ABSTRACT

GONG, YANLIN. Impacts of Eddies, Topography, and Topographic Rossby Waves on North Brazil Deep Current Variability: an Observation and Modeling Synthesis Study. (Under the direction of Dr. Ruoying He.)

In situ observations of the north Brazil deep current, one made in May-December 2011 and the other in May 2011- May 2013, revealed drastic differences in bottom circulation. It is found that current measured by the offshore buoy (Orenoque B, 2000 m), showing persistent southwestward motion, is a part of the Deep Western Boundary Current (DWBC). In contrast, current measured 15 miles apart (Orenoque A, 1800 m) displays much larger flow oscillations and variability. The goal of this study is to understand specific roles of surface eddy/ring, topography and topographic Rossby waves (TRWs) in affecting such a bottom current variability.

Based on 14-year satellite altimeter data, a NBC ring census analysis is firstly carried out, showing that most (>50%) eddies in the region have around 130 km radius, 4-8 weeks lifespan, and >7 cm sea level anomaly. During the two-year buoy observation time, twelve surface eddy events are identified near the mooring location. Among them, there are six periods when the surface and bottom meridional currents are significantly correlated (abs(r) >0.60), suggesting surface eddies have an important impact on the bottom current. Aside from local eddy influence, analyses of power spectra of the slope mooring (Orenoque A) current data and a ray tracing calculation show the bottom current variability can also be introduced by TRW propagated from a remote site north of mooring location.

Idealized numerical model experiments are subsequently used to understand three-dimensional circulation associated with eddies. Both surface-trapped and deep-reaching eddies are considered over either a flat bottom or a sloping bottom setting. In the flat bottom
model simulation, significant negative (positive) correlation is found between surface and bottom meridional currents when surface-trapped (deep-reaching) eddy is present. Compared to the surface-trapped eddy, deep-reaching eddy is able to generate significantly larger deep current and material transport. In the sloping bottom experiment, model simulations show the ocean mixing is further enhanced over the slope at the bottom boundary layer and inside the eddy. Additionally, a strong bottom jet and TRWs develop south of the eddy center, resulting in a larger bottom transport on the slope. Adding a sloping bottom also generates complex sub-mesoscale circulation near the slope.

Finally, a realistic hindcast modeling approach, taking surface eddy, topography, boundary current and TRWs all into consideration, is used to simulate circulation during the mooring observing period. Model-data comparisons show the model is able to resolve most of current variability, but underestimates the strength and energy of the observed bottom current. Deterministic prediction of the deep circulation in the region clearly needs further model refinements, more extensive observations, together with advanced data assimilation schemes.
Impacts of Eddies, Topography, and Topographic Rossby Waves on North Brazil Deep Current Variability: an Observation and Modeling Synthesis Study

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To my beloved family.
BIOGRAPHY

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Chapter 1: Motivation

Mesoscale eddies, with diameters of tens to hundreds of kilometers and periods of weeks to months, are a common feature in the open ocean (e.g., Chelton et al., 2011a). They are known to influence the distribution of ocean physical, biological and biogeochemical properties (e.g., Chelton et al., 2011b). Surface eddies have a potential impact on the deep ocean environment during their propagation. Mooring observations showed that surface eddies influenced hydrothermal vent efflux transport at the East Pacific Rise (Adams et al., 2011) and deep sea sediment transport in the northern South China Sea (Zhang et al., 2014). In the Gulf of Mexico, intermittent deep intensified currents were observed and some showed an enhanced coherence between the upper and lower layer currents (Kolodziejczyk et al., 2012). On the continental slope, surface eddies’ impact on the bottom current became more complex due to topographic Rossby waves (TRWs) propagation and deep western boundary current (DWBC) variability (Johns et al., 1993).

1.1 Circulation in the Western Tropical Atlantic Ocean

In the western tropical Atlantic Ocean, the circulation is very complex (Figure 1.1). Near the surface, the North Brazil Current (NBC) flows northward along the northeastern coast of South America. When it reaches French Guiana (4° N, 53° W), part of the NBC separates from the coast and retroflects to join the North Equatorial Counter Current (NECC), while the rest of the NBC continues flowing northwestward to form the Guiana Current (GC, Condie, 1991). At the same time, the Atlantic North Equatorial Current (NEC) flows westward, approaches the shelf region of the Americas, and bifurcates, with a portion of its transport turning back to the east after being entrained into the NECC, and the other
portion continuing to flow westward (Bourles et al., 1999). Then the NEC and GC couple together and enter into the Caribbean Sea and feed the Caribbean Current (CC) (Johns et al., 2002). In this region, the thermocline is located at around 200 m (Figure 1.2) and salinity has the maximum value of 36.5 psu at 100 m. The squared Brunt-Vaisala frequency ($N^2$) ranges from $2.5 \times 10^{-4}$ $s^{-2}$ at upper 200 m and to $1.5 \times 10^{-6}$ $s^{-2}$ at 2000 m. The first baroclinic Rossby radius of deformation in this region is around 80-150 km (Chelton et al., 1998).

This region is dominated by large anticyclonic rings/eddies called NBC Rings. They are shed by the retroreflection of the NBC (John et al., 1990), and move northwestward toward the Caribbean Sea on a course parallel to the South American coastline (Goni and Johns, 2001). After translating northwestward for several months, NBC rings decompose in the vicinity of the Lesser Antilles (Fratantoni et al., 2006). The physical properties and evolutions of NBC rings have been investigated with extensive hydrographic and direct-velocity survey cruises, current meter arrays and inverted echo sounders, and satellite observations during the NBC Rings Experiments (e.g. Goni and Johns, 2001; Wilson et al., 2002; Johns et al., 2003; Fratantoni et al., 2006). Using six years of TOPEX/POSEIDON (T/P) altimetry, Goni and Johns (2001) gave a census of the NBC rings’ properties. They estimated that an average of more than five rings and eddies were generated per year, with the translation speed of 14 km/day and mean length scale of approximately 100 km. Wilson et al. (2002) described three fundamentally different types of NBC ring structures (Figure 1.3): (1) a shallow, surface-trapped structure with velocities confined to the top 200 m (Ring 3 and Ring 4), (2) a deep-reaching structure with significant swirl velocities extending to 2000 m (Ring 2), and (3) a thermocline-intensified structure with almost no detectable
surface signature (Ring 1). Numerical modeling has been successful in producing NBC rings and studying NBC rings’ properties. For example, using the Miami Isopycnic Coordinate Ocean Model (MICOM), Garraffo et al. (2003) classified the NBC rings into ‘shallow’, ‘intermediate’, ‘deep’, and ‘subsurface’ categories. They showed that NBC rings contributed 40% of the total meridional transport from the surface to the intermediate water layers.

Near the bottom, circulation in the western tropical Atlantic Ocean is dominated by the Deep Western Boundary Current (DWBC). The DWBC transports North Atlantic Deep Water from the overflow regions between Greenland and Scotland and from the Labrador Sea into the South Atlantic and Antarctic circumpolar currents. The DWBC acts as a primary pathway for the cold, lower limb of the meridional overturning circulation (MOC), and its transport variability has been well-documented in several observation networks in the North Atlantic Ocean (e.g. Kanzow et al., 2008; Peña-Molino et al., 2012; Meinen et al., 2013). By analyzing Line W array data, Peña-Molino (2012) showed that fluctuations of the DWBC with periods from 10 to 60 days were bottom-intensified and associated with TRWs. A fraction (~15%) of flow anomalies were surface-intensified and fluctuated at frequencies lower than the TRWs, which were caused by Gulf Stream rings and meanders. Furthermore, interannual variability in the velocity field was related to changes in the hydrographic properties of the deep Labrador Sea water.

1.2 Eddy Dynamics and its Impact on Deep Circulation

Mesoscale eddies are a common feature in the ocean. They are characterized by temperature and salinity anomalies with associated rotational flow anomalies. Large eddies in
the Atlantic Ocean include Gulf Stream warm-core rings, Loop Current eddies, and NBC rings.

The dynamics of large isolated eddies has been extensively studied using numerical models. In the open ocean, the propagation and decay of isolated eddies are influenced by eddy size, strength and vertical structure (McWilliams and Flierl, 1979). The anticyclonic (cyclonic) eddies generally move in the southwest (northwest) direction in the northern hemisphere under the combined planetary β and nonlinear effects. The westward propagation of a feature is a consequence of planetary β, while the rate of southward (northward) propagation of an anticyclonic (cyclonic) vortex is determined by the strength of the current (McWilliams and Flierl, 1979; Mied and Lindemann, 1979). Under different dynamic frameworks, eddies show different features during their propagation (Firing and Beardsley, 1976). When eddies are governed by linear dynamics, they disperse and their energy is radiated away by Rossby waves. Under nonlinear dynamics, barotropic or nearly barotropic cyclonic eddies, which have the same or similar upper layer and lower layer velocities, become elongated on their western side, and an anticyclonic feature forms to the east of original vortex. Eddies evolve into different structures in the upper layer and lower layer (Mied and Lindemann, 1979). Upper ocean eddies are stable, while deep ocean eddies exhibit rapid barotropic-like dispersion, and no vortex is generated in the lower ocean that connects with the upper ocean.

Since eddies primarily move westward, they will eventually reach the regime of the continental slope or western boundary current. Model studies of eddy–slope interaction have yielded insight on the behavior of eddies when encountering the continental slope. Eddies on
the slope tend to evolve into upper-layer features quickly, and lower-layer features are obliterated through radiation of TRWs (LaCasce, 1998). Nof (1999) showed that, in the presence of vertical-walled boundaries, eddy migration was governed by three processes: the image effect, under which eddies move northward; the β-induced self-advection effect, under which eddies move southward; and the rocket effect, under which eddies move northward. The image effect is the most dominant factor in determining the final migration along the wall. Furthermore, eddy-topography interactions often produce smaller-scale vortices and generate coastally trapped waves (e.g., Robinson, 1991; Wei and Wang 2009; Sutyrin and Grimshaw, 2010; Akuetevi and Wirth, 2015). Sutyrin and Grimshaw (2010) showed that secondary cyclonic eddies were generated due to off-shelf advection of water with high potential vorticity under the inviscid setting. The viscous response at smaller scales implies the formation of a frictional bottom boundary layer on the continental slope, where strong lateral shears develop and shear instability occurs, generating subsurface vortices (Oey and Zhang, 2004; Molemaker et al., 2015). Using both idealized and realistic simulations, Vic et al. (2015) revealed the formation of intense frictional boundary layers and generation of submesoscale coherent vortices, leading to Persian Gulf Water shedding in the Gulf of Oman.

From observation and numerical simulation studies, surface eddies were shown to have a potential impact on the deep ocean environment during their propagation. Early studies in the Synoptic Ocean Prediction Experiment showed that strong cyclones in the deep ocean were generated beneath large amplitude Gulf Stream meander troughs and rings (Luyten, 1977). Recent in situ observations demonstrated an unexpected impact of surface-
generated mesoscale eddies on the transport of hydrothermal vent efflux at the deep East Pacific Rise (Adams et al., 2010), and transport of deep-sea sediments in the South China Sea (Zhang et al., 2014). South China Sea results showed that the surface and bottom currents had a positive correlation coefficient, while East Pacific Rise results showed the correlation was negative. Figure 1.4 shows the instance of a strong negative correlation (Ru = −0.56, Pu < 0.0001; Rv = −0.86, Pv < 0.0001 with an 8-day lag) between a current anomaly observed on the ridge crest and inferred geostrophic velocities at the surface detected from May through June 2007 at the East Pacific Rise. In the Gulf of Mexico, intermittent deep intensified currents were observed and some showed an enhanced coherence between the upper and lower layer currents (Kołodziejczyk et al., 2012). Numerical simulations also showed that surface eddies could impact bottom currents to some extent. For example, using a quasi-geostrophic model, Adams and Flierl (2011) showed coherent cyclones developed in the deep layer and the dispersal potential of passive particles was significantly increased. Eddy interaction with ridge topography further enhanced tracer dispersal along the ridge axis through shearing and elongation of the eddy core.

1.3 Topographic Rossby Waves

Topographic Rossby Waves (TRWs) are transverse, quasi-geostrophic waves found in regions of sloping topography (Rhines, 1970). TRW dispersion can be derived from the primitive ocean-governing equations as follows. Assume a uniform sloping bottom defined by \( z = -H + \alpha y \), where \( z \) is the depth and \( \alpha \) is the bottom slope. The right-handed coordinate set is \( (x, y, z) \), with \( z \) directed vertically upwards; \( p \) is the pressure; \( \omega \) is the frequency of
the wave; k, l are the horizontal wave numbers in the x and y directions, respectively

\( K^2 = k^2 + l^2 \); f is the Coriolis parameter; and N is the Brunt-Vaisala frequency.

First, the linear, hydrostatic, Boussinesq equations are written as follows:

\[
\frac{\partial u}{\partial t} - fv = -\frac{1}{\rho_0} \frac{\partial p}{\partial x}
\]

\[
\frac{\partial v}{\partial t} + fu = -\frac{1}{\rho_0} \frac{\partial p}{\partial y}
\]

\[0 = -\frac{1}{\rho_0} \frac{\partial p}{\partial z} - \frac{g \rho}{\rho_0}
\]

\[
\frac{\partial \rho}{\partial t} + w \frac{\partial \rho}{\partial z} = 0
\]

\[
\frac{\partial u}{\partial x} + \frac{\partial v}{\partial x} + \frac{\partial w}{\partial z} = 0
\]

From these equations we can derive a single governing equation:

\[
\frac{\partial}{\partial t} \left( \frac{\partial^2 p}{\partial x^2} + \frac{\partial^2 p}{\partial x^2} + \frac{f^2}{N^2} \frac{\partial^2 p}{\partial z^2} \right) = 0
\]

where \( N^2 = -\frac{g}{\rho_0} \frac{\partial \rho_0}{\partial z} \)

The boundary conditions are described as follows:

at \( z = 0 \), \( w = 0 \) and we get \( \frac{\partial}{\partial t} \frac{\partial p}{\partial z} = 0 \)

at \( z = -H + \alpha y \), \( w = \alpha v \) and we get \( \frac{1}{N^2} \frac{\partial}{\partial t} \frac{\partial p}{\partial z} = -\alpha \frac{\partial p}{\partial x} \)

Wave solution is assumed and the solution has the form:

\[
p(x, y, z, t) = p(z) \exp(i(kx + ly - \omega t))
\]
The vertical distribution $p(z)$ is the solution of an eigenvalue equation. If the stratification is uniform, the eigenvalue equation has a solution as follows:

$$p(z) = p_0 \cosh\left(\frac{NKz}{f}\right)$$

$$\omega = -N\alpha \frac{k}{K} \coth\left(\frac{NHK}{f}\right)$$

From this solution, we can see that as $z$ increases, the magnitude of $p$ also increases, so TRWs are bottom-intensified. From the dispersion relation equation, we see that the maximum frequency or cutoff frequency is $N\alpha$.

TRWs play an important role in the deep current field on the continental slope. Numerous moored arrays along the continental slope of the Mid-Atlantic Bight (MAB) (e.g., Hogg, 1981, Hamilton, 1984) and Gulf of Mexico (e.g., Hamilton, 1990, Hamilton, 2009) indicated that current variability below 1000 m is dominated by TRWs, since current variability conformed to TRW theory of high coherence through the lower water column, bottom intensification, and period agreement with the dispersion relation. The observed periods of TRWs in the MAB range from 8 to 60 days and wavelengths range from 40 to 300 km (Hogg, 1981). In the Gulf of Mexico, TRWs have characteristics of periods ranging from 10 to 100 days and wavelengths ranging from 50 to 200 km, and are almost depth-independent or slightly bottom-intensified in the weakly stratified lower water column (Hamilton, 2009).

The generation mechanism for TRWs is complex. In the MAB, the source of the waves is believed to be the deep Gulf Stream, because the observed phase propagation of the waves is offshore and so the energy propagation would be onshore. Several generation
mechanisms are proposed: (1) Louis and Smith (1982) presented evidence that bursts of TRWs observed off Nova Scotia were associated with the formation of warm Gulf Stream eddies. Furthermore, Shaw and Divakar (1991) suggested TRWs can be generated when there is lower-layer circulation in a ring. Such a deep circulation can be produced by adjustment of the velocity field to the density structure if a ring was not initially in geostrophic balance. (2) Pickart (1995) showed wavenumber coupling between propagating surface features such as Gulf Stream meanders and deep flows as a generation mechanism for TRWs; and (3) Hogg (2000) indicated TRWs were generated by broadband radiation of the Gulf Stream and rings. In the Gulf of Mexico, TRWs were suggested to be generated through energy transfer from upper-layer eddies to the lower layer by potential vorticity adjustments as a result of changing depths of the bottom and the interface between upper layer and lower layer (Hamilton, 2009). Model simulations also indicated that TRWs and deep lower-layer eddies may be generated by the Loop Current (LC) and LC eddy shedding processes (Oey and Lee, 2002). In a 3-dimensional numerical modeling study, Oey et al. (2009) put temperature and velocity anomalies on the ocean bottom (Figure 1.5), and the pressure produced strong localized eddies. Away from the source, TRWs were excited and disturbances propagated north–northwestward and along-isobaths. They also investigated the focusing and accumulation of TRW energy by slopes coupled with bends in the isobaths.

TRW ray tracing method is described and used to determine the TRWs’ energy path by the following steps (Meinen et al., 1993). The method is based on the TRWs’ governing equations and dispersion relation, which describe the dependence of frequency on environmental parameters such as bottom slope and water depth. First, the initial values for
position, wavelength, and frequency are chosen and the corresponding group velocity is computed. Then the wave packet is computed using group velocity and a short time interval to get the new location. At the new location, the method computes the new wavelength, frequency and group velocity, and the wave packet progresses again. This computation repeats and TRW ray path is obtained. Pickart (1995) used this method and found TRWs in the MAB were generated by the deep Gulf Stream.

1.4 Cold Seeps and the North Brazil Deep Current

Though light only penetrates the upper layers of the water column, high local diversity and low-density biomass ecosystems exist on the deep-sea floor at hydrothermal vents and cold seeps. These systems are largely driven by chemosynthetically derived energy, without direct dependence on photosynthesis (Vanreusel et al., 2010). Cold seeps are places where reduced sulphur and methane emerge from seafloor sediments without an appreciable temperature rise (Levin, 2005). Cold seeps were first found on the Florida Escarpment in the Gulf of Mexico in 1983 (Paull et al., 1984), and since then have been discovered in other parts of the world's oceans. Figure 1.6 shows map of cold seep locations in the Atlantic Equatorial Belt.

To investigate connectivity in western Atlantic seep populations, the deep sea survey OC 472-2 was conducted during May 17-20, 2011, and two bottom current meter moorings (Orenoque A and Orenoque B, Figure 1.1) were deployed in the North Brazil basin in May, 2011. Each mooring was equipped with an Acoustic Current Meter (ACM) with a sampling interval of 3 hours. Orenoque A and B are located at the 1800 m and 2000 m isobath,

Though the distance between two moorings is only 30 km, both their mean current and current variability are very different. Figure 1.7 shows the bottom current variability observed by the two moorings and their inferred particle trajectories. The inertial period at the mooring location is 67 hours, so we applied 72-hr low-pass filter to the current data to exclude tidal and inertial wave signals. At mooring A, the mean current in the meridional direction is -0.7 cm/s, with the standard deviation of 7.4 cm/s. At mooring B, the mean current in the meridional direction is -8.7 cm/s, with the standard deviation of 4.7 cm/s. Mooring B shows a strong persistent southward bottom current, as expected for the DWBC along the continental slope, while mooring A shows a very weak mean current with large variations. Bottom currents at these moorings have different potential impact on the transport of particles at local cold seeps. A simple progressive vector diagram calculation shows that a particle will move back and forth at mooring A, while it will advect southward quickly at mooring B. The correlation coefficient for the zonal current component of these two moorings is 0.19, and for the meridional current component is only -0.03. Therefore, the mechanism of bottom current variability is different at these two mooring locations.

1.5 Research Objectives

From this background review, we see that North Brazil bottom currents are complex and several processes can impact their variability, such as NBC rings and TRWs. In addition, topography plays an essential role in sustaining or impacting these processes. The overall objective of this Ph.D. research is to better understand bottom velocity observations, and by
doing so, seek a deeper dynamical insight on impacts of eddies, slope topography, and topographic Rossby waves on North Brazil deep current variability.

This dissertation study has several significant research facets. First, eddy-topography interaction is a fundamental research topic in physical oceanography. Second, studying eddy/ring properties in the north Brazil region will deepen the understanding of circulation dynamics in the tropical western Atlantic Ocean. Third, understanding deep current variability can provide crucial insight about deep sea environment and marine ecology (such as cold seeps ecosystem and sediment transport).

Previous studies have addressed several scientific questions about eddies’ and TRWs’ impact on bottom currents. However, there are still elements that have not well-resolved. For example, whether eddies can significantly impact bottom currents has not reached a convincing conclusion, and the difference between surface-trapped eddies’ and deep-reaching eddies’ impact on bottom currents is unknown. Previous model studies used a two-layer model to study eddies’ impact on bottom currents, but impact from the three-dimensional baroclinic eddy structures is not well addressed. TRW studies have mainly focused on the MAB and Gulf of Mexico region, but little is known about their properties in the western tropical Atlantic Ocean. Therefore, in this dissertation study, the following three central scientific questions will be investigated about the north Brazil deep circulation:

Q1: How do surface eddies influence bottom current variability?

Q2: What role does the continental slope play in the presence of eddies’ impact on the bottom current?

Q3: Can bottom current variability be introduced by TRWs?
To study these three questions, the following hypotheses are proposed:

H1: Convergence/divergence introduced by surface eddies can lead to bottom current variability through vortex stretching and mass conservation. Deep currents are affected more significantly by deep-reaching eddies than by surface-trapped eddies.

H2: Stronger mixing is generated in the presence of both surface eddies and continental slope. TRWs and submesoscale features develop and lead to enhancement of the bottom current variability under the eddy impact.

H3: In addition to local processes, energy propagated from remote sites through topographic Rossby waves also contributes to bottom current variability.

In this dissertation study, both data analysis and numerical modeling will be used to test hypotheses and address questions. The numerical modeling includes both idealized modeling and realistic modeling. Several numerical experiments are designed and performed to study the dynamics of eddies’ impact on bottom current variability. The structure of this dissertation is as follows: in Chapter 2, we utilized observation data to study impacts of eddies, topography, and topographic Rossby waves on North Brazil deep current variability. In Chapter 3, idealized numerical modeling experiments are used to study impacts of eddies, topography, and topographic Rossby waves on North Brazil deep current variability. In Chapter 4, a realistic hindcast modeling approach, taking surface eddy, topography, boundary current and TRWs all into consideration, is used to simulate circulation during the mooring observing period. Chapter 5 provides a summary of this dissertation.
References


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Chapter 2: Impacts of Eddies, Topography, and Topographic Rossby Waves on North Brazil Deep Current Variability: 1. Observation Analysis

Abstract:

Two in situ observations of the north Brazil current (NBC) were made in May-December 2011 and May 2011- May 2013, and revealed drastic differences in bottom circulation. It is found that current measured by the offshore buoy (Orenoque B) is a part of the Deep Western Boundary Current (DWBC), showing persistent southwestward motion. In contrast, current measured 15 miles apart (Orenoque A) displays much larger variability. Based on 14-year satellite altimeter data, an eddy census analysis shows that most (>50%) eddies in the region have around 130 km radius, 4-8 weeks life spans, and >7 cm sea level anomaly. During the two-year observation period, twelve eddy periods passing mooring sites are identified. In four (two) periods, correlation coefficients between surface and bottom meridional current at Orenoque A are significant and larger (smaller) than 0.6 (-0.6). Therefore, surface eddies potentially have a direct impact on the deep ocean circulation. In addition to impact from local eddy, power spectra analyses on slope mooring (Orenoque A) current data and ray tracing calculation show the bottom current variability can be introduced from a remote site north of mooring location via the TRW propagating along the slope.
2.1 Introduction

North Brazil Current (NBC) rings are large, anticyclonic eddies shed by the North Brazil Current, an intense western boundary current and the dominant surface circulation feature in the western tropical Atlantic Ocean. At the place of NBC retroflection, NBC rings are shed and move northwestward toward the Caribbean Sea on a course parallel to the South American coastline (Goni and Johns, 2001). After translating northwestward for several months, NBC rings decompose in the vicinity of the Lesser Antilles (Fratantoni et al., 2006).

Near the bottom, circulation in the western tropical Atlantic Ocean is dominated by the Deep Western Boundary Current (DWBC). The DWBC acts as a primary pathway for the cold, lower limb of the meridional overturning circulation (MOC), and its transport variability has been well-documented in several observation networks in the North Atlantic Ocean (e.g. Kanzow et al., 2008; Peña-Molino et al., 2012; Meinen et al., 2013).

The physical properties and evolution of NBC rings have been investigated with extensive hydrographic and direct-velocity survey cruises, current meter arrays and inverted echo sounders, and satellite observations during the NBC Rings Experiments (e.g. Wilson et al., 2002; Johns et al., 2003; Fratantoni et al., 2006). Wilson et al. (2002) showed three fundamentally different types of NBC ring structures: (1) a shallow, surface-trapped structure with velocities confined to the top 200 m, (2) a deep-reaching structure with significant swirl velocities extending to 2000 m, and (3) a thermocline-intensified structure with almost no detectable surface signature. Using six years of TOPEX/POSEIDON (T/P) altimetry, Goni and Johns (2001) gave a census of NBC ring properties, such as the average ring number per year, trajectories, and translation speed.
Surface eddies have a potential impact on the deep ocean environment during their propagation. Mooring observations show that surface eddies influence hydrothermal vent efflux transport at the East Pacific Rise (Adams et al., 2011) and deep-sea sediment transport in the northern South China Sea (Zhang et al., 2014). In the Gulf of Mexico, intermittent deep intensified currents are observed and some events show an enhanced coherence between the upper and lower layer (>1000 m) currents (Kolodziejczyk et al., 2012).

Topographic Rossby Waves (TRWs) play an important role in the deep current field on the continental slope. Numerous moored arrays along the continental slope of the Mid-Atlantic Bight (MAB) (e.g., Hogg, 1981; Hamilton, 1984) and Gulf of Mexico (e.g., Hamilton, 1990; Hamilton, 2009) indicated that current variability below 1000 m is dominated by TRWs, since current variability conformed to TRW theory of high coherence through the lower water column, bottom intensification, and period agreement with the dispersion relation. The observed periods of TRWs in the MAB range from 8 to 60 days and wavelengths range from 40 to 300 km (Hogg, 1981). In the Gulf of Mexico, TRWs have characteristics of periods ranging from 10 to 100 days and wavelengths ranging from 50 to 200 km, and are almost depth-independent or slightly bottom-intensified in the weakly stratified lower water column (Hamilton, 2009).

Considering NBC rings and eddies exist all year in the western tropical Atlantic Ocean, whether surface eddies have a similar influence on the deep ocean current there is unknown. Since sea surface height (SSH) fields of the merged altimetry data from T/P, Jason-1, Jason-2 and Envisat et al. reveal a fundamentally different perspective on SSH variability from the low-resolution SSH fields from T/P data alone (Chelton et al., 2011), an
updated and more detailed investigation of NBC ring surface properties will be presented using the most recent 14 years (2000–2013) merged altimetry data. Accordingly, the eddy vertical temperature and salinity structure will be investigated using Argo float data during this period. Previous TRW studies have focused on the MAB and Gulf of Mexico regions; little is known about their properties in the western tropical Atlantic Ocean. In this chapter, we will investigate impacts of eddies, topography, and TRWs on North Brazil deep current variability based on observation data. The rest of this chapter is organized as follows: section 2.2 describes the data and methods. Section 2.3 presents eddies’ impact on the bottom current variability. Section 2.4 provides a census study of eddy properties. Section 2.5 investigates the TRW signal from the mooring observation. Section 2.6 is the discussion and conclusions.

2.2 Data and Methods

Two bottom current meter moorings (Orenoque A and Orenoque B), shown in Figure 1.1, were deployed southeast of Barbados in May 2011. Each mooring was equipped with a Falmouth Scientific 2-Dimensional Acoustic Current Meter (2D-ACM) with a sampling interval of 3 hours. The height of observation above bottom was 2 m. Orenoque A and B were located at the 1800 m and 2000 m isobath, respectively, and the distance between these two moorings was only 30 km. Orenoque B mooring collected data from May 20 to December 20, 2011, and Orenoque A mooring collected data from May 20, 2011 to May 8, 2013.

To investigate the surface properties of NBC rings and cyclonic eddies, we used gridded sea level anomaly (SLA) fields constructed by SSALTO/DUACS with 7 day intervals and 1/4°×1/4° resolution distributed by AVISO project
(http://www.aviso.oceanobs.com) from January 2000 to December 2013. To investigate the vertical temperature and salinity structure of NBC rings and cyclonic eddies in this region, we used autonomous CTD profiling floats from the Argo program (http://www.argo.ucsd.edu) over the same time as the altimetry observations, and World Ocean Atlas 2013 (WOA13) climatology data (Locarnini et al., 2013; Zweng et al., 2013).

We use two methods to study eddy properties: the eddy identification and tracking method, and the eddy composite analysis method. The TRW ray tracing method is used to determine the TRW energy path.

Several methods are available for identifying eddies from satellite altimetry (Kurian et al., 2011), including physics-based methods, which are based on closed contours of a threshold value of SSH anomaly, vorticity, velocity gradient tensor (Q), or the Okubo-Weiss parameter (W); geometrical methods; and other methods based on wavelets or Lagrangian coherent structures. In this work, we used a new eddy identification method (Chelton et al., 2011). Since there is no need for differentiation of the SLA field, this method result is smoother than Q and W methods. For each mapping of SLA, the eddy detection method finds regions (i.e., a set of connected pixels) that contain a local maximum (minimum) in SLA and then identifies pixels in the region which are above (below) a given SLA threshold as anticyclonic (cycloonic) eddies. We set the parameters so that more than 6 and fewer than 600 pixels comprise one eddy, the amplitude of an eddy is at least 3 cm, and the distance between any pair of points within an eddy must be less than 300 km. These values are similar with criteria shown in Chelton et al. (2011), with minor adjustment to avoid detecting eddies which enclose elongated features or broad amoeba-like regions.
After eddies are identified for each time step in the sequence of SLA mapping, an automated tracking procedure (Chelton et al., 2011) is applied to determine the trajectory of each eddy. In this method, for each eddy identified at time step k, eddies identified at the next time step k + 1 are searched to find the closest eddy lying within a restricted search region. The restricted search region in this study is 150 km in both the east-west and south-north directions. Only eddies with amplitude and area that fall between 0.25 and 2.5 times of that in the k step are considered in the k+1 step. These values are also based on criteria shown in Chelton et al. (2011) to obtain the trajectory of the same eddy at different date and avoid connecting different eddy together.

Following Chaigneau et al. (2011), eddy vertical structure was constructed using the eddy composite analysis method. First, we obtained the good quality Argo data and discarded suspicious Argo profiles. The criteria are as follows (Chaigneau et al. 2011): (1) the shallowest data are located between the surface and 10 m depth and the deepest acquisition is below 1000 m; (2) the depth difference between two consecutive data does not exceed a given limit which depends on the considered depth (25 m for the 0–100 m layer; 50 m for the 100–300 m layer; and 100 m for the 300–1000 m layer); and (3) the number of data levels in the 0–1000 m layer is higher than 30. After obtaining good quality float data, we calculated the temperature and salinity anomaly for each profile by subtracting the WOA13 climatology field. Third, if the Argo profile was located inside an eddy boundary identified and tracked from the AVISO data, then we extracted its temperature and salinity anomaly and calculated its distance from the eddy center. Last, we averaged temperature and salinity anomaly as a function of depth and distance from the eddy center. By doing so, the eddy vertical
temperature and salinity structure were constructed. The surface properties and vertical structure of rings are well-studied in the South Atlantic Bight (SAB) and the Gulf Stream recirculation region (Castelao and He, 2013; Castelao, 2014), and results of NBC rings and cyclonic eddies will be compared with those from the SAB.

The TRW ray tracing method (Meinen et al., 1993) is used to determine the TRW energy path by the following steps. First, the initial values for position, wavelength, and frequency are chosen and the corresponding group velocity is computed. Then the wave packet is computed using group velocity and a short time interval to get the new location. At the new location, the method computes the new wavelength, frequency and group velocity, and the wave packet progresses again. This computation repeats and TRW ray path is obtained. Pickart (1995) used this method and found TRWs in the MAB were generated by the deep Gulf Stream. We will use this method to calculate the TRW ray paths in the western tropical Atlantic Ocean.

2.3 Eddy and Bottom Current Observation Results

Though the distance between two moorings is only 30 km, both their mean current and current variability are very different. Figure 1.7 shows the bottom current variability observed by the two moorings and their inferred particle trajectories. At mooring A, the mean current in the meridional direction is -0.7 cm/s, with the standard deviation of 7.4 cm/s. At mooring B, the mean current in the meridional direction is -8.7 cm/s, with the standard deviation of 4.7 cm/s. Mooring B shows a strong persistent southward bottom current, as expected for the DWBC along the continental slope, while mooring A shows a very weak mean current with large variations. The correlation coefficient for the zonal current
component of these two moorings is 0.19, and for the meridional current component is only -0.03. Therefore, the mechanism of bottom current variability is different at these two mooring locations.

During the observation period, several anticyclonic and cyclonic eddies passed over the two moorings (Figure 2.1). Using the eddy detection and tracking procedures, we detected five eddies that passed the mooring location during the first six months period. We calculated the eddy properties and found that the first anticyclonic eddy was stronger, with mean radius over 150 km and amplitude over 13 cm, while the cyclonic eddies and other anticyclonic eddies were much weaker, with radius around 100 km and amplitude around 7 cm. Another feature of these eddies is that the first anticyclonic eddy decayed quickly after passing the mooring location, while the other anticyclonic eddies remained observable for at least another seven days.

To study the relationship between surface eddies and bottom currents, we applied a 72 hr low-pass filter to the daily AVISO SLA data and interpolated the 3 hourly mooring current data into daily data. Figure 2.2 shows the time series of sea level anomaly, surface geostrophic velocity anomaly and bottom current anomaly at the mooring A location. The surface geostrophic velocity was calculated from SLA based on geostrophic balance equation, and applied to the mooring location using the nearest AVSIO gridded point data. Since eddies could have a combined effect on sea level and surface velocity variability, i.e., the departure of the anticyclonic eddy and arrival of the cyclonic eddy would induce sea level decrease and a southward surface velocity anomaly, we combined eddies into twelve periods so that each period had a whole cycle of increasing/decreasing SLA and
southward/northward variability of surface velocity. The length of each period varied from four weeks to eleven weeks. Correlation coefficients between the surface current and bottom current are shown at Table 2.1. In most periods, correlation coefficients in the zonal direction were much smaller than that in the meridional direction. This is probably due to the meridional orientation of the bathymetry constraining eddies’ effect on bottom zonal current variability. For the meridional component, correlation coefficients in four periods were significant at the 95% confidence interval and greater than 0.6, indicating eddies potentially had a strong positive impact on deep current. In another two periods, correlation coefficients were significant and smaller than -0.6, indicating eddies potentially had a strong negative impact on deep current. In another two periods, correlation coefficients were significant but the magnitude was relatively small. In other four periods, correlation coefficients were not significant at the 95% confidence interval. This shows a much more complex situation compared with that in the East Pacific Rise region (Adam et al., 2011), where the correlation coefficient was significantly negative between surface current and bottom current.

There are some aspects difficult to explain solely with surface eddies’ impact on bottom current variability, such as during the period in November 2011, there was strong bottom current variability without eddy impact. Since the mooring was located on the continental slope, we expect that the bottom current is also impacted by other factors such as TRWs.

2.4 Eddy Property Census

In the following, we discuss the eddy property census using the AVISO data and Argo data. Studying eddy properties can not only provide an update of eddy census given by
Goni and Johns (2001), but also provide a basis for designing eddy structure in the numerical modeling study.

2.4.1 Eddy Trajectories

NBC rings are anticyclonic eddies shed at NBC retroflection, move northwestward toward the Caribbean Sea and decompose in the vicinity of the Lesser Antilles. Figure 2.3 shows trajectories and spatial distribution of both anticyclonic and cyclonic eddies with lifespans of no less than four weeks tracked by the automated eddy tracking procedure. Though most previous studies focused on anticyclonic NBC rings, cyclonic eddies were also a common feature in this region. From the eddy trajectories, we can see there are three regions with a high eddy density: the western part, located in the Caribbean Sea; the north part, located north of 15°N and east of the Lesser Antilles; and the south part, located south of 15°N and east of Lesser Antilles. Since we focused on the properties of NBC rings, only eddies in the southern area will be analyzed. Furthermore, we also excluded eddies located shoreward of the 200 m isobath because altimetry data have much larger error on the shelf than in the deep ocean.

From eddy trajectories, we can see both anticyclonic NBC rings and cyclonic eddies tend to propagate toward the west. Eddy trajectory clusters are separated by the Lesser Antilles, indicating that eddies decay and are destroyed after reaching the Lesser Antilles. Eddy number differs greatly spatially. Most eddies occur seaward of the 2000 m isobath. Eddies located between 60°W and 50°W are closer to the slope, while eddies between 50°W and 40°W spread out in the open ocean. The total number of individual anticyclonic (cyclonic) eddies identified from 2000 to 2013 is 1571 (1683). This number includes
repetition of the same eddy over multiple time steps. Counting each NBC ring only once over the course of its lifetime, the total number of anticyclonic (cyclonic) eddies identified from 2000 to 2013 is 208 (230), and the average number of both anticyclonic and cyclonic eddies is about 15 per year. This number is much larger than from a previous study that showed an average of 5 rings per year (Goni and Johns, 2001). The major reason is that Goni and Johns (2011) focused on the long-lived eddies, with the average lifespan of NBC rings in their study of 3.8 months (~16 weeks), whereas in our study, the average is ~ 7.5 weeks per track.

2.4.2 Eddy Radius, Lifespan and Amplitude

The probability distributions of the radius, lifespan, and amplitude of the anticyclonic and cyclonic eddies are shown in Figure 2.4. The radius of an eddy is defined as the radius of a circle with the area equal to the area of the closed ring boundary. The amplitude of an eddy is defined as the difference between the maximum SLA within the ring and the averaged height of the ring boundary.

Anticyclonic (cyclonic) eddy radius ranges from 50 km to 200 km, with 79% (72%) of the ring radius falling between 100 km and 175 km. The probability distribution is relatively symmetric. The mean anticyclonic (cyclonic) eddy radius is 130 (124) km and the peak value is between 125 and 150 km.

Since our criterion for identifying eddy is a lifespan of no less than 4 weeks, all ring lifespans are larger than 4 weeks. Some ring lifespans are as long as 28 weeks, while 86% (87%) of the lifespans of anticyclonic (cyclonic) eddy are shorter than 12 weeks. The mean anticyclonic (cyclonic) eddy lifespan is 7.6 (7.3) weeks. Unlike that of the ring radius, the probability density distribution of the lifespan is skewed to the right.
All eddy amplitudes are smaller than 0.3 m and larger than 0.03 m, and about 84% (90%) of measured anticyclonic (cyclonic) eddy amplitudes are smaller than 0.1 m. Similar to that of ring lifespan, the probability density of eddy amplitude decreases as ring amplitude increases. The mean anticyclonic (cyclonic) eddy amplitude is 7 (6.2) cm and is relatively small, because this detection procedure strives to identify eddies as compact mesoscale features (Chelton et al., 2011).

In summary, the probability distributions of the radius, lifespan, and amplitude of the anticyclonic and cyclonic eddies are similar, though the anticyclonic eddies have a little larger radius, longer lifespan and larger amplitude.

2.4.3 Ring Movement

Rings move primarily westward due to the planetary β effect. Figure 2.5 shows the spatial distribution of meridional velocity and zonal velocity averaged in the 1°×1° pixel grid. We only averaged the velocity in the pixels that contained at least five individual eddies in order to reduce random error. The mean zonal velocity of anticyclonic (cyclonic) eddy is -10 (-10) cm/s, while the mean meridional velocity is 3.8 (1.6) cm/s. The mean total velocity of anticyclonic (cyclonic) eddy is 11.7 (10.1) cm/s or 10 (8.7) km/day, smaller than previous remote sensing results of 14 km/day (Goni and Johns, 2011). There is no obvious relationship between the latitude of eddy location and westward propagation velocity. However, westward propagation velocity is faster in the 60°W-50°W region than in 50°W-40°W. A strong northward velocity close to the Lesser Antilles shows that both anticyclonic and cyclonic eddies are deflected northward due to coastline and topographic effects.

2.4.4 Vertical Temperature and Salinity Structure
The altimetry data are mainly used to study eddy surface structure. By combining them with Argo float data, we can investigate eddy vertical temperature and salinity structure. The Argo spatial distribution is more or less uniform in the western tropical Atlantic Ocean, while temporal distribution varies a lot, with many more Argo floats during 2009-2012 than other years. Because the Argo data are too sparse, temperature and salinity anomaly readings are not smooth (Figure 2.6). The maximum temperature anomaly of anticyclonic (cyclonic) eddy is around $5^\circ$C ($-5^\circ$C), and maximum salinity anomaly is around 0.5 (-0.5) psu. The largest temperature and salinity anomalies are located at 200 m depth at the eddy center. The anomaly decreases with larger distance from the eddy center, though this feature of the cyclonic eddy temperature is less obvious. The depth of maximum temperature and salinity anomaly is much shallower than that reported for the SAB and Gulf Stream Recirculation region (Castelao, 2014), because the thermocline depth is shallower in the tropical ocean than that in the subtropical ocean.

2.5 TRW Signals from Mooring Observations

To study whether TRW signals exist on the continental slope in the western tropical Atlantic Ocean, we first checked whether the wave energy spectrum satisfied the TRW dispersion relation. Figure 2.7 shows the variance-preserving spectra at mooring A, located in the North Brazil basin as shown in Figure 1.1. For the variance-preserving spectra, the area under the spectral curve between two frequencies represents the spectral signal variance in that frequency band. The energy spectra are similar between the u and v components. At mooring A, the variance is largest at 20~50 days and the peak occurs at around 30 days. The TRW cutoff period is calculated to be about 5.3 days, as we estimate the bottom slope $\alpha$ is
0.012 from ETOPO5 bathymetry data and the bottom water value of N is 1.2×10^{-3} s^{-1} from WOA13 climatology temperature and salinity data. Wave motions above the cut-off frequency are not supported by TRWs dynamics. At mooring A, the energy falls by an order of magnitude from the peak at 30-day period to periods shorter than 5.3 days. Therefore, an energy spectrum dominated by TRWs at mooring A is consistent with TRWs theory. At mooring B with the location shown in Figure 1.1, the spectra (not shown) focus at 3~10 days and the peak occurs at 3.5 and 6 days. Since the dominant frequencies at mooring B are higher than the TRW cutoff frequency, TRWs are not the dominant signal at mooring B.

In this area, TRWs would be expected to propagate from north to south along the isobath. In the following, we calculated the TRWs’ ray path to investigate the TRWs’ energy source. The bathymetry used for ray tracing was ETOPO5 data, smoothed using a filter width of 150 km to avoid small-scale undulations in the bottom. Then the topography was fitted with a series of B-splines, so it was much smoother than the realistic bathymetry. The B-splines were first fit to lines of constant latitude, then along lines of constant longitude, and a two dimensional B-spline expression was created for the bathymetry. Figure 2.7 shows the 30-day period TRWs’ ray path, which is consistent with TRW theory. The wave energy source at low frequency, such as this 30-day period, is located on the continental slope north of the mooring location. The energy source of TRWs is not clear. Deep-reaching eddies could be one potential source. At similar isobaths at 16°N, moorings from the Meridional Overturning Variability Experiment also showed large current variability (Kanzow et al., 2008), which would propagate southward to the mooring A location.
2.6 Discussion and Conclusions

The North Brazil deep current and its relationship with surface currents were investigated from mooring observation data and satellite altimetry data. The mean current at one mooring showed the southward deep water boundary current (DWBC) exists at the bottom, though the magnitude is smaller than that at other transects (e.g. Kanzow et al., 2008; Meinen et al., 2013). One interesting finding is that two nearby moorings have different mean current and variability. A possible reason is that these two moorings are located at the boundary of the DWBC front, i.e., mooring A is located outside of the DWBC, while mooring B is located inside of the DWBC. For the relationship between the surface current and bottom current, the correlation coefficient between surface eddies and bottom current in the meridional direction is much larger than zonal direction. The correlation coefficient between surface eddies and bottom current is generally positive, indicating eddies have a barotropic vertical structure. The correlation coefficient varies a lot in different periods. One possible reason is that different eddies have different vertical structure, and some eddies could penetrate much deeper than other eddies (Wilson et al., 2001).

Statistical properties, spatial distribution, and vertical structure of NBC rings were studied from satellite altimetry data and Argo float data. NBC rings were identified and tracked using automated eddy identification and tracking procedures. NBC ring vertical structure was constructed from averaged Argo temperature and salinity anomaly data. Mean ring radius was 130 km and mean amplitude was 7 cm. Rings moved northwestward with a mean velocity of 11.7 cm/s. The largest temperature anomaly was 5°C and salinity anomaly was 0.5 psu, located near 200 m depth at the eddy center. The anomaly generally decreased
upward and downward from 200 m depth and away from the eddy center. This study updates the NBC ring surface property analysis conducted by Goni and Johns (2001). Results differed between the lifespan (short-lived and long-lived) of NBC rings being studied. For example, many more rings per year were identified in this study, and the ring propagation speed was smaller than that shown by Goni and Johns (2001). Though most previous studies focused on anticyclonic NBC rings, cyclonic eddies were also a common feature in this region and their properties were similar with anticyclonic eddies.

TRW signal in the western tropical Atlantic Ocean was investigated from mooring current meter observation data. The spectrum at mooring A was dominated by motions at 20~50 days and the peak occurred at around 30 days. The power spectrum at mooring A was consistent with TRW theory. Since current data at other depths were not available, we were not able to check whether the current variability satisfies other aspects of TRW theory. Though the wavelength from these two moorings can be estimated, it would be not statistically meaningful considering the presence of many non-TRW signals. The TRW ray path shows that energy of the 30-day period waves propagated along the continental slope north of the mooring location. The ray path can extend further north with a larger domain. However, the present data are incomplete to get detailed TRW characteristics. It is also hard to distinguish TRW signal from other factors such as deep eddies. A further comprehensive observation system would be needed to provide a better picture of TRWs in this region.

In reality, eddy impact on the deep ocean current is very complex due to eddy-slope interaction. We will examine the idealized eddy modeling in Chapter 3, and eddies in the realistic ocean in Chapter 4. During the NBC Rings experiments, the evolution of NBC rings
was investigated with extensive and integrated observations. However, the mechanism of NBC ring generation, evolution and decay is still not well understood. Deep ocean observation data are not enough to support a detailed process study of the surface eddies’ impact on the deep ocean. High resolution coastal ocean models, combined with extensive observation data, can be a good approach to address these questions in the future.
References


Chapter 3: Impacts of Eddies, Topography, and Topographic Rossby Waves on North Brazil Deep Current Variability: 2. Idealized Numerical Modeling

Abstract

Idealized numerical model experiments are used to understand three-dimensional circulation associated with surface-trapped and deep-reaching eddies over a flat bottom and a slope bottom. It is found that vertical shear of the surface-trapped eddy velocity is much larger than deep-reaching eddy, induced larger vertical viscosity, more potential vorticity loss, and more southward motion. During the eddy horizontal propagation, that negative (positive) correlation exists between surface and bottom north-south velocity component when surface-trapped (deep-reaching) eddy is present. Convergence/divergence and downwelling/upwelling introduced by surface eddies lead to bottom current variability through mass conservation. Compared to the surface-trapped eddy, deep-reaching eddy generates stronger deep current energy variability and larger material transport. Follow-up model analyses adding the continental slope topography reveal that strong vertical mixing is generated at the bottom boundary layer and inside of the eddy. Strong bottom jet and TRWs develop south of the eddy center over the slope, and larger transport is produced than that on the flat bottom. Furthermore, the subsurface and bottom ocean have obvious submesoscale features when eddies interact with the slope and relative vorticity transfers from the large/meso-scale to submesoscale.
3.1 Introduction

Mesoscale eddies are a common feature in the open ocean (e.g., Chelton et al., 2011). They are characterized by temperature and salinity anomalies with associated rotational flow anomalies. They move primarily westward due to the planetary β effect (McWilliams and Flierl, 1979). Large eddies in the Atlantic Ocean include Gulf Stream warm-core rings, Loop Current eddies, and NBC rings. They eventually reach a continental slope or western boundary current. For example, in the tropical Atlantic Ocean, North Brazil Current rings interact with the coast of South America (Richardson et al., 1994; Fratantoni et al., 1995). During the propagation, surface eddies have a potential impact on the deep ocean environment during their propagation. Mooring observations show that surface eddies influence hydrothermal vent efflux transport at the East Pacific Rise (Adams et al., 2011) and deep-sea sediment transport in the northern South China Sea (Zhang et al., 2014).

The dynamics of large isolated eddies has been extensively studied using numerical models. In the open ocean, the propagation and decay of isolated eddies are influenced by eddy size, strength and vertical structure (McWilliams and Flierl, 1979). The anticyclonic (cyclonic) eddies generally move in the southwest (northwest) direction in the northern hemisphere under the combined planetary β and nonlinear effects. The westward propagation of a feature is a consequence of planetary β, while the rate of southward (northward) propagation of an anticyclonic (cyclonic) vortex is determined by the strength of the current (McWilliams and Flierl, 1979; Mied and Lindemann, 1979). Under different dynamic frameworks, eddies show different features during their propagation (Firing and Beardsley, 1976). When eddies are governed by linear dynamics, they disperse and their energy is
radiated away by Rossby waves. Under nonlinear dynamics, barotropic or nearly barotropic cyclonic eddies, which have the same or similar upper layer and lower layer velocities, become elongated on their western side, and an anticyclonic feature forms to the east of original vortex. Eddies evolve into different structures in the upper layer and lower layer (Mied and Lindemann, 1979). Upper ocean eddies are stable, while deep ocean eddies exhibit rapid barotropic-like dispersion, and no vortex is generated in the lower ocean that connects with the upper ocean. Numerical models show that surface eddies could impact bottom currents to some extent. Using a quasi-geostrophic model, Adams and Flierl (2011) showed coherent cyclones with relatively strong current velocities developed in the deep layer associated with simulated surface anticyclones, and coherent cyclones significantly increased the dispersal potential of passive particles. Eddy interactions with ridge topography further enhanced tracer dispersal along the ridge axis through shearing and elongation of the eddy core.

Eddy-slope interaction dynamics has also been extensively studied using numerical models. Smith and O’Brien (1983) and Smith (1986) showed that β-dispersion caused asymmetry in the pressure distribution around an eddy leading to non-linear self-advection. The movement of the eddy then depended on its strength and orientation, planetary β, topographic β, and nonlinear self-adve ctive propagation. Nof (1999) showed eddy migration was governed by three processes in the presence of vertical-walled boundaries: eddies tended to move northward under the image effect, southward due to the β-induced self-advection effect, and northward due to the rocket effect. The image effect was the dominant factor in determining eddies’ final migration along the wall. Eddy-topography interactions often
produce smaller-scale vortices and generate coastally trapped waves, according to numerical modeling studies (e.g., Robinson, 1991; Wei and Wang 2009; Sutyrin and Grimshaw, 2010; Akuetevi and Wirth, 2015). LaCasce (1998) showed that eddies on a slope quickly evolved into upper-layer features and lower-layer features that were obliterated through radiation of topographic Rossby waves (TRWs) in a two-layer model study. Oey and Zhang (2004) described the generation of subsurface cyclones and jets when eddies smashed onto a continental slope and shelf. Wei and Wang (2009) showed that Gulf Stream rings became stalled and bounce on and off the shelf/slope, and small cyclones marked by strong upwelling are generated near the shelfbreak when Gulf Stream rings collided with the slope. Sutyrin and Grimshaw (2010) showed that secondary cyclonic eddies were generated due to off-shelf advection of water with high potential vorticity under the inviscid setting. Vic et al. (2015) revealed the formation of intense frictional boundary layers and generation of submesoscale coherent vortices, leading to Persian Gulf Water shedding in the Gulf of Oman using both idealized and realistic simulations.

Previous modeling studies have focused on isolated eddy dynamics and eddies’ impact on bottom current variability. However, not much attention has been paid to the impact difference under different eddy structure using 3-dimensional numerical model. Previous modeling studies have showed the complex interactions between eddy and slope. However, most simulation studies have focused on upper ocean processes, while little attention has been paid to the slope effect on the eddy-inducing bottom current variability. In this chapter, we will investigate impacts of eddies, topography, and topographic waves on deep current variability utilized idealized modeling approach. The organization of this
Chapter 3 is as follows: Section 3.2 describes the numerical model design, Section 3.3 presents model results on the flat bottom, and Section 3.4 presents model results on the slope bottom. Section 3.5 is the discussion and conclusions.

3.2 Numerical Model Design

In this chapter, a 3-dimensional idealized model is used to study different eddies’ impact on bottom current processes on the flat bottom and slope bottom. The model is based on the Regional Ocean Modeling System (ROMS) (Shchepetkin and McWilliams, 2005), a free-surface, terrain-following, primitive-equations ocean model in widespread use for estuarine, coastal and basin-scale applications.

ROMS solves Navier Stokes equations using the Boussinesq and hydrostatic approximation. The horizontal momentum equations are:

\[
\frac{\partial u}{\partial t} + u \frac{\partial u}{\partial x} + v \frac{\partial u}{\partial y} + w \frac{\partial u}{\partial z} - f v = - \frac{1}{\rho_0} \frac{\partial p}{\partial x} + \frac{\partial}{\partial z} \left( K_m \frac{\partial u}{\partial z} \right) + D_u
\]

\[
\frac{\partial v}{\partial t} + u \frac{\partial v}{\partial x} + v \frac{\partial v}{\partial y} + w \frac{\partial v}{\partial z} + f u = - \frac{1}{\rho_0} \frac{\partial p}{\partial y} + \frac{\partial}{\partial z} \left( K_m \frac{\partial v}{\partial z} \right) + D_v
\]

Surface and bottom boundary conditions are defined as:

at the surface \( z = \zeta(x, y, t) \)

\[
\rho K_m \frac{\partial u}{\partial z} = \tau^x_s(x, y, t) \quad \rho K_m \frac{\partial v}{\partial z} = \tau^y_s(x, y, t)
\]

at the bottom \( z = -h(x, y) \)

\[
\rho K_m \frac{\partial u}{\partial z} = \tau^x_b(x, y, t) \quad \rho K_m \frac{\partial v}{\partial z} = \tau^y_b(x, y, t)
\]
where $D_u$ and $D_v$ are horizontal mixing, $\tau^x_s$ and $\tau^y_s$ are surface wind stress, $\tau^x_b$ and $\tau^y_b$ are bottom stress, $K_m$ is the vertical viscosity coefficient, $h(x, y)$ is bottom depth and $\zeta(x, y, t)$ is the surface elevation.

The model domain in this idealized study was a closed basin, 2000 km wide in the east-west direction and 1200 km wide in the north-south direction. Two kinds of model bathymetry were included in the model: the first one was flat bottom, and the second was the continental slope. The continental slope was modeled as a hyperbolic function defined as:

$$H = -2000 \times [(1 + \tanh(\frac{x-800}{120}))],$$

where $x$ is the distance from the western boundary in km. The continental slope was located at $x=600$ to $x=1000$ km in the model domain. The slope $(\frac{\Delta H}{\Delta x})$ was about 0.01, which was approximately equal to the bottom slope at two mooring locations estimated from ETOPO5 bathymetry data. The model horizontal resolution was 4 km and vertically there were 30 terrain-following levels in the flat bottom experiment and 60 terrain-following levels in the slope bottom experiment. The model used the beta-plane, and the Coriolis parameter varied as $f = 0.7 \times 10^{-5} + 2.25 \times 10^{-11} \times y$, which is located at 4°N - 16°N.

The model initial background temperature data were from the WOA13 profile (Locarnini et al., 2013; Zweng et al., 2013) near Barbados in the western tropical Atlantic Ocean. The model initial background salinity data was set to be constant at 35 psu. The background current and sea surface height were set at zero. To study the impact of surface eddies, different idealized eddies were injected into the background environment. The
idealized eddy was initialized with a Gaussian function temperature anomaly profile defined as \[ \Delta T = T_0 \exp\left(-\frac{r}{r_0}^2\right) \exp\left(-\frac{z-z_s-200}{z_s}\right), \] where \( r \) is the distance from the eddy center, \( z \) is the depth, \( r_0, z_s \) and \( T_0 \) are parameters adjusting the eddy strength and depth. The maximum temperature anomaly was located at 200 m, which was approximately equal to the thermocline depth in this area. The eddy velocity and sea level anomaly were computed using the geostrophic relationship. Two kinds of eddies were defined: the surface-trapped eddy and the deep-reaching eddy. Parameters of these eddies are shown in Table 3.1. Temperature anomaly of the surface-trapped eddy was mainly confined in the upper ocean, and decayed much more quickly than the deep-reaching eddy. We first ran the model using this idealized profile and corresponding geostrophic velocity. After a 10 day adjustment, the model reached an equilibrium state. The final output (Figure 3.1) was used as the following model’s initial conditions. The eddy maximum temperature anomaly was located near the thermocline depth, and the eddy maximum velocity was located at the surface, with a magnitude of 0.92 m/s for the surface-trapped eddy and 0.99 m/s for the deep-reaching eddy. The surface-trapped eddy was mainly concentrated in the upper 400 m, while the deep-reaching eddy had a velocity of 0.3 m/s at 500 m depth and 0.1 m/s at 1500 m depth.

The Mellor-Yamada turbulence closure scheme (Mellor and Yamada, 1982) was used for vertical eddy viscosity and diffusivity. Horizontal viscosity was modeled using constant harmonic diffusion and viscosity coefficients of 20 m$^2$/s. The sponge area was defined to be 20 grids near the boundaries to enhance the viscosity and diffusion by 20 times to reduce wave reflection. The model had no wind stress, heat flux or salt flux input, to isolate other...
factors that have the potential to impact eddy dynamics. The flat bottom experiment ran for 50 days, and the slope bottom experiment ran for 100 days.

3.3 Flat Bottom Results

3.3.1 Eddy Evolution and Trajectories

We ran two experiments on the flat bottom with the bathymetry of 2000 m: a surface-trapped eddy experiment and a deep-reaching eddy experiment. Both experiments ran for 50 days. Figure 3.2 shows model output vorticity and velocity at day 20. Eddies maintained a coherent structure in the upper ocean. At 300 m, the surface-trapped eddy had a much smaller vorticity anomaly than deep-reaching eddy. The bottom ocean was set into motion, with a formation of an even weaker eddy fields for either the surface-trapped eddy or the deep-reaching eddy. In the upper ocean (surface to 300 m), there was a positive vorticity anomaly outside of the eddy boundary. The positive vorticity anomaly was not symmetric, and was stronger in the southern and eastern sides of the eddy. On the bottom, cyclonic eddies developed due to upper layer eddy propagation on the beta plane. Rossby wave feature radiation dominated the bottom variability, as shown from the Rossby wave interference patterns (Early et al., 2011).

The translational motion of an eddy and its related deep current will have a potential impact on the particle motion both in the upper and deep ocean. Figure 3.3 shows experimental float trajectories of a surface-trapped eddy and a deep-reaching eddy over a flat bottom. Five floats were located inside the eddy at the initial time, and then moved passively with the current. For the surface-trapped eddy, floats in the upper 250 m had a nearly closed rotational structure as they moved to the southwest, but at 500 m depth, there was no eddy
structure. Deeper than 750 m, floats remained almost quiescent. The rotational structure of the deep-reaching eddy reached down to 750 m. Flat trajectories at eddy center varied at different depths. Westward displacement over 50 days decreased from 600 km at the surface to 200 km at 750 m depth, while the southward displacement remained almost constant at 200 km. This is because eddy displacement decreases as the rotational velocity decreases. Between 1000 and 1500 m depth, float trajectories showed a transition zone from the surface eddy effect to deep eddy/Rossby wave effect. In the deep ocean, floats mainly moved northward/northwestward about 200 km over 50 days. Therefore, there was a stronger influence of the deep-reaching eddy on the bottom particle displacement.

From the eddy trajectory, we can see that the surface-trapped eddy moved more southward than the deep-reaching eddy. This can be also seen from the planetary vorticity anomaly evolution at the eddy center (Figure 3.4). The planetary vorticity of both the surface-trapped eddy and deep-reaching eddy decreased with time, and the surface-trapped eddy had greater reduction. Potential vorticity (PV), the quantity \((\zeta + f)/h\), should be conserved following the fluid parcel without diffusion, where \(\zeta\) is the relative vorticity, and \(f\) is the planetary vorticity, and \(h\) is the total depth of the fluid. For the anticyclonic eddy, relative vorticity \(\zeta\) was negative and increased during propagation (Figure 3.4). Since the experiment was on a flat bottom and \(h\) was constant, \(f\) must decrease and thereby the eddy moved southward to satisfy PV conservation. The surface-trapped eddy moved more southward than deep-reaching eddy, and PV was not balanced due to diffusion. Therefore, the surface-trapped eddy had more PV loss than deep-reaching eddy.
We compared the vertical viscosity between the surface-trapped eddy and deep-reaching eddy. Though the bottom friction in the deep-reaching eddy experiment was larger than surface-trapped eddy, however, the bottom friction had little impact on the surface eddy evolution. In the deep ocean, the vertical viscosity term \( \frac{\partial}{\partial z} \left( K \frac{\partial v}{\partial z} \right) \) in the momentum equation was smaller for four orders of magnitude than that in the surface ocean (Figure 3.5). At the surface, vertical viscosity term was much larger in the surface-trapped eddy than deep-reaching eddies. This is because the vertical shear of the velocity was larger in the surface-trapped eddy than deep-reaching eddy. Therefore, surface-trapped eddy had larger vertical viscosity, more potential vorticity loss, and moved more southward.

In two-layer experiments, Smith and O’Brien (1983) found that as vortex strength increased, eddy westward propagation speed and the meridional component of its motion were augmented. We had run several more sensitivity experiments in our 3D model, and found that the eddy center displacement increased as the rotational velocity increased, so it was consistent with the 2D model result overall.

3.3.2 Surface Current and Bottom Current Relation

To study the relationship between surface current and bottom current variability, we calculated the correlation coefficient between them for both the surface-trapped eddy and deep-reaching eddy. Figure 3.6 shows the correlation coefficient map of meridional velocity calculated over 50 days. In the surface-trapped eddy experiment, the correlation was mostly negative, due to the cyclonic eddy generation on the bottom. This result was consistent with previous 2D modeling results (Adams and Flierl, 2011). For the deep-reaching eddy, the
correlation was less negative, because the negative correlation was offset by the barotropic mode in the deep-reaching eddy.

We also ran the idealized model on the f-plane, and found that no cyclonic eddy developed on the bottom. Therefore, the generation of bottom cyclonic eddies was caused by upper-layer eddy propagation on the beta-plane. Figure 3.7 shows velocity, convergence and divergence for both the surface-trapped eddy and the deep-reaching eddy on the flat bottom. At the surface, eddies had convergence in their southwestern areas and divergence in their northeastern area. This was because an anticyclonic eddy tends to propagate southwestward. In the bottom, the ocean current mainly had a convergence pattern, especially in the northeastern area below the surface eddy. The convergence and divergence patterns were more persistent for the deep-reaching eddy case.

Consistent with the convergence and divergence, model results showed strong upwelling and downwelling (Figure 3.8) during eddy horizontal propagation. For both the surface-trapped eddy and deep-reaching eddy, their propagation had induced strong downwelling in eddy’s third quadrant and upwelling in eddy’s first quadrant. The upwelling and downwelling could extend to the ocean bottom, resulting in bottom current motion due to mass conservation. The vertically velocity pattern was not symmetric around the eddy center and tilted eastward, especially for the deep-reaching eddy. Therefore, this downwelling and upwelling was responsible for the bottom current variability.

Surface-trapped eddy and deep-reaching eddy had different energy evolution and different impact on the deep current energy variability. The total kinetic energy was calculated over the different water column depth (Figure 3.9). In the whole water column,
both the surface-trapped eddy and deep-reaching eddy had 20% kinetic energy loss over the 50 days. The evolution of the kinetic energy at different depths was different. At the upper 200 m, kinetic energy in the surface-trapped eddy experiment decreased greatly over time, while energy in the deep-reaching experiment showed much less change. Between 200 m to 1000 m, the energy was stable for surface-trapped eddy, while decreased greatly in the deep-reaching eddy experiment. At the bottom 1000 m, the energy in both experimental eddies rapidly increased. It is clear that energy had transferred from the upper ocean to the bottom ocean during eddy propagation, and the deep-reaching eddy induced larger bottom current energy variability.

3.4 Slope Bottom Results

3.4.1 Eddy Evolution and Trajectory

We ran two experiments over the slope bottom for both the surface-trapped eddy and the deep-reaching eddy. We also ran another two experiments as the control experiments on the flat bottom with the bathymetry of 4000 m. Each experiment ran for 100 days. Both the surface-trapped eddy and the deep-reaching eddy were initially located in the open ocean, and moved southwestward to approach the continental slope. To compare surface-trapped eddy and deep-reaching eddy over a slope, Figure 3.10 shows the model output vorticity anomaly and velocity results of these two different eddies at day 40. At the surface, there was no obvious interaction between surface eddy and slope. Similar to that over a flat bottom, the center of the deep-reaching eddy moved less southward and more westward than the surface-trapped eddy. At 300 m, the deep-reaching eddy had a much larger vorticity anomaly than surface-trapped eddy in the open ocean. When impacting the continental slope, the vorticity
anomaly showed propagation along the slope. South of the eddy center, the eddy had a positive vorticity anomaly, which stretched along the slope. In the deep ocean, the deep eddy and Rossby wave were generated. Over a slope, the deep eddy and Rossby wave were impacted by the slope, especially for the deep-reaching eddy, since there were strong positive vorticity and strong velocity anomaly at the south along the slope.

When eddy eventually reached a continental slope, their trajectories would be altered by the slope. Figure 3.11 shows the eddy trajectories for both surface-trapped eddy and deep-reaching eddy. We can see that both the surface-trapped eddy and deep-reaching eddy slowed down when reaching the slope, and later had a southward net movement. The deep-reaching eddy had a larger net movement. This result was different with previous results. For example, Shi and Nof (1994) showed that an eddy migrated alongshore under the influence of the beta force, image effect and ‘rocket’ effect when collided with a wall, and found that the image effect usually dominated and eddy moved northward. Wei and Wang (2011) found the ring became stalled and bounced on and off the shelf/slope with little net movement when colliding with a northeast-southwest-oriented shelf/slope. Since our model had different eddy structure, located at different latitude, had different slope orientation et al., so these factors probably accounted for the trajectory difference compared with previous model results.

3.4.2 Slope Jet and TRWs

To get a clearer view of the continental slope effect on eddies and deep currents, we compared model results of a deep-reaching eddy on a flat bottom and on a slope (Figure 3.12). In the upper ocean, the velocity anomalies were similar, with several northward and southward velocity cores. In the deep ocean, the current was very weak over the flat bottom;
however, with the impact of the slope, the eddy had induced strong velocity anomalies reaching the ocean bottom. Corresponding to the bottom-reaching velocity, there was a large temperature anomaly along the slope. Between X=1400 km and X=1600 km, the current had deep-reaching feature over both the flat bottom and slope. Though the nonlinear eddy evolution was dominated by the coherent westward-propagating sea surface anomaly, it also showed Rossby wave interference patterns (Early et al., 2011). Therefore, this observed deep-reaching feature can be explained as the barotropic Rossby wave induced behind of the eddy. Slope topography generated strong mixing over the slope (Figure 3.13). We used the vertical temperature diffusion term as the indicator of the vertical mixing magnitude. Compared with the flat bottom, there were two areas with enhanced mixing, one was located at the bottom boundary layer, and the other was inside the eddy at the upper 100 m closed to the slope.

TRWs were expected to be excited and propagate along isobaths if there were vorticity anomalies over the continental slope. Figure 3.14 shows the deep current map for the flat bottom and over a slope. The continental slope had the effect of sustaining the anomaly propagating along the slope. On the flat bottom, the bottom current was relatively symmetric and had the Rossby wave feature. With the slope, not only did the current near the bottom become much stronger than over the flat bottom, but also the current anomaly extended southward along the slope. This was expected as the topographic waves propagated southward. Since TRWs propagated southward, the anomaly was very weak in the area north of the eddy center. TRW propagation resulted in a relatively northward jet, i.e., the strong
current along the slope, as large as 3~5 cm/s. This could have a strong impact on material property transport at the ocean bottom.

3.4.3 Submesoscale Features

During interaction with the slope, eddies were deformed and submesoscale features were generated. Figure 3.15 shows the relative vorticity of a deep-reaching eddy on a flat bottom and on the slope. On the flat bottom, the eddy had a compact structure at 300 m depth with the radius around 100 km. A strong positive vorticity anomaly existed south of the eddy and a Rossby wave pattern existed behind the eddy. In the deep ocean, the main structure was the Rossby wave. With the slope, several submesoscale vortices appeared north of the eddy center. The length of the submesoscale vortex was around 30~50 km. The submesoscale vorticity magnitude was of the same order as the original mesoscale eddy. Both positive vorticity and negative vorticity in the submesoscale vortexes existed. At 2000 m depth, although there was no rotational vortex structure, the slope jet had relatively strong velocity shear and vorticity anomaly. This vorticity anomaly had a width of around 30~50 km, which was also in the range of the submesoscale processes.

Not only relative vorticity but also divergence, shear and strain would be impacted by topography. Figure 3.16 shows these eddy properties between over a flat bottom and over a slope. In both experiments, the relative vorticity, shear and strain were of the same order, while the divergence was smaller over the flat bottom. Therefore, eddy-slope interaction had enhanced the upwelling and downwelling process. On the flat bottom, the shear and strain were symmetric in the eddy center. However, on the slope, noisy shear and strain features in
the sub-mesoscale range appeared during the interaction. Thus both shear and strain contributed to the existence and evolution of the submesoscale vortex.

We applied 2D spatial Butterworth filter to the vorticity field, and distinguished the large/meso-scale feature from sub-mesoscale feature clearly (Figure 3.17) using 50 km spatial filter cutoff value. The large/meso-scale mainly had rotational vortex structure, and the submesoscale had the filament structure. We used the vorticity root mean square (rms) over the whole domain as the indicator to study the temporal evolution of the large/meso-scale and submesoscale features. The overall surface vorticity magnitude decreased by 30% over the 100 days. The large/meso-scale decayed by 50%, which was much more than overall decay magnitude. The submesoscale vorticity rms initially was small, and increased during the 50 days, and then decayed over the next 50 days. At the day 50, the submesoscale feature was as large as the large/meso-scale feature. Therefore, there is strong vorticity transfer from the large/meso-scale to submesoscale. Since our model resolution was 4 km, we cannot use the spatial filter to study the small scale variability directly.

3.4.4 Bottom Float Trajectories

To study eddy effect on bottom particle displacement, we put fourteen floats on the bottom along the slope. Figure 3.18 shows float trajectories over both the flat bottom and slope. The slope had an impact on the motion of bottom particles. Over a flat bottom, there was little displacement between Y=300 km and Y=600 km in the model domain, while floats moved offshore to the north between Y=600 km and Y=800 km. Over a slope, there was strong northward particle motion along the slope. Floats moved 50 km in 10 days, or 5 cm/s, which was consistent with bottom current magnitude.
Particle trajectories can be used to estimate eddy diffusivity. Taylor (1921) showed that in steady homogeneous turbulence, the particle dispersion function would be a straight line proportional to time for a short period. The relation can be expressed as $<d(t)^2> = 4Dt$, where $<d(t)^2>$ is the mean square distance between particles and their center of mass, $t$ is the release time, and $D$ is eddy diffusivity. It reveals how a cluster of tracers spread about its center of mass in a random walk process (LaCasce, 2008). To quantify this diffusion, we computed the absolute dispersion of particles on the slope around their mean position as a function of the release absolute time (Figure 3.19). These particles were deployed at Y=400 km and X=800 km location with a time interval of 10 days. We can see that from the release time to about 10 days, $<d(t)^2>$ could be approximated by a linear function of time $t$. Assuming this model was valid and using a linear regression, we estimated the eddy diffusivity at the bottom of the slope as 1.8 m$^2$ s$^{-1}$. Our horizontal eddy diffusivity agreed well with reported observed horizontal diffusivity from Ledwell et al. (1991), who found open-ocean horizontal eddy diffusivity is 1~3 m$^2$ s$^{-1}$.

3.5 Discussion and conclusions

In summary, idealized ocean models were used to investigate surface-trapped and deep-reaching anticyclonic eddies’ impact on deep current variability. Model results showed that surface eddies only kept coherent structure at the surface, while their bottom currents were much weaker on a flat bottom. However, cyclonic eddies developed due to propagation in the upper layer eddy on the beta plane, and then Rossby wave radiation dominated bottom variability. A previous two-layer numerical modeling study (Adams and Flierl, 2010)
showed the presence of a cyclone at a lower layer associated with the surface anticyclone, significantly increasing the dispersal potential of passive particles. Our 3D idealized model showed some similar features, such as eddies’ impact on the displacement of bottom particles. More complex features also appeared in this 3D model experiment, such as the differing impact between surface-trapped eddies and deep-reaching eddies. The surface-trapped eddy had much smaller impact on the deep current variability, while the deep-reaching eddy could disperse bottom particles about 200 km over 50 days.

The idealized 3D primitive equation model was further used to expand the simulation study over a flat bottom and investigate topographic effect on eddies and bottom current processes. The continental slope was introduced with a smooth hyperbolic function. Several features appeared with the inclusion of bottom slope. First, the current was more likely to have deep-reaching features. There was the strong mixing generated over the slope near the bottom boundary layer and inside of the eddy. Second, bottom cyclonic eddies and Rossby waves were impacted by the slope, and disturbances stretched along the slope, resulting in stronger northward bottom currents. This bottom current anomaly had a strong impact on bottom particle motion and could move particles as fast as 50 km in 10 days. These results were also consistent with an early linear model study by Chapman and Brink (1987). They found that the shoreward flow in the eddy could not move onto the shelf and instead formed an alongshore jet near the shelf break. The jet extended away from the eddy in the direction toward which free coastally trapped waves propagated. Third, the subsurface and bottom ocean had obvious submesoscale features when eddies interacted with the slope.
The model result showed the generation of jets during eddy-bottom interaction. In previous studies, the jet was mainly located in the upper ocean (e.g., Oey and Zhang, 2004). In our study, the jet could be as deep as 2000 m. Since the bottom current was dominated by the Rossby wave pattern on the flat bottom, the jet in the deep ocean was caused by the interaction of the Rossby wave and the slope, not directly by the eddy and slope interaction. The jet could be also viewed as propagation of TRWs, which propagated in a direction with the shallower water on the left. Therefore, the eddy had a more complex impact the bottom current with the presence of the slope, and the slope sustained the propagation of the TRWs.

In this chapter, we have investigated the eddy impact on the bottom current variability on the flat bottom and slope using the idealized model. Further study will be needed using the realistic ocean setting to reveal the complex interaction process between the eddy, slope, TRWs and background current, and we will explore this in Chapter 4.
References


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Oey, L.-Y. and H.C. Zhang, 2004: The generation of subsurface cyclones and jets through eddy-slope interaction. Continental Shelf Research, 24, 2109-2131


Chapter 4: Impacts of Eddies, Topography, and Topographic Rossby Waves on North Brazil Deep Current Variability: 3. Realistic Numerical Modeling

Abstract:

A realistic, 3D, primitive-equation model was used to hindcast NBC rings and deep current variability in the western tropical Atlantic Ocean. The model was able to reproduce many observed features, such as the westward propagation of three large surface rings during May to December 2011, and southward propagation of TRWs in the deep ocean. Our model results showed that some NBC rings had deep-reaching features and others had surface-trapped features. Significant TRWs signals propagated from north to south with a distance of 200 km over 2 months. Bottom current variability was impacted by surface eddies locally and TRW propagation remotely. Particle motion on the continental slope depended on whether it was located inside or outside of the Deep Western Boundary Current (DWBC).
4.1 Introduction

The circulation in the western tropical Atlantic Ocean is very complex. In the upper ocean, the North Brazil Current (NBC) flows northward along the northeastern coast of South America. When reaching French Guiana, part of the NBC separates from the coast and retroflects to join the North Equatorial Counter Current (NECC), while the rest of the NBC continues flowing northwestward to form the Guiana Current (GC, Condie, 1991). During the retrofection into the NECC, large anticyclonic NBC rings are shed and move northwestward toward the Caribbean Sea on a course parallel to the South American coastline (Goni and Johns, 2001). After translating northwestward for several months, NBC rings decompose in the vicinity of the Lesser Antilles (Fratantoni et al, 2006). Near the bottom, circulation features the Deep Western Boundary Current (DWBC), transporting North Atlantic Deep Water from the overflow region between Greenland and Scotland and from the Labrador Sea into the South Atlantic and the Antarctic Circumpolar Current (Kanzow et al., 2008; Peña-Molino et al., 2012; Meinen et al., 2013).

With the development of numerical modeling in recent decades, several realistic models have been applied to study tropical Atlantic circulation. For example, Garraffo et al. (2003) analyzed ring shedding, classification of ring types, and NBC rings’ volume transport using the Miami Isopycnic Coordinate Ocean Model (MICOM). Their modeled NBC generated a variety of rings, classified as ‘shallow’, ‘intermediate’, ‘deep’, and ‘subsurface’. NBC rings contributed 40% of the total meridional transport from the surface to the intermediate water layers. Jochumsen et al. (2010) investigated ring shedding, propagation and the complex interaction with the Lesser Antilles in the 1/12° Family of Linked Atlantic
Model Experiments (FLAME) model. Their modeling results showed that the deep and surface-intensified rings dominated in winter and spring and the subsurface rings were predominantly formed during fall. The interaction of NBC rings with the topography of the Lesser Antilles was also investigated, showing that rings approached the Lesser Antilles from the southeast, moved northwestward around Tobago, were trapped in the topographic triangle of Tobago, St. Lucia and Barbados until their diameter was reduced to approximately 180 km, and then passed through the St. Lucia-Barbados gap.

Topographic Rossby Waves (TRWs) are transverse, quasi-geostrophic waves found in regions of sloping topography (Rhines, 1970). Some key features of TRWs are that the wave motions are bottom-intensified, the maximum frequency for TRW is $N\alpha$, where $\alpha$ is the bottom slope and $N$ is the Brunt-Vaisala frequency, and the along-isobath component of the wave vector is always negative. TRWs are important processes that dominate the deep current field on the continental slope. Numerous moored arrays along the continental slope of the Mid-Atlantic Bight (MAB) (e.g., Hogg, 1981, Hamilton, 1984) and Gulf of Mexico (e.g., Hamilton, 1990, Hamilton, 2009) revealed that the current variability below 1000 m was dominated by TRWs, since current variability conformed to the TRW dispersion relation.

Previous hindcast models mainly studied upper ocean dynamics in the western tropical Atlantic Ocean, while little attention has been paid to bottom current variability. In the real ocean, the deep ocean environment is very complex, and potentially impacted by surface eddies and TRWs. To study impacts of eddies and TRWs on North Brazil deep current variability in the realistic ocean, a hindcast experiment was implemented in the
western tropical Atlantic Ocean using the Regional Ocean Modeling System (Shchepetkin and McWilliams, 2005). The organization of this Chapter is follows: Section 4.2 describes the model configuration, Section 4.3 presents the model result analysis, and Section 4.4 is the discussion and conclusions.

4.2 Model Configuration

The model domain covered the area between 65°W and 52°W, 2°N and 20°N (Figure 4.1). The horizontal resolution of this model was about 4 km, much higher than previous 1/12° FLAME model or HYCOM model. Model bathymetry was interpolated from National Geophysical Data Center (NGDC) 2-Minute Gridded Global Relief Data, with the minimum water depth of 10 m and maximum water depth of 5500 m in the model domain. Vertically, it had 36 terrain-following layers spaced to better resolve surface and bottom boundary layers. The vertical coordinate transformation was:

\[
z(x, y, \sigma, t) = \zeta(x, y, t) + \left[ \zeta(x, y, t) + h(x, y) \right] \frac{h \sigma + h(x, y) C(\sigma)}{h_c + h(x, y)}
\]

\[
C(\sigma) = (1 - \theta_s) \frac{\sinh(\theta_s \sigma)}{\sinh \theta_s} + \theta_b \left[ \frac{\tanh(\theta_b (\sigma + \frac{1}{2}))}{2 \tanh(\frac{1}{2} \theta_b)} - \frac{1}{2} \right]
\]

where \( \sigma \) is a fractional vertical stretching coordinate ranging from -1 to 0, \( \theta_s \) and \( \theta_b \) are surface and bottom control parameters, and \( h_c \) is the positive thickness controlling the stretching. In this model, vertical s-coordinate model parameters are \( \theta_s = 5 \), \( \theta_b = 0.4 \) and \( h_c = 10 \). Momentum advections were solved using a 3rd-order upstream bias scheme for 3D velocity, and a 4th-order centered scheme for 2D transport. Tracer (temperature and salinity)
advections were solved with a 3rd-order upstream scheme in the horizontal direction, and a 4th-order centered scheme in vertical direction. The horizontal mixing for both the momentum and tracers utilized the harmonic formulation with momentum and tracer mixing coefficients of 50 m²/s.

The hindcast experiment ran from May 1 to December 31, 2011. The initial and boundary conditions were from global data assimilative HYCOM/NCODA output (http://www.hycom.org/). Within 10 grid points of near the eastern and western boundaries and 50 grid points near the northern boundary, 3-dimesional tracer and momentum fields were nudged to corresponding HYCOM model fields. Surface forcing data were from NCEP data, with the 3-hour temporal interval and the 2.5° horizontal resolution. There were no rivers or tidal forcing in this realistic model, since we did not focus on coastal or shelf processes. To be consistent with the mooring observation data presented in Chapter 2, we only analyzed the hindcast results from May 20 to December 20, 2011.

4.3 Model Results Analysis

The model was able to reproduce both surface circulation and bottom circulation features in the western tropical Atlantic Ocean. Figure 4.2 shows vorticity and velocity at the surface, 300 m and 2000 m on August 13, 2011 from model output. The current was the typical western tropical Atlantic Ocean environment. At the surface, the current was northwestward and there was a large NBC ring located at 10°N, 58°W. The maximum meridional velocity was 1.4 m/s, and zonal velocity was 1.0 m/s. The eddy amplitude was of the same order as previous idealized model results in Chapters 3. There was strong negative relative vorticity inside of the eddy, and strong positive relative vorticity near its boundary.
The eddy moved northwestward, and the direction was consistent with the census analysis in Chapter 2. The eddy moved with a speed of around 11 km per day in the zonal direction and 4 km per day in the meridional direction, with the total velocity around 12 km/day. The result was a bit larger than the result estimated from the altimetry census data of 10 km/day, also a bit larger than the idealized model data of 10 km/day, while a little smaller than previous remote sensing results of 14 km/day (Goni and Johns, 2001). Since the ring translational speed depended on the ring magnitude and radius, and this ring was larger and stronger than census and idealized modeled rings, the ring translational speed was also larger than previous observational and idealized modeling results. Compared with idealized model results, some additional features appeared, such as the largest rotational velocity being located in the eddy’s third quadrant. This was probably due to the northwestward flow of the Guiana current and eddy-slope interaction. The NBC rings’ diameter was around 300 km, and it had subsurface features at 300 m depth. At 2000 m, there was a strong southward current flowing along the continental slope, which represented the DWBC. Close to the DWBC, there existed several sub-mesoscale eddies; this was also shown in the idealized model in Chapter 3. In the realistic ocean, the sub-mesoscale eddies would interact with the background current and evolved more complex structures than that without background currents.

To compare model results with observations and investigate the surface eddy impact on bottom current variability, we extracted model surface and bottom currents at the mooring locations (Figure 4.3). The A and B points were the mooring A and mooring B location, and X was a point located further shoreward of mooring A. Because of the mismatch between topography used in the model simulation and that in reality, points in the model had location
bias compared with mooring locations due to smoother model topography. We therefore compared the actual mooring A with model point X, and actual mooring B with model point A. Temporal variation in surface velocity shows that three rings passed the moorings during the simulation period. The first was from mid-June to mid-July, the second was during August, and the third was during October. Modeled surface velocity of the rings was as large as 50 cm/s, the same as that estimated from altimetry data. Modeled bottom current variability was similar to mooring current meter observations in that the seaward point showed a persistent southward current, while the shoreward point showed variation with mean velocity of zero. The variation amplitude is much smaller than the observed current. The range of modeled meridional velocities was less than 10 cm/s, while the range of observed meridional velocities was over 20 cm/s. This is also seen from the energy-preserving power spectra comparison between simulated velocity and observed velocity (Figure 4.4). The spectra magnitude of modeled meridional velocity was smaller for one order of magnitude than that of the observed meridional velocity. Some possible reasons could be the bathymetry mismatch between observation and model, the model resolution and the boundary forcing quality from HYCOM data. The pattern of the energy spectra of meridional velocities was similar between model results and mooring observation results. Both observation and model spectra showed that at the onshore location, the energy spectra were dominant in the low-frequency band with the period of 20-100 day, while at the offshore location, the energy spectra had large amplitude at high-frequency band with the period of 5-10 day.
To show the structure and propagation of the NBC rings, the meridional velocity anomaly at the 10.3°N transect was examined. This transect was the location of the two Barbados moorings. Two NBC rings passed this transect in July and August, as shown in Figure 4.5. The eddy structure varied between the dates, and was different from the idealized eddy. The first eddy reached deeper, with a magnitude of over 5 cm/s at 1000 m, while the second eddy was mainly confined to the upper 200 m. Near the continental slope, the eddy was impacted by the slope, causing a large velocity anomaly to stretch downward and reach deeper than in the open ocean. This was consistent with the previous idealized modeling study, which showed the deep-reaching feature and slope jet which appeared during eddy-slope interaction.

To investigate bottom particle behavior, five floats were modeled on the slope close to the moorings, then allowed to flow passively with the current. Figure 4.6 shows the float trajectories on the bottom of the continental slope. The result was consistent with the mooring velocity observations. In the deep ocean, two shoreward floats moved back and forth around their initial locations. Seaward floats were impacted by the DWBC and flowed southeastward along the slope for 80 km in 30 days. Therefore, the DWBC had a direct impact on bottom particle transport and seep ecology. This result was more complex than that from the idealized model, which showed bottom floats mainly flowing along the slope. In this realistic model, some floats had cross-isobath motion and moved seaward at a later time. This was probably caused by sub-mesoscale vortex features in the deep ocean as shown in Figure 4.2. At other months (not shown), float trajectories showed similar patterns, though
the number of southward-moving floats was different. This demonstrated the onshore and offshore variation of the DWBC boundary.

Since TRWs are important features that impact the deep current field on the continental slope, we extracted model hindcast velocity data in 2011 along the ray path calculated in Chapter 2 to examine TRW propagation in this area. Figure 4.7 shows the velocity anomaly along the TRWs’ ray path at 2000 m. Significant TRWs signals propagated from north to south, as shown from the line with arrow in the figure, with a distance of 200 km over 2 months. The estimated propagation speed was around 4 cm/s. TRW signals also decayed quickly, such as the strong anomaly at 15°N in June became much weaker when the anomaly propagated at 14°N in July. Other anomaly signals were trapped locally and did not propagate southward, such as the strong velocity anomaly at 15°N in December. The southward propagating TRWs did not last for longer than 300 km from north to south. Other local factors, such as baroclinic instability, had a potential impact on the propagation and caused the decay of TRWs signals. The energy source of TRWs is not clear, but surface eddies could be one potential source.

4.4 Discussion and Conclusions

In summary, a realistic, 3D, primitive-equation model was used to hindcast NBC rings and deep current variability in the western tropical Atlantic Ocean. One NBC ring we examined showed deep-reaching features and another showed surface-trapped features. Eddies were impacted by seafloor slope during propagation. Since current variability showed a TRW signal and some eddies reached deep, bottom current variability was impacted by surface eddies locally and TRW propagation remotely.
The model was able to reproduce many observed features, such as the westward propagation of three large surface rings during May to December 2011, and southward propagation of TRWs in the deep ocean. The pattern of the energy spectra of meridional velocities was similar between model results and mooring observation results. Particle motion on the continental slope depended on whether it was located inside or outside of the DWBC. Furthermore, eddy properties such as translational speed and radius were similar to the census observation results.

Compared with the idealized model results, this realistic model revealed much more complex features. For example, in the upper ocean, eddy structure was not compact and symmetric, and varied greatly both horizontally and vertically. Rings in this realistic model moved northwestward, while in the idealized model they moved southwestward. Near the bottom, sub-mesoscale eddies existed near the DWBC, and bottom floats had cross-isobath motion and moved seaward. These features are caused by eddy interaction with surface forcing, bottom currents, realistic topography and the complex initial and boundary conditions in the realistic model.

Either surface eddy impact or TRWs could explain bottom current variability from our model results. Bottom current variability could be impacted by the combined effects of surface eddies, TRWs propagation, topography, and other factor such as deep ocean eddies. This was similar to the subinertial velocity variability at East Pacific Rise (Liang and Thurnherr, 2011), which was concluded to be a superposition of velocities associated with eddies propagating westward across the ridge and topographic flows. Johns et al. (1993) also pointed out that this process was difficult to document because deep ocean current motion
was in general a random superposition of responses to forcing at different times and places over a broad region.

Though the model can reproduce many observational features, there is still some inconsistency between observations and modeling results. Due to the scarcity of observations, it is difficult to evaluate the model hindcast skill comprehensively. Modeling the bottom current variability is challenging; other approaches, such as high resolution data assimilative ocean modeling (e.g., Chen et al., 2014; Li et al., 2015), combined with extensive observational data, will provide more power to address deep ocean circulation dynamics in the future.
References


Chapter 5: Summary

In this dissertation, impacts of eddies, topography, and topographic Rossby waves on North Brazil deep current variability are investigated utilizing an observation and modeling synthesis study. The observation study mainly includes satellite altimetry data, Argo float data, and mooring current meter data analysis. Numerical modeling includes both the idealized model and realistic model approaches. Some major findings of this dissertation study are as follows:

In Chapter 1, we show that two in situ observations of north Brazil deep current were made in May-December 2011, and May 2011- May 2013, revealing drastic difference of bottom circulation. It is found that current measured by the offshore buoy (Orenoque B) is a part of the deep western boundary current (DWBC), showing persistent southwestward motion. In contrast, current measured 15 miles apart (Orenoque A) displays much larger variability.

In Chapter 2, we show that during the two-year buoy observation time at the mooring A location, twelve eddy periods are identified. In four (two) periods, the correlation coefficient between surface and bottom meridional current is significant and larger (smaller) than 0.6 (-0.6). Therefore, surface eddies potentially have a direct impact on the deep ocean circulation. Based on 14-year satellite altimeter data, an eddy census analysis shows that most (>50%) eddies in the region have around 130 km radius, 4-8 week life spans, and >7 cm sea level anomaly. In addition to impact from local eddy and topography, power spectra analyses on slope mooring (Orenoque A) current data and ray tracing calculation show the
bottom current variability can be introduced from a remote site north of mooring location via the TRW propagating along the slope.

In Chapter 3, idealized numerical model experiments were used to understand three-dimensional circulation associated with surface-trapped and deep-reaching eddies over a flat bottom and slope bottom. It is found that negative (positive) correlation exists between surface and bottom north-south velocity component when surface-trapped (deep-reaching) eddy is present. Convergence/divergence introduced by surface eddies lead to bottom current variability through mass conservation. Compared to the surface-trapped eddy, deep-reaching eddy generates stronger deep current and larger material transport. Model analyses adding the continental slope topography reveal that strong vertical mixing is generated over the slope inside of the eddy. Strong bottom jet and TRWs develop at the south of the eddy center over the slope, and larger transport is produced than that on the flat bottom. Furthermore, the subsurface and bottom ocean have obvious submesoscale features when eddies interact with the slope.

In Chapter 4, a realistic modeling taking eddy, topography, boundary current and TRWs into consideration is used to hindcast circulation during the study period. Model-data comparisons show the model is able to resolve most of observed variability. Model results showed that bottom current variability was impacted by surface eddies locally and TRW propagation remotely. Particle motion on the continental slope depended on whether particles were located inside or outside of the DWBC. Deterministic prediction of bottom current variability clearly needs further model refinements, better observations, together with using advanced data assimilation schemes.
This dissertation study has several significant research facets. First, eddy-topography interaction is a fundamental research topic in physical oceanography. Second, studying eddy/ring properties in the north Brazil region will deepen the understanding of circulation dynamics in the tropical western Atlantic Ocean. Third, understanding deep current variability can provide crucial insight about deep sea environment and marine ecology (such as cold seeps ecosystem and sediment transport). Through this data analysis and numerical modeling study, an improved understanding of impacts of eddies, topography, and topographic Rossby waves on North Brazil deep current variability has been achieved. This study has deepened the understanding of circulation dynamics in the tropical western Atlantic Ocean and deep ocean environmental processes. The research provides a valuable information and toolsets for the future observational data analysis and modeling study related to eddy-TRWs-DWBC interaction. Future efforts utilizing high resolution data assimilative ocean model combined with more extensive observation data hold promise to further elucidate research questions addressed herein.
Figure 1.1: Schematic representation of mean currents in the western tropical Atlantic. The North Brazil Current (NBC), the North Equatorial Current (NEC), the North Equatorial Counter Current (NECC), Guiana Current (GC), and the Caribbean Current (CC) are labeled. The bottom current indicated is the Deep Western Boundary Current (DWBC). A and B represent Orenoque A mooring and Orenoque B mooring location, respectively.
Figure 1.2: Vertical profile of temperature, salinity and squared Brunt-Vaisala frequency at the mooring location from WOA13 data.
Figure 1.3: Velocity structure of North Brazil Current rings (Wilson et al., 2002). Top panels show maps of the shipboard ADCP current vectors for each of the four rings that were surveyed; bottom panels show the velocity sections across each ring derived from the lowered ADCP data. Main isotherms from the CTD profiles concurrent with the lowered ADCP's are superimposed on the velocity maps. Velocity contour level is 20 cm/s, Temperature contour level is 5°C.
Figure 1.4: Left panel: regional map showing the bathymetry of the East Pacific. The star marks the location of the study site at the 9°N vent field. Right panel (A): Sea Level Anomalies and inferred geostrophic currents show the passage of two anticyclonic eddies over the 9–10°N vent field from May through June 2007. Right panel (B): Near-bottom current anomalies on the East Pacific Rise at 2430 m depth. (Adams et al., 2011)
Figure 1.5: Velocity vectors (blue is cyclonic and red is anti-cyclonic) superimposed on a color map of relative vorticity non-dimensionalized by the Coriolis parameter $\zeta/f$ at the indicated depth and time. Strong forcing is visible in the southeast and disturbances to the north and west. (Oey et al., 2009)
Figure 1.6: Map of cold seeps in the Atlantic Equatorial Belt. Cold seeps are found at the Barbados Trench (BT), Blake Ridge diaper (BR), Orenoque sectors (OR), El Pilar sector (EP), Nigerian slope (NIG), Guiness area (GUI), Regab pockmark (REG), and several other locations in the Gulf of Mexico. Circles show our research vessel investigation locations, labeled with the year of the investigation.
Figure 1.7: Bottom current variability from May 20, 2011 to December 20, 2011 with 72 hr low-pass filter at Orenoque A mooring (upper left panel) and its inferred particle motion (upper right panel), and current variability at Orenoque B mooring (lower left panel) and its inferred particle motion (lower right panel).
Figure 2.1: Map of SLA (unit: m), with detected anticyclonic eddies (red dots) and cyclonic eddies (blue dots) and weekly eddy tracking locations (purple dots and line) during the first six months when eddies were located near the moorings (black dots).
**Figure 2.2:** Time series of sea level anomaly (upper panel), surface-inferred geostrophic currents (middle panel), and variability of bottom current anomaly (lower panel) at the mooring A location. Twelve periods are divided by vertical lines.
Table 2.1: Correlation coefficients between surface and bottom currents at mooring A at different period. The U and V mean the zonal and meridional current, respectively.

<table>
<thead>
<tr>
<th>Period</th>
<th>Start Date End Date</th>
<th>Degree of Freedom</th>
<th>Correlation Coefficient with 95% Confidence Interval</th>
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<td></td>
<td></td>
<td></td>
<td><strong>U</strong></td>
</tr>
<tr>
<td>Period 1</td>
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<td>40</td>
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<tr>
<td></td>
<td></td>
<td></td>
<td>-0.78</td>
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<td></td>
<td>0.66</td>
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<td>0.64</td>
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<td></td>
<td></td>
<td>0.50</td>
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Figure 2.3: Eddy center locations and trajectories of eddies with lifespans of no less than 4 weeks (left panels). The red dots denote eddy birth locations. The spatial distribution of the number of eddy over a 1°×1° pixel grid (right panels). The gray line shows the 2000 m isobath.
Figure 2.4: Probability distribution of anticyclonic eddy (upper panels) and cyclonic eddy (lower panels) properties. The left panels are the eddy radius distribution, the middle panels are the eddy lifespan distribution, and the right panels are the eddy amplitude distribution.
Figure 2.5: Eddy zonal propagation speed (m/s) (left panels) and meridional propagation speed (m/s) (right panels) of anticyclonic eddy (upper panels) and cyclonic eddy (lower panels).
Figure 2.6: Ensemble temperature anomaly (upper panels) and salinity anomaly (lower panels) of anticyclonic eddy (left panels) and cyclonic eddy (right panels).
Figure 2.7: Energy-preserving power spectra of bottom u and v velocities at mooring A (left panel). Smoothed bathymetry (unit: m) and the ray path of 30-day period TRWs (right panel).
Table 3.1: Parameters of the surface-trapped eddy and deep-reaching eddy.

<table>
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<th>$r_0$</th>
<th>$z_s$</th>
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<th>$z&gt;=200$</th>
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<td>Deep-reaching</td>
<td>1.3</td>
<td>100</td>
<td>100</td>
<td>750</td>
<td></td>
</tr>
</tbody>
</table>
Figure 3.1: Ocean temperature, temperature anomaly and swirling velocity of the surface-trapped eddy (upper panel) and the deep-reaching eddy (lower panel).
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