obtained by a blend of satellite altimetric measurements and Expendable Bathythermographs (XBT) temperature profiles, using the method described by Montenegro and Weatherly (submitted to JGR).

Section 2 presents a brief review of forcing mechanisms associated with WBC variability, concentrating on the ones pertinent to the present study. In section 3, Empirical Orthogonal Function analysis of surface temperature and altimetric data is used to demonstrate the feasibility of using the last in studies of WBC barotropic currents. The method used for the
current estimates is described in section 4. Results derived from application of the method in two areas of the Atlantic WBC are presented in sections 5 and 6. These results are then discussed and summarized in section 7.

3.2 Possible forcings of WBC barotropic variability

Here we summarize previous studies dealing with WBC barotropic variability. Given the temporal resolution of our estimated time series (approximately 1 to 2.5 months), attention is focused on the seasonal to interannual range.

Willebrand et al. [6], based on analytical and numerical results, conclude that large scale, stochastic wind forcing with periods between 1 and 300 days generate a predominantly barotropic oceanic response. According to these authors, for periods above 30 days, this response is in the form of a wind stress curl forced, Sverdrup like, interior flow compensated by a return in the WBC. Their conclusions are confirmed by further numerical and analytical studies conducted by [13]. This would imply a negative correlation between basin wide wind stress curl and the WBC. A series of observational efforts corroborate this relationship ([10], [8], [19], [14], [11]).

On a local scale, Lee and Williams [19] (based on results from [18]) use a simple coastal response analytical model to relate current variability and along-channel wind stress in the Florida Strait for time scales of a few days. In their model, the along-channel wind affects the along-shore currents via convergence and divergence generated by Ekman transport at the coastal boundaries. The authors identify an annual correlation between along-channel wind stress and WBC transport, which they attribute to modulation of the signal by seasonal differences in storm incidence. Other works state that the same process influences the yearly variability of the Kuroshio east or Taiwan ([14]).

The mechanism is described and observed for flows between two boundaries and, as such, would not fit the characteristics of the areas where our data comes from. Still, while the presence of the second boundary reinforces cross-bathymetry flow, it is not a necessary condition for its establishment. The anomaly generated by coastal convergence should have a horizontal scale equivalent to the barotropic Rossby radius. This means that the sea level signal should extend across most shelves and be present over the continental slope. Some evidence for such is observed in 5.8 years of current meter data east of the Bahamas ([11]).
A series of modelling efforts in the Tokara Strait, south of Japan, also indicate the existence of a barotropic response to along-shore wind forcing ([16], [17] [15]). Of particular interest is the work of [17], who observe the effects of convergence at the coast forcing the WBC in a manner similar to the process proposed by Lee and Williams [19]. An important difference is that their numerical model uses monthly winds, indicating this type of response is not necessarily restricted to time scale of days.

Based on the above, we deem it worthwhile to compare our estimated currents to local along-shore winds. The goal is to test the hypothesis that seasonal to interannual divergence of the Ekman flow on the coast generates a WBC barotropic response even when only one boundary is present.

Observed WBC barotropic variability with periods of 150 days have been attributed to reflected Rossby waves ([9]). This process is not analyzed here.

Before presenting the current time series and the method used in their estimation, we discuss some Empirical Orthogonal Function (EOF) analysis results that indicate altimetric data is a viable source of information for the study of WBC barotropic variability.

### 3.3 SSH and SST EOFs

EOF analysis of SSH and sea surface temperature (SST) data in the Brazil Basin are presented to test the feasibility of using TP data for the study of depth independent flows. The EOFs are also used to verify the existence of an apparent yearly signal observed by [22] in 18 months of current meter data in the Brazil Basin DWBC.

The data for the SSH EOF analysis were monthly, 1°x1° gridded fields spanning the period between Jan. 1993 to Dec. 1999, provided by NASA ([56]). $|SSH| \geq 1.2$m were discarded. The SST data were also monthly grids with the same spatial resolution. The temperature set is a blend of ship, buoy and bias-corrected satellite measurements provided by the Integrated Global Ocean Services System (IGOSS).

Prior to the EOF calculation, both SSH and SST fields were smoothed by a 5° x 5° 2D Hanning filter. All analysis were restricted to the Brazil Basin in the western South Atlantic (10° – 25°S and 15° – 45°W). Only areas deeper than 200 m were evaluated.

Results from the first SSH and SST EOFs are displayed in Fig.3.1. Both are dominated by an annual signal that represent 30% of SSH and 58% of SST total variance. The yearly
oscillation of the temperature field is disrupted in the end of the period between mid-1997 and 1999. The temperature variability presents a zonal structure. The SSH pattern west of the 4000m isobath, specially in the northern part of the domain, is characterized by cross-slope gradients. There is some indication that the offshore SSH gradient is accompanying the topographic inflection of the Trindade-Vitória Sea-Mount Chain near 20°S.

The SSH slope and phase provided by the first EOF over the continental slope around 18°S agrees qualitatively with the peak in integrated DWBC transport between January and March observed by Weatherly et al. [22]. The same general pattern could be observed in the amplitude of the 12 months signal given by spectral analysis of the SSH data. The amplitude of the current annual oscillation (0.2 m/s) resulting from the SSH slopes in the spectral analysis was similar to the directly measured variability. Both results were in good agreement with modelled seasonal patterns of the global bottom pressure ([54]).

For the above analysis, no effort was made to separate the SSH values into their depth dependent and depth independent portions. This being said, comparisons between SSH slopes and deep or near bottom speeds are, given the well documented correlation between altimetric slopes and above thermocline velocity fields ([70], [71] [65], for some examples), to be taken carefully.

More pertinent to this study are the marked differences between SSH and SST EOFs spatial patterns. These are taken as an indication that, in this frequency range, regional SSH is not predominantly dictated by thermal steric effects. Also, they suggest that annual barotropic variability in the DWB has a surface signature above the noise level of the instrument aboard TP. The method used to subtract surface variability from this signal is described below.

### 3.4 The barotropic height method

The estimated current time series analyzed here are obtained using the method described and tested by Montenegro and Weatherly (submitted to JGR). Its goal is to make possible the inference of depth independent variability from altimetric data by removing the effects of near surface phenomena from SSH values. The procedure is based on “barotropic height” slopes across the Western Boundary. Barotropic height (BRT) is operationally defined as the local difference between sea surface height (SSH) and dynamic height anomaly (DYH)
Figure 3.1. Brazil Basin EOFs. Top two figures: Spatial pattern and respective time series of the first SSH EOF (30% of the variance). The blue lines are 4000m isobaths. The units of the time series y axis are meters. Bottom two figures: Same as top but for the SST first EOF (58% of the variance). Units in the time series are °C.
The altimetric SSH is defined by:

\[ SSH(s, t) = \eta_a(s, t) - \Phi(s) \]  

(3.1)

with \( \eta_a \) the position of the sea surface given by the altimeter and \( \Phi \) the geoid. As indicated, SSH and \( \eta_a \) are functions of time and space, while \( \Phi \) is a temporal mean.

The DYH is also a deviation from a time mean state, given by:

\[ DYH(s, t) = \int_{RL}^{\eta} \delta(s, t) \, dz - \int_{RL}^{\eta} \bar{\delta}(s) \, dz \]  

(3.2)

where \( \delta \) is the specific volume anomaly, \( \eta \) the sea surface and RL a depth reference level. The over bar denotes time mean. Substituting for the symbols in Fig.3.2:

\[ DYH = \eta_d - \varphi \]  

(3.3)

We have then:

\[ BRT = SSH - (DYH/g) \]  

(3.4)

where \( g \) is gravity.

Given BRT values in two or more distinct points in space, one can use the geostrophic balance to estimate the mean barotropic velocity between them:

\[ M_b = \frac{g}{f} \cdot \frac{\delta_{BRT}}{\delta_s} \]  

(3.5)

where \( f \) is the Coriolis parameter and \( \delta_{BRT}/\delta_s \) the BRT spatial gradient.

Equation (3.4) is solved with SSH measured by the altimeter on board TOPEX/POSEIDON (TP). The SSH data used are along-track values provided by NASA ([56], [57]). DYH’s are calculated using XBT temperature profiles from which density is estimated by a temperature density relationship. The XBT data comes from the WOCE Upper Ocean Thermal Data Assembly Center (WOCE-UOT). This choice of input data imposes two basic limitations. First, geoid uncertainties in the SSH data render impossible the use of BRT slopes in absolute
current estimations. Consequently, results are restrained to temporal variability. We have then, from (3.5):

$$\frac{\Delta M_b}{\Delta t} = \frac{g}{f} \frac{\Delta[\Delta BRT]}{\Delta t}$$

(3.6)

where t is time and \( \Delta \) indicates use of discrete values. Equation (3.6) makes the procedure not only independent from the geoid (\( \Phi \)), but also from the mean dynamic height (\( \varphi \)). Current time series are obtained by integrating (3.6) in time, or, since values are discrete:

$$M_b(n) = \sum_{k=1}^{n} \frac{\Delta M_b}{\Delta t}(k)$$

(3.7)
The other limitation is that, as defined by (3.4), BRT would only represent the total barotropic signal when the reference level (RL) for the DYH calculation is at the bottom. Otherwise, BRT refers to a partial barotropic height (relative, in the present case, to the deepest XBT reading). Velocities estimated from a partial BRT slope contain the desired depth independent component plus the baroclinic velocity at the RL.

### 3.4.1 Method application

The choice of XBTs as the source of density information for the method is based on the fact that XBT profiles are, by far, the most abundant type of sub-surface oceanographic measurement. Still, they are incomparably scarcer than the satellite information. Consequently, while in theory the method’s results are restricted by both SSH and DYH data characteristics, any practical application of the technique will, in truth, be constrained by limitations on XBT temporal and spatial availability.

With this in mind, and given the present project’s goals, the WOCE Upper Ocean Thermal data set is searched for ideal locations, as far as XBT availability. Two are selected. One is in the South Atlantic, where no DWBC measurements longer than 2 years exist. The other is near 38°N, to the north of the Gulf Stream separation. This location presents a dynamical setting different from the area sampled for almost 10 years near 26° ([7]). At both sites, specially at the northern one, XBT casts sample approximately the same cross-western boundary transects for more than 6 years with somewhat constant frequency. The methodology performed well in tests conducted with model data in both areas. (*Montenegro and Weatherly*, submitted to JGR)

### 3.5 South Atlantic WBC

The present inspection of WBC variability in the South Atlantic is centered around a time series generated by the BRT method between 6 – 10°S (Fig. 3.3). As stated above, the goal is to analyze the relative importance of two forcing mechanisms: local along-shore wind stress and basin wide wind stress curl. A comparison between the estimated currents and transport measured at 18°S is also conducted.
3.5.1 Estimated time series

3.5.1.1 Method implementation

Here we describe how the DYH and SSH values used as input in equation (3.4) are obtained. First the DYH:

In the chosen area (rectangle in Fig.3.3), XBT casts that sample the slope (from 800m and 4200m) are selected. The mean maximum profile depth of these XBTs is about 400m. Shallower casts are discarded and the ones that extend below it are truncated. Density profiles for each cast are estimated using a temperature-density relationship based on local historical CTD data from the National Oceanographic Data Center (NODC).

DYH values are calculated for individual density profiles. Next, DYHs are divided into shallow (800m to 2500m) and deep (2500m to 4200m) values. These are then averaged into 40 day blocks (coinciding with 4 TP orbits). The results are two (one shallow, one deep) DYH times series. Not all 40 day blocks contained enough data. As a result, the intervals between values in the DYH (and consequently current) time series are irregular. The mean interval between estimates is around 2.5 months.

The SSH data comes from TP’s orbit 252 along track data (Fig.3.3, insert). The measurements for each orbit are divided into deep and shallow values using the same depth limits of the DYH procedure. Next, they are averaged in time so that there is a mean SSH for each value in the DYH time series.

Even though the DYH averaging interval correspond to the period of 4 satellite passes, the XBT casts are not evenly distributed within the 40 day blocks. Therefore, when calculating the mean SSHs, only orbits that coincide with XBT casts are used. On average, each SSH and DYH value is based on a ~ 30 day mean. An example of the spatial distribution of the data used for one BRT estimation is given in the insert of Fig.3.3.

The SSH and DYH time series are used for estimating BRT values. Equations (3.6) and (3.7) are then used to produce a time series of along-slope barotropic current ($M_b$) variability containing 30 estimates from Feb. 1993 to Apr. 1999.

3.5.1.2 Error estimates

Here we analyze SSH and DYH errors and estimate their effect on the $M_b$ time series. The SSH uncertainty, $e_{ssh} = 2cm$, is determined by the altimeter’s precision. DYH errors can...
arise from errors in the XBT temperature and errors generated by the temperature density (T-D) conversion. The magnitude of the first is 0.1°C (according to the WOCE-UOT). These errors are considered insignificant compared to the latter effect and are not analyzed.

The errors in the T-D relationship depend on the scatter of the T-D curves and are a function of temperature. It is assumed that, for all temperatures, this error has a normal distribution. The following steps are taken to determine their impact on DYH:

First, based on the T-D curves from all CTD data, the standard deviation of density as a function of temperature, \( \sigma_d(T) \) is established. Next, 500 density anomaly profiles are created using: \( Ap(T) = 2 \times \sigma_d(T) \times \text{rand} \), where \( \text{rand} \) is a normally distributed random number and \( (T) \) indicates the temperature dependency. Then, the previously obtained polynomial relating temperature and salinity is applied to the mean XBT temperature profile, establishing a mean estimated density profile.

This profile is added to the 500 \( Ap \)s, generating 500 perturbed density profiles which are used to calculate 500 perturbed DYHs. Finally, the DYH error caused by inaccuracy of the T-D relationship is calculated by \( e_{dyh} = 2 \times \sigma_{dyh} \sim 1 \text{cm} \), with \( \sigma_{dyh} \) being the standard deviation of perturbed DYHs. For all error calculations DYH is really \( DYH \times g \), hence the value has length units.

As defined, \( e_{ssh} \) and \( e_{dyh} \) refer to individual measurements. It is assumed that both errors have normal distributions and are independent in time and space from each other. Noting that each BRT estimate is based on a average of 12 (6 deep and 6 shallow) DYH and 25 (15 deep and 10 shallow) SSH observations, the mean final errors are given by:

\[
fe_{dyh} = e_{dyh} / \sqrt{N} = 0.42 \text{cm}
\]

where \( N \) is the number of observations. The same expression applied to the SSH error will result in:

\[
Sfe_{ssh} = e_{ssh} / \sqrt{N} = 0.63 \text{cm}
\]

and

\[
Dfe_{ssh} = 0.52 \text{cm}
\]

this results from the distinct \( N \) values for shallow and deep areas.

The estimated \( fe_{dyh}, Sfe_{ssh} \) and \( Dfe_{ssh} \) are used to obtain the BRT slope error and finally the velocity error for the time series used in all present analysis: \( e_{vel} \pm 0.01 \text{m/s} \).
As already mentioned, currents obtained through the BRT method are good estimates of barotropic flow only if the density variability below the reference level of the dynamic height calculations is small. By examining temperature profiles from the XBT casts and the climatological monthly temperatures at and near 400m in the area ([72]) we conclude that the 400m reference level is deep enough to capture most of the baroclinic signal. As a result, no effort is made to quantify errors associated with baroclinic contamination of the estimated barotropic currents in this area.

3.5.1.3 Results

Before analyzing the comparisons between estimated \( M_b \) currents and the wind forcing (Fig.3.4), some facts about the wind information used are presented. The wind data comes from the scatterometers on board the European Remote Sensing satellite systems (ERS) 1 and 2. They provide constant coverage during the period of interest. The formats used are monthly mean 1°x1° gridded wind stress and wind stress curl fields obtained from CERSAT, at IFREMER, Plouzané (France).

These monthly data are filtered and interpolated to the periods where BRT slopes are available. In the analysis, WSC refers to the mean wind stress curl in the area between 6° – 10°S from the shelf to 15°W. This last value is taken as the mean position of the Mid-Atlantic Ridge. LWS (for Local Wind Stress) is the mean along-shore component of the wind stress in the area delineated in Fig.3.3.

The adopted wind error is chosen as the larger value between two estimates: the error provided by CERSAT and the RMS difference between the ERS and the pseudo-wind stress from the Servain wind field over the Tropical Atlantic (the data is obtained at www.coaps.fsu.edu/woce/SAC/atlantic). This results in a LWS error of \( \pm 6 \times 10^{-2} Pa \) and WSC uncertainties of \( \pm 10 \times 10^{-8} Pa/m \).

The bottom panel of Fig.3.4 shows the number of XBT casts used in each current estimation plotted above. They serve as an indirect measurement of \( M_b \) reliability. Attention is called to the small number of casts around Jan. 1996. Remembering that \( M_b \) values are a time integral, the impact of this lack of data should be lower quality estimates from late 1995 to early 1996 and some shift in the estimates that follow this period. The potentially
erroneous values are not considered in the quantitative analysis, but the shifted ones are included.

The $M_b$ amplitude (Fig.3.4) results in an estimated barotropic transport variability of $\pm 9$Sv. The top panel indicates a negative correlation, with near annual frequency, between $M_b$ and WSC. This is consistent with the boundary response to an interior in Sverdrup balance. The central panel shows a positive correlation, also in the yearly range, between LWS and $M_b$ and indicates a possible local control mechanism for some of the barotropic variability.

Simple zero lag correlation between current and the wind data favor LWS ($r_{LWS} = 0.83$) as the main forcing. In fact the correlation between $M_b$ and WSC ($r_{WSC} = 0.44$) is just above the 95% significance level of $r = 0.41$.

The above significance level of $r = 0.41$ is based on the total number of $M_b$ estimates. A more stringent test uses not all observations but only the ones that are independent from each other. The measure of independence is given by the zero crossing of the autocorrelation function. Using this procedure the new 95% significance level is $r = 0.48$, above the correlation value observed between WSC and $M_b$.

Results are the same when a linear regression model of the form

$$M_b = a W_S + b W_{SC}$$

(3.8)

is applied to the data. The residual, or model error, is in the order of 53% of the total current variance. When (3.8) is inverted taking only LWS into account ($b \equiv 0$), the error only changes to 55%. On the other hand, if only WSC is taken into consideration the residual rises to 85%.

On the top panel of Fig.3.5, the estimated currents are plotted together with the WBC transport anomaly measured by at 18°S. Given the differences in location, method, spatial and temporal resolution between the two series, any comparisons must be made with caution. This being said, both series present the same general configuration, with intensification of southern flows during the Southern Hemisphere summer and more northerly fluxes during winter and spring.

In their study, Weatherly et al. [22] did not observe a clear correlation between WBC transport and basin wind stress curl from National Center for Environmental Prediction
(NCEP) reanalysis data. Similar results are obtained when the ERS winds are used. As seen in the center panel of Fig.3.5, for most of the record, the correlation has the wrong sign and currents tend to precede the wind. Still, from the end of 1994 onwards, transport and WSC relate in a way that is coherent with Sverdrup response. In this figure, WSC and LWS are not the same values used in 3.4, but refer to spatial averages relevant to the location of the transport estimate.

The relationship between LWS and transport is shown by the bottom plate of Fig.3.5. The positive correlation is like the one detected by the estimated series further north. Still, there are some phase differences, specially in beginning of 1994, with the LWS leading the currents.

The qualitative interpretation of the results in Fig.3.5 tends to confirm the prevalence of local wind stress over basin wind curl as a controlling mechanism of WBC barotropic variability. Still, as inferences about an annual signal are being made based upon 18 months of data, we feel further tests are needed. For this we turn to a numerical model.

### 3.5.2 Model results

This section deals with data from the CLIPPER Atlantic Ocean circulation model, a project developed by a French consortium that includes IFREMER and the universities of Paris, Grenoble and Toulouse ([67]). Results come from a version with 1/6° horizontal resolution and 42 vertical levels. The data span the period between Jan. 1995 to Dec. 2000 with outputs every 5 days. The ocean is forced by European Centre for Medium-Range Weather Forecasts (ECMWF) reanalysis winds and fluxes.

For the comparisons, the near bottom flows in the Deep Western Boundary Current (DWBC) are used as a representation of the western boundary barotropic velocity. In order to minimize topographic effects, near-bottom currents are taken as the velocities in the two grid boxes lying immediately above the deepest bin. On average, given the code’s configuration, this is a 400m layer located 200m above the bottom.

The DWBC values are obtained by the following steps. First, the near-bottom currents over the slope (local depth between 1000m and 4000) are averaged for all available years. The DWBC is then defined as the mean, between 6° – 10°S, of all slope bins that exhibit
poleward near-bottom flow during the 5 year average. The LWS and WSC are calculated like before, but based on the ECMWF winds used to force the model.

The magnitude of model DWBC variability (Fig.3.6) is similar to the amplitude of the estimated \( M_b \) time series. Anomalies tend to be to the south during the Austral Summer and to the north during winter and fall, which agrees with the annual phase of the \( M_b \) and transport data. Visual comparison with the wind data indicates, again in agreement with the estimated currents, that the LWS seems more important than WSC in controlling the barotropic variability.

This LWS dominance is further demonstrated by cross spectral analysis of the synthetic data (Fig.3.7). Both LWS and WSC are significantly correlated with the currents at the annual frequency. Still, the coherence between the LWS-DWB correlation (~ 0.8) is considerably larger than the WSC-DWB one (~ 0.45). Note also that phase relationships (near 0 for LWS-DWBC and nearly 180° for WSC-DWBC) are in accordance with the expected physical mechanisms relating the forcing to the currents.

3.6 North Atlantic WBC

3.6.1 Estimated time series

3.6.1.1 Method implementation

Data sources and procedures for obtaining the \( M_b \), WSC and LWS time series are the same ones used at the southern site. Parameters are adjusted according to local characteristics and data availability.

The BRTs are calculated for the area lying between 37.7°-39.2°N and 70.6°-73.1°W (Fig.3.8, area A). For the DYHs, XBT profiles are truncated at 450m. The SSH data comes from TP orbit 50. The shallow area includes depths from 1000m to 2800m and the deep area includes depths from 2800m to 3800m.

This area is better sampled than the south. Local XBT observations generate BRT values that represent, on average, a 10 day mean. The resulting \( M_b \) time series is composed of 82 estimates with nearly monthly resolution from May 1993 to July 1999.

Area A is relatively small when compared to the ERS data 1°x1° grid. As a result, LWS values are averaged over an area slightly larger that the one used for the current estimation.
Its limits are from 36° to 40°N, extending zonally from the coast to 67°W. The WSC band has the same meridional boundaries and goes from the coast to 40°W.

### 3.6.1.2 Error estimates

Repeating the steps described above, the following errors are obtained:

- $M_b$ error - $e_{vel} \pm 0.07 m/s$
- LWS error - $e_{ws} \pm 10 \times 10^{-2} Pa/m$
- WSC error - $e_{wsc} \pm 18 \times 10^{-8} Pa/m$

The LWS error is given by CERSAT and the WSC error is the RMS between the ERS and NCEP reanalysis wind fields.

In contrast to the south, the local temperature field is not stable below the maximum XBT profile depth. This can be inferred from the time variability of temperature near 450m (Fig.3.10,bottom panels). These changes might be associated with passage of Gulf-Stream rings, well documented in the area ([73]).

In order to quantify the impact this has on the $M_b$ velocities, we estimate the amplitude of the baroclinic current variability immediately below 450m. The implicit concept is that this parameter is related to the portion of the baroclinic signal not captured by the DYH calculation.

This is done by projecting current profiles on the first two baroclinic dynamic modes obtained from local CTD derived densities (Fig.3.9, left). The CTD data are the same used for the temperature density relationship.

Velocity values come from the NODC Coastal Ocean Time Series Database. They are current meter observations from a mooring located inside area A (38.65°N and 72.5°W) at 2650m depth. There are five instruments positioned at 196, 294, 587, 1824 and 2417m. The current time series is 6 months long (Nov. 1990 to Jun. 1991). Original hourly values are smoothed with a 30 day Hanning filter and interpolated to daily profiles. These are then projected into the modes (Fig.3.9, right).

The sum of first and second mode standard deviations at 500m is 0.018 m/s for the U projection and 0.026 m/s for V. The standard deviation of the $M_b$ data is 0.11. This would
indicate that the baroclinic contamination of the estimated barotropic signal varies from 16-24%.

Taking into account the lack of vertical resolution, and especially the restricted spatial and temporal scope of the current meter data, this is considered to be a rough estimate.

### 3.6.1.3 Results

The top two panels on Fig.3.10 show $M_b$ and wind anomalies. All values are smoothed with a 7 month Hanning filter. The $M_b$ amplitude results in estimated transport variability of $\pm40$Sv. Visual inspection does not indicate a clear relationship between currents and LWS. In the WSC case, an inverse, interannual correlation is apparent.

The temperature profiles on the two lower panels of Fig.3.10, like the number of XBT casts from Fig.3.4, offer a qualitative indication of $M_b$ reliability. The dynamic mode analysis from the last section provides a mean value, but the baroclinic contamination varies in time. It should be larger for periods of rapid temperature change near 450m.

Cross-spectral analysis between the $M_b$ and wind data (Fig.3.11) indicate a clear annual correlation between currents and LWS. According to the phase diagram, the local winds lead the currents by 28 days, near the 30 day resolution of the input data series (for the spectrum calculations, $M_b$ values are linearly interpolated to the ERS wind dates). The WSC x $M_b$ cross-spectrum presents two significant peaks, a dominant one in the interannual range and a less energetic peak with yearly period. In both cases, the phase is coherent with interior Sverdrup forcing.

### 3.6.2 Model results

All analyses based on observations are repeated using data from the CLIPPER model and ECMWF winds. Here, however, model and wind data do not come from the same site used in the $M_b$ estimation, but from a region further north (B in Fig.3.8). In the real ocean, area A lies inshore of the mean Gulf Stream path. In the model, however, this region is occupied by the Gulf Stream, that separates too far upstream (A. Treguier, personal communication). Apart from this spatial displacement, data sources and treatment are the same as in the south.
No correlations are evident in the model DWBC and LWS or WSC time series (Fig. 3.12). If any parallel can be drawn between model and estimated currents it is the absence of a clear annual signal with constant phase.

Some similarities though, arise in the model cross-spectrum (Fig. 3.13. As observed for the estimated $M_b$, there is a significant positive correlation between LWS and model DWBC in the yearly range. Also present in both cases is the negative correlation between currents and WSC over interannual frequencies. Model currents and WSC are not significantly correlated over a one year period. Not present in the observed values but very prominent in both LWS and WSC for the model data are the correlation peaks near 90 days.

As can be noted in Fig. 3.10, CLIPPER currents are nearly one order of magnitude smaller than the northern $M_b$s. The first explanation would be that results come from geographically distinct areas. Area B is only chosen as the southernmost area after the model Gulf Stream separation. No attempt is made to verify if it is similar to area A in other ways.

A second possible cause is distinctions in spatial averaging. The model currents represent means over a larger area. The segment of the western boundary covered by area B is much longer than the one in area A. Adding to this difference is the disposition of the XBT casts and TP track inside area A. This ends up producing many $M_b$ measurements that are better described as transects than as along-slope means.

Another possibility is the already mentioned presence of near-surface variability in the $M_b$ values. Still, assuming our estimate of such impact is correct, this would be too small to explain the observed differences.

No matter what the reason, the goal of the model analysis is not to check the precision of the BRT method in reproducing velocity values. Model DWBC and estimated $M_b$ represent, albeit expressed by the same physical quantity, different parameters. What they do have in common, and the reason they are compared here, is that both can be treated as proxies of the barotropic variability at the western boundary.

The experiments with synthetic data demonstrate whether observed relationships between $M_b$ and the wind are also seen in the model. The conclusion is that, with the exception of the annual correlation between $M_b$ and the WSC, all of the relationships are observed.
3.7 Discussion and conclusions

In both areas, there are indications of a correlation between the interior wind curl annual oscillation and the estimated barotropic currents. Care must be taken in the southern site, where the correlation is borderline significant at best. While this may be the first observational evidence of this relationship in the low latitude South Atlantic, this is a well established phenomenon, previously described in many other studies. In this sense, maybe more than a novel result, it serves as further validation of the methodology.

References to an interannual signal relating the interior Sverdrup transport to WBC flows, as observed in the north location, also exist, but are less abundant ([12]). An immediate explanation is the paucity of long term records. Also, while not explicitly mentioning an interannual response, [11] and [14], detect large year to year changes within the annual barotropic transport cycles of their 5.8 and 7 year time series. Lastly, it must be noted that there is not much energy in the north wind stress curl signal at annual frequencies.

According to Willebrand et al. [6], wind energy becomes available to baroclinic waves for periods above 300 days. Decadal scale variability observed in Gulf Stream transport estimates have been associated with first and second mode baroclinic Rossby waves forced by interior curl ([5]). An argument can be made that the interannual signal in the north is in fact baroclinic in nature.

Still, the amplitude of the observed interannual oscillation is much larger than the estimated error (16-24%) due to residual baroclinic effects. Also, a lag between wind and currents would be expected in a baroclinic response. Further support is given by the modelled results, where the interannual signal is also present.

Comparisons between Sverdrup flow and estimated barotropic transports confirm and lend some quantitative credence to the described WSC-$M_b$ relationship. The amplitude of the interior Sverdrup transport variability is given by the area integral, over the same region used for the WSC average, of $S_{\sigma} = \frac{WSC_{\sigma}}{\rho \beta}$. With $WSC_{\sigma}$ being the standard deviation of the WSC, $\rho$ the water density and $\beta$ the meridional derivative of $f$. $M_b$ transport variability is calculated by $V_{\sigma} = AM_{\sigma}$, where $M_{\sigma}$ is the $M_b$ standard deviation and $A$ is the cross-section area over which the estimates are obtained.
For the south, this generates $M_{h\sigma}/Sv_{\sigma} = 4.5/7.4$, or 61%. The same ratio for the north is 20.3/35.7, or 57%. Values are in Sverdrups. These results indicate that measured variabilities are within the magnitude of the interior forcing. Given the simplicity of the Sverdrup transport estimation and the presence of the local along-shore forcing, a one to one correlation between transports is not expected.

The variability in the north is dominated by the interannual signal of the wind curl. Impacts of local stress are secondary. Analysis of the time series (Fig.3.10) indicates a clear relationship between currents and LWS during the first years of the record. The correlation, though, becomes untraceable after early 1996, when the amplitude of the local wind becomes smaller and anomalies tend to be more positive. A tentative explanation is that Ekman induced convergence at the coast is not as efficient as divergence in generating the cross-bathymetry surface slopes.

The south area currents, on the other hand, seem primarily influenced by local processes. The narrow shelf and lower latitude make the southern site more sensitive to the along-shore wind stress than the north. Based on the time dependent model of Lee and Williams [19], if forced by the same wind stress value, currents in the south are 50% larger than their northern counterparts. Given the much stronger northern winds, this cannot serve as an explanation for the distinct responses. Still, this local predominance is not only present in the $M_h$ estimates, it is also very clear in the model results and some indication of it can be inferred from the Brazil Basin transport.
Figure 3.3. South Atlantic site. Black dots represent location of all available WOCE XBT profiles between 09/1992 and 12/2000. Method was applied to data from area delineated in black rectangle. **Insert:** example of data spatial distribution for one BRT calculation. Thick line is the satellite track (SSH), diamonds are XBT casts (DYH).
Figure 3.4. South site estimated time series. **Top:** Solid line, Estimated $M_b$ time series for the South Atlantic site; dot-dash line, mean cross-basin wind stress curl anomaly. Vertical lines on left and right are current and wind error bars. Half circles on the current line mark time of BRT slope estimates. **Center:** Same as top, but with dot-dash line being the local along-shore wind stress anomaly. All series smoothed with a 5 month block filter. **Bottom:** Number of XBT casts used in each BRT slope estimate. Circles, casts in shallow area; asterisks, casts in deep area.
Figure 3.5. Analysis of measured DWBC variability. **Top**: Solid line, transport anomaly at 18°S; dot-dash line, estimated $M_b$ time series. Vertical line on left is current error. Circles on $M_b$ line mark time of BRT slope estimates. **Center**: Solid line, same as top; dot-dash line, mean cross-basin wind stress curl anomaly. Vertical line on left is wind error. **Bottom**: Same as center, with dot-dash line being the local along-shore wind stress anomaly.
Figure 3.6. Clipper time series for south site. **Top:** Solid line, model DWBC anomaly; dot-dash line, wind stress curl anomaly. **Bottom:** Same as top, with dot-dash line being the local along-shore wind stress anomaly. All series are smoothed with a 6 months Hanning filter.
Figure 3.7. Clipper spectral analysis for south site. **Left-top**: model DWBC x LWS cross spectral density, vertical lines are error estimates, \( p \) is the peak’s mean period and \( f \) its frequency. **Left-center** model DWBC x LWS coherence, the dash-dot line marks the significant level. **Left-bottom**: model DWBC x LWS phase. **Right column**: same as left, but for DWBC x WSC cross spectrum.
Figure 3.8. North Atlantic site. Black dots represent location of all available WOCE XBT profiles between 09/1992 and 12/2000. Method was applied to data from area A, delineated in black rectangle near 38°N. Model results come from area B, near 43°N. Insert: Blowup of area A with an example of data spatial distribution for one BRT calculation. Plus signs are satellite altimeter readings (SSH), diamonds are XBT casts (DYH).
Figure 3.9. Baroclinic contamination. **Left:** First, blue, and second, red, dynamic modes for area A based on CTD data. **Right-top:** Projection of daily average zonal current profiles on first, blue, and second, red, dynamic modes. Velocity values are in cm/s. **Right-bottom:** same as top, for meridional currents.
Figure 3.10. North site estimated time series. **Top**: Solid line, Estimated $M_b$ time series for the North Atlantic site; dot-dash line, mean cross-basin wind stress curl anomaly (WSC). Vertical lines on left and right are current and wind error bars. **Center**: Same as top, but with dot-dash line being the local along-shore wind stress anomaly (LWS). All series smoothed with a 7 month Hanning filter. **Bottom contour plots**: Temporal evolution of the XBT temperature profiles used in estimating shallow and deep BRTs. Values on gray scale are °C.
Figure 3.11. North site spectral analysis. **Left-top**: $M_b \times LWS$ cross spectral density, vertical lines are error estimates, $p$ is the peak’s mean period and $f$ its frequency. **Left-center**: $M_b \times LWS$ coherence, the dash-dot line marks the significant level. **Left-bottom**: $M_b \times LWS$ phase. **Right column** same as left, but for $M_b \times WSC$ cross spectrum.
Figure 3.12. Clipper time series for north site. **Top**: Solid line, model DWBC anomaly; dot-dash line, wind stress curl anomaly (WSC). **Bottom**: Same as top, with dot-dash line being the local along-shore wind stress anomaly (LWS). All series are smoothed with a 6 months Hanning filter.
Figure 3.13. Clipper spectral analysis for north site. **Left-top**: model DWBC x LWS cross spectral density, vertical lines are error estimates, \(p\) is the peak’s mean period and \(f\) its frequency. **Left-center** model DWBC x LWS coherence, the dash-dot line marks the significant level. **Left-bottom**: model DWBC x LWS phase. **Right column**: same as left, but for DWBC x WSC cross spectrum.
CHAPTER 4

CONCLUSIONS

The barotropic method is capable of reproducing WBC deep velocity variability, taken as a proxy of the barotropic component, in a satisfactory manner for a series of tests using observed and synthetic data. While the comparisons are generally more favorable for the synthetic data, it is important to remember the observed values were not obtained in a way that best suites the method’s needs. We conclude this technique is a viable, operationally inexpensive option for studies concerned with barotropic flows.

Application of the method to areas in the Atlantic WBC generates two time series, each approximately 6 years in length. Temporal resolution range from 1 (north site) to 2.5 (south site) months. This is a much longer sampling period than most available WBC eulerian measurements, specially of the deep flows.

In both areas, current variability is significantly correlated to basin averaged wind stress curl and and also to local along-shore wind stress. The relationship between currents and wind curl is coherent with the WBC response to interior Sverdrup flow. We propose the local winds stress exerts control over the flow by Ekman layer convergence at the coast, in a physical mechanism similar to the barotropic shelf response to along-shore winds.

In the north (40°N), the variability is dominated by interannual oscillations of the wind curl. The effects of the local stress are secondary and have annual frequency. Both wind stress curl and along-shore wind are significantly correlated to the currents in the southern site (8°S), but local effects appears to be the dominant forcing. The main observed results are confirmed by data from a numerical model.
REFERENCES


BIOGRAPHICAL SKETCH

Álvaro Montenegro

Álvaro Montenegro was born in the city of Botucatu (SP, Brazil) on March 20, 1970. He graduated with a Bachelor of Oceanography degree from the University of Rio Grande (RS, Brazil) in January of 1997. In June of 1999 he obtained the degree of Master of Physical Oceanography from the University of São Paulo (SP, Brazil). Since August of 1999 he has been enrolled in the Physical Oceanography Graduate program at the Florida State University Department of Oceanography.