Dynamics of Wind-driven Atlantic Meridional Overturning Circulation Variability

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DYNAMICS OF WIND-DRIVEN ATLANTIC MERIDIONAL OVERTURNING CIRCULATION VARIABILITY

By

Jian Zhao

A DISSERTATION

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DYNAMICS OF WIND-DRIVEN ATLANTIC MERIDIONAL OVERTURNING
CIRCULATION VARIABILITY

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The dynamical processes governing the wind-driven Atlantic Meridional Overtuning (AMOC) variability are studied using observations and a variety of models, ranging from a simple forced Rossby wave model to an eddy-resolving Ocean General Circulation Model (OGCM). To better identify the mechanisms for the AMOC variability at different time scales, the AMOC is decomposed into Ekman and geostrophic transport components. On seasonal time scale, the AMOC variability is determined by both components in the extratropics but dominated by the Ekman transport in the tropics. While the Ekman transport is directly related to zonal wind stress seasonality, the comparison between different numerical models shows that the geostrophic transport involves a complex oceanic adjustment to the wind forcing. The oceanic adjustment is further evaluated by separating the zonally integrated geostrophic transport into eastern and western boundary currents and interior flows. Our results indicate that the seasonal AMOC cycle in the extratropics is controlled mainly by local boundary effects, where either the western or eastern boundary can be dominant at different latitudes, while in the northern tropics it is the interior flow and its lagged compensation by the western boundary current that determines the seasonal AMOC variability.
The AMOC interannual variability is quantified in both in-situ observations at 26.5°N and numerical models. The observed AMOC interannual anomalies consist of an increase from early 2004 to late 2005 and a following downtrend which reaches a minimum in the winter of 2009/2010. These interannual AMOC fluctuations are dominated by changes in the upper mid-ocean geostrophic flow except during the winter of 2009/2010, when the anomalous wind-driven Ekman transport also has a significant contribution. The physical mechanisms for the interannual changes of the AMOC are proposed and evaluated in a two-layer model. While the Ekman transport is linked to the North Atlantic Oscillation (NAO), the anomalous geostrophic transport involves the oceanic adjustment to surface wind forcing. In particular, the intensification and weakening of the southward interior geostrophic flow is modulated by the internal Rossby wave adjustment to the surface wind forcing. The Gulf Stream, on the other hand, is controlled by both topographic waves along the US coast and westward propagating planetary waves. Our study suggests that a large part of the observed AMOC interannual variability at 26.5°N can be explained by wind-driven dynamics.

In addition, the basinwide AMOC responses to the interannual wind forcing are investigated in an eddy-resolving OGCM. The diagnostic analysis suggests that topographic waves and interior baroclinic Rossby waves play essential roles in modulating the AMOC interannual variability throughout the Atlantic basin. The proposed mechanisms are evaluated in a simple two-layer model. The high-latitude anomalies are communicated into the lower latitudes by topographic waves and account for about 50% of the AMOC interannual variability in the subtropics. The topographic waves and the large scale Rossby waves excited by wind forcing set up coherent AMOC
interannual variability across the tropics and subtropics. The comparisons between simple model and OGCM results suggest that a large fraction of interannual AMOC variations in the OGCM can be explained by wind-driven dynamics.
Dedication

This dissertation is dedicated to my parents, Dexin Zhao and Xuezhen Yu,
whose support and guidance that made me a better human being.
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Chapter 1

Introduction

The global meridional overturning circulation (MOC), also generally called the thermohaline circulation, is a large-scale, inter-basin circulation (Fig.1.1). To better view or diagnose the MOC, the zonal integration of meridional mass (volume) transport is usually used to eliminate the wind-driven gyre circulation and isolate the MOC. Figure 1.2 displays the schematic of the zonal integrated meridional flow in meridional and vertical dimensions. In this meridional cell, different water masses are redistributed and transformed: cold water at depths greater than 1000m is upwelled into the upper ocean and recirculated back to the deep water formation regions (Fig. 1.2).

The Atlantic meridional overturning circulation (AMOC) is a crucial component of the global ocean circulation and meridionally transports a large amount of heat and freshwater (Ganachaud and Wunsch 2003; Lumpkin and Speer 2007; Talley 2003; 2008). At 26°N where the oceanic circulation carries a northward heat flux of 1.3PW (1 PW =10^15W) (Lavin et al. 1998), the AMOC contributes about 88% of the basinwide meridional heat transport (Johns et al. 2011). The fluctuations of the heat transport associated with the AMOC lead to changes in the North Atlantic Ocean heat content that can have regional and large-scale climatic impacts (Cunningham et al., 2013). Previous studies showed that changes of the AMOC are directly linked to the inter-hemispheric sea surface temperature (SST) anomalies in the Atlantic and summertime climate of both North America and Western Europe (Knight et al., 2005; Sutton and Hodson, 2005; Zhang and Wang, 2013; Zhang et al. 2014). Therefore, understanding the variability of
AMOC and the corresponding mechanisms is vital to further progress in climate change studies.

Figure 1.1. The pathway of MOC and their net transport in isopycnal layers across hydrographic sections. Four layers are adopted: surface (red), intermediate (green), deep (blue) and bottom (dark gray). (Adapted from Talley 2008).
1.1 AMOC variability

Owing to incomplete long-term ocean observations, our knowledge about AMOC variations heavily relies on theoretical analysis and numerical models. After many years of study, it is well established from both in-situ observations and numerical simulations that the AMOC varies over broad time scales, ranging from daily to (multi)decadal and centennial (e.g. Delworth et al., 1993; Häkkinen, 1999; Köhl and Stammer, 2008). The variations on different frequency can be produced by external forcing factors, such as
momentum and buoyancy fluxes (Eden and Willebrand, 2001; Biastoch et al., 2008; Yeager and Danabasoglu, 2014), or by intrinsic variability linked to instability or mesoscale eddies (Hirschi et al., 2012; Thomas and Zhai, 2013). Much attention has been focused on determining the response of the AMOC to external forcing, with an aim to understand the mechanisms accounting for the AMOC variability. Buoyancy flux anomalies associated with different phases of the North Atlantic Oscillation (NAO) have been found to significantly affect the convective intensity and hence the deep-water formation in the subarctic region. Model studies suggest that variations in the formation rate of the water masses can trigger dynamical responses in the Deep Western Boundary Current (Böning et al., 2006) and consequently modulate the AMOC at lower latitudes either through a fast boundary wave response (Kawase, 1987; Johson and Marshall, 2002; Getzlaff et al., 2005) or by slower advective processes (Bower et al., 2009; Zhang, 2010; van Sebille et al., 2011). On the other hand, changes in the air-sea momentum flux can also generate substantial AMOC variability. The role of local wind forcing in driving the AMOC seasonal changes was noted and discussed by Bryan (1982) and Jayne and Marotzke (2001). They showed that an AMOC seasonal cycle can be produced by meridional Ekman transport at the surface and depth-independent compensational flow below. Recent observational and numerical studies point out that geostrophic transport also contributes substantially to the AMOC seasonality and that the anomalous geostrophic flow in the extratropics is controlled mainly by local wind forcing near the basin boundaries (Dong et al. 2009; Kanzow et al. 2010). In addition, wind stress curl variations excite baroclinic Rossby waves in the ocean interior that can amplify as they across the basin. These oceanic adjustments lead to wind driven Sverdrup transport...
anomalies and modify the meridional geostrophic transport in the upper 1000 m (Eden and Willebrand, 2001; Köhl, 2005). Importantly, as the westward propagating Rossby waves impinging on the western boundary they modify the pressure profiles and hence the AMOC strength (Hirschi et al., 2007; Cabanes et al., 2008; Sinha et al. 2013).

It should be emphasized that the AMOC responses to thermohaline and momentum forcing have different dominant time scales. The AMOC anomalies associated with subpolar deep water formation mostly occur on (multi) decadal time scales, whereas the wind-driven variations dominate the interannual and seasonal variability (Biastoch et al., 2008; Yeager and Danabasoglu, 2014). In order to understand the AMOC variations in observations and model hindcast simulations, the signals induced by buoyancy and wind forcing needs to be quantified separately. On the other hand, the magnitude, frequency and dominant dynamics of the thermohaline driven variability is model dependent due to the sub-grid scale parameterizations and model resolutions, which determine the representation of (sub)mesoscale flows and the Nordic Sea overflows (Willebrand et al. 2001; Yeager and Danabasoglu 2012; Danabasoglu et al. 2014). However, the AMOC fluctuations induced by momentum flux are more deterministic because the involved wind driven dynamics are generally linear process (Kanzow et al. 2010). This suggests that the dynamics responsible for the wind-driven AMOC variability are fundamental and can be understood regardless of model details.

1.2 Observations for the AMOC

Although the AMOC was found in observations several decades ago, its structure and variability were mainly derived either from synoptic trans-ocean basin ship-based
estimates or partial measurements at critical regions, such as the Deep Western Boundary Current (DWBC). The AMOC variability drawn from these data is subject to serious sampling errors. Since April 2004, the U.K.-U.S. Rapid Climate Change (RAPID-WATCH)/Meridional Overturning Circulation and Heat flux Array (MOCHA) basin-wide array (hereafter the 26.5°N array) has been deployed along 26.5°N in the Atlantic Ocean to measure the mean structure and temporal variability of AMOC. These unprecedented observational times series show that the AMOC has considerable fluctuations (Cunningham et al., 2007; Kanzow et al., 2007; McCarthy et al., 2012). With about 10 years of data to date, the variance spectrum of the AMOC at this latitude is, at least partly, established, so that it is necessary to fully understand the fundamental mechanisms responsible for the variability at different frequencies. Using the first four years of the time series, Kanzow et al. (2010) documented the seasonal cycle of the AMOC and concluded that the geostrophic mid-ocean and Gulf Stream transports dominate its seasonal variability. They also proposed that the wind stress curl forcing at the eastern boundary of the Atlantic largely controls the seasonal fluctuations of geostrophic mid-ocean transport. Interannual fluctuations of the AMOC at 26.5°N were described by McCarthy et al. (2012) who reported that the AMOC significantly decreased in 2009/2010. They further pointed out that the decline was caused by both a temporarily reversed Ekman transport and strengthening of the southward mid-ocean geostrophic flow.

1.3 Objectives

This purpose of this study is to quantify the wind-forced AMOC variability at different time scales and explore the corresponding physical mechanisms. In particular,
the AMOC variability derived from the 26.5°N array will be investigated. The dynamics diagnosed from the 26.5°N section will be evaluated throughout the Atlantic basin using a hierarchy of numerical models.

Chapter 3 explores the AMOC seasonal cycle in an eddy resolving Ocean General Circulation Model (OGCM), with a focus on the contribution from the geostrophic transport. In addition, a set of simpler models, including a forced Rossby wave model and a linear two-layer model, are used to help understand the dynamical processes accounting for the seasonal variability. Our results indicate that the seasonal AMOC cycle in the extratropics is controlled mainly by local boundary effects, where either the western or eastern boundary can be dominant at different latitudes, while in the northern tropics it is the interior flow and its lagged compensation by the western boundary current that determines the seasonal AMOC variability.

Chapter 4 focuses in detail on the physical mechanisms of the AMOC interannual variability at 26.5°N, within the framework of a relatively simple wind-driven two-layer model, to understand what aspects of the observed variations can be directly attributed to wind forcing, and what the underlying dynamics are. The observed AMOC interannual variability at 26.5°N, especially the downward trend between 2005 and 2009, is shown to be explained by wind-driven dynamics.

Chapter 5 is devoted to explore the nature of the basinwide AMOC interannual variability in an eddy-resolving OGCM. The meridional coherence of AMOC fluctuations are diagnosed, and the proposed mechanisms are evaluated in a simple two-layer model to help reveal the fundamental dynamics of the basinwide AMOC response to the interannual wind forcing. This chapter also investigates both local and remote
forcing effects, particularly regarding the AMOC remote responses in the subtropics and tropics to subpolar wind forcing. The ocean adjustments, including the topographic waves and interior baroclinic Rossby waves, to the interannual wind forcing is found to play essential roles in modulating the AMOC interannual variability throughout the Atlantic basin.
Chapter 2

Methodology

2.1 Observations

Daily measurements of AMOC at 26.5°N are obtained from the 26.5°N array. At this latitude, the upper branch of the AMOC is measured by three separate components: 1) The Gulf Stream (GS) transport $T_{GS}$, through the Straits of Florida; 2) the surface wind-driven Ekman transport, $T_{EK}$; and 3) the upper mid-ocean transport (UMO) across the basin between the Bahamas and Africa, $T_{UMO}$ (Kanzow et al., 2007; Rayner et al., 2011). $T_{GS}$ is estimated through the voltage records obtained from the submerged telephone cables between West Palm Beach and Grand Bahama Island (Baringer and Larsen, 2001). Direct observations of $T_{GS}$ from shipboard Lowered Acoustic Doppler Current Profiler and Dropsonde floats are also used to calibrate the cable measurements. $T_{EK}$ is derived from Cross-Calibrated Multi-platform Product (CCMP) winds (Atlas et al., 2011) according to $T_{EK} = - \int_{x_w}^{x_e} \frac{\tau_x}{\rho_f} dx$ where $x_w$, $x_e$, $\tau_x$, $\rho$ and $f$ are the western and eastern basin boundary, zonal wind stress, sea water density and Coriolis parameter, respectively.

In the trans-Atlantic mooring array, the mid ocean transport ($T_{MO}$) is calculated as the sum of three parts: 1) the meridional absolute transport over the Bahamian continental shelf (Johns et al., 2008); 2) the internal geostrophic flow calculated from the difference in full depth density profiles on either side of the basin; and 3) an external flow component which is the zonally integrated reference-level contribution to the mid-ocean geostrophic flow. It is computed from a constraint of no net mass transport across the
section, validated by independent bottom pressure measurements (Kanzow et al., 2007).

The part of the mid-ocean flow that occurs in the upper ocean, shallower than the depth of the AMOC streamfunction maximum (at about 1100 m) is referred to as $T_{UMO}$ (Kanzow et al., 2007). Therefore, at any given time the AMOC ($T_{AMOC}$) strength is given by the sum of $T_{GS}$, $T_{EK}$, and $T_{UMO}$.

2.2 Models

2.2.1 OFES

The general circulation model used here is the Ocean general circulation model For the Earth Simulator (OFES) based on the Modular Ocean Model (MOM3). It is configured nearly globally (75°S to 75°N) with horizontal resolution of 0.1°. There are 54 vertical z levels with thicknesses ranging from 5 m at the surface to 330 m for the bottom level (Masumoto et al. 2004). The model is spun up from rest for 50 years and forced with monthly mean climatological atmospheric fluxes derived from 1950-1998 NCEP/NCAR reanalysis (Kalnay et al. 1996). After that, the model is driven by daily mean NCEP/NCAR reanalysis data from 1950 to 2010 (Sasaki et al. 2008; Masumoto 2010). In the present study, the output over 1950-2010 is used to study the AMOC variability. The data is downloaded from Asia-Pacific Data-Research Center (APDRC) at http://apdrc.soest.hawaii.edu.
2.2.2 Two-layer model

This study also uses a two-layer model to simulate the wind-driven circulation in the Atlantic Ocean. The two-layer model consists of continuity and momentum equations as follows:

\[ \frac{\partial u_i}{\partial t} + u_i \frac{\partial u_i}{\partial x} + v_i \frac{\partial u_i}{\partial y} - f v_i = -P(x) + A V^2 u_i + F(x) - \frac{\lambda u_i}{H_i} \]

\[ \frac{\partial v_i}{\partial t} + u_i \frac{\partial v_i}{\partial x} + v_i \frac{\partial v_i}{\partial y} + f u_i = -P(y) + A V^2 v_i + F(y) - \frac{\lambda v_i}{H_i} \]

\[ \frac{\partial H_i}{\partial t} = \frac{\partial u_i H_i}{\partial x} + \frac{\partial v_i H_i}{\partial y} \]

where \( i=1 \) (upper layer) or 2 (lower layer), \( u,v \) are zonal and meridional velocity, \( f \) is Coriolis parameter, \( A=1.2 \times 10^3 m^2/s \) is a lateral viscosity coefficient, \( \lambda=-5 \times 10^{-5} \) is a bottom friction coefficient, and \( H \) is layer thickness. The bottom friction is usually applied in the lower layer and works in the top layer only when \( H_2 \) is zero. The pressure gradient term (\( P \)) is:

\[ P(k) = \begin{cases} \frac{g \partial h_i}{\partial k} & \text{for layer 1} \\ g \frac{\partial h_i}{\partial k} + g' \frac{\partial h_2}{\partial k} & \text{for layer 2} \end{cases} \]

\( k=x, \) or \( y. \)

where \( g \) is gravitational acceleration, \( g' = g \frac{\rho_2 - \rho_1}{\rho_2}, \rho_i \) is layer density \((i = 1 \text{ or } 2), h_1 \) is the sea surface height and \( h_2 \) is the interface anomaly.

The forcing term is concentrated in the upper layer

\[ F(k) = \begin{cases} \tau_k & \text{for layer 1} \\ 0 & \text{for layer 2} \end{cases} \]
where \( k = x \) or \( y \). The initial upper layer thickness is 1000m, and the density difference between the upper and lower layers is 1 kgm\(^{-3}\). The model domain covers the region from 40°S to 65°N and 100°W to 25°E with a uniform spatial resolution of 0.25°. Buffer zones are implemented at the southern and northern boundaries to damp artificial boundary waves. Bottom topography comes from the ETOPO5 (Earth Topography-5min) bathymetry data set interpolated onto the model grid, and the minimum water depth is 200 m.

2.3 Calculation of the AMOC

In both OFES and the \textit{in situ} data, the AMOC strength at certain latitude is defined as the maximum of the vertical streamfunction:

\[
\Psi (z, y, t) = \int_{x_w}^{x_e} \int_{z}^{0} \nu(x, y, z, t)dzdx
\]

where \( x_w, x_e \) are the western and eastern boundaries, and \( \nu(x, y, z, t) \) is the meridional velocity. The AMOC in the two-layer model is calculated from the upper layer meridional velocity. Similar to that derived from hydrographic sections (Talley et al. 2003; Lumpkin and Speer 2007), the mean structure of the meridional streamfunction in OFES (Fig. 2.1) is comprised of two meridional overturning cells located between the surface and about 3000 m, and from 3000 m to the bottom, respectively. The 1980-2009 averaged AMOC strength has maximum of 18 Sv at 35°N. At 26.5°N, the mean AMOC is 16 Sv which is slightly weaker than the measured seven-year mean value of 17.4 Sv from the 26.5°N array between Apr. 2004 and Apr. 2011 (Zhao and Johns 2014a). This study is mainly focused on the upper cell which involves the northward flow within approximately the top 1100 m and the compensating southward flow from 1100 m to 3000 m.
Figure 2.1. The mean structure of the AMOC simulated in OFES (1980-2010). Unit: Sv.
Chapter 3

Wind driven seasonal cycle of the AMOC

In the variance spectrum of the AMOC, the seasonal variability has a prominent contribution. Various numerical models with different resolutions and forcings suggest that the fluctuations of AMOC on time scales shorter than a year are more energetic than at other frequencies, especially south of 40°N (e.g. Bingham et al. 2007). Based on the in-situ measurements, Kanzow et al. (2010) showed that local wind stress curl along the eastern boundary is the main factor modifying the density profiles there on seasonal time scales, and the subsequent seasonal cycle of the AMOC at 26.5°N. However, it remains unknown if similar dynamics occur in other regions such as in the tropical ocean or the South Atlantic. In this section, the AMOC seasonal cycle simulated by the OFES model is studied. The physics governing the seasonal fluctuations of the AMOC are explored in detail at three latitudes (26.5°N, 6°N, and 34.5°S).

To help interpret the dynamics involved in the seasonal AMOC cycle, the two-layer model described in chapter 2 is linearized and forced by the same winds used in the OFES simulation (i.e. NCEP/NCAR reanalysis). The two-layer model is integrated for 25 years and the results from last 5 years are used to derive a climatological annual cycle. There is no thermohaline forcing in the two-layer model, and therefore its time-varying AMOC is driven only by winds, and it is also missing the large scale thermohaline circulation present in the OFES model.
In addition, daily measurements of AMOC from 26.5°N array between April 2004 and April 2011 are used to evaluate the numerical results. Following the AMOC decomposition of the observational data (Rayner et al. 2011), the northward upper ocean flow of the AMOC is decomposed into its three main components: the Gulf Stream transport through the Straits of Florida (GS), the Ekman transport in the top 100 m (EK), and the upper mid-ocean transport between the Bahamas and Africa (UMO). Further details about the estimation of each component may be found in Rayner et al. (2011).

3.1 Seasonal cycle of the AMOC

3.1.1 Seasonal variability in OFES

The AMOC estimated from the OFES between 1980 and 2010 is used to derive a climatological annual cycle. Figure 3.1 shows the distribution of the AMOC anomaly with respect to latitude and month. The most significant seasonal variability takes place in the tropical Atlantic region with 11 Sv peak to peak amplitude, and a maximum (minimum) in boreal winter (fall). Outside of the tropics, the amplitudes are much weaker (4-6 Sv) and the phases are also different from that in the tropical ocean. While the northern hemisphere subtropics has broad maximum in boreal summer and fall and minimum in winter and spring, the southern subtropical ocean reaches its maximum in austral winter and minimum in austral fall. In most latitudes, the phases are generally coherent within about 10-15°, meaning that the seasonal cycle along one latitude is representative for about a 10-15° degree meridional extent. The amplitudes of the seasonal variability are significant in comparison to the AMOC variability on interannual
time scales in the model, which have r.m.s. values ranging from 0.5 Sv - 1.5 Sv at different latitudes.

On seasonal time scales, the meridional flow can be split into Ekman and geostrophic components provided that the non-Ekman ageostrophic flow is negligible. This assumption is confirmed by Baehr et al. (2004) and Hirschi and Marotzke (2007) who demonstrated that the thermal wind and the Ekman contributions are the dominant terms in the force balance governing the meridional flow, with generally smaller contributions from other terms.

The Ekman contribution is calculated from zonal wind stress:

\[ \Psi_{ek}(y, t) = \int_{x_w}^{x_e} \frac{1}{\rho_f} \tau_x(x, y, t) dx \]

where \( \tau_x, \rho, f, x_w \) and \( x_e \) are the zonal wind stress, reference density, Coriolis parameter, and western and eastern boundaries, respectively. The geostrophic contribution to the AMOC is estimated by removal of the Ekman transport from the total meridional transport above \( h_{moc} \), where \( h_{moc} \) is the depth where the maximum \( \Psi \) occurs.

As shown in Figure 3.1, both the Ekman and the geostrophic anomalies have substantial changes throughout the year, indicating that both contribute to the seasonal cycle of the AMOC. Similar to the AMOC, their peak to peak amplitudes are largest in the tropical ocean and decrease toward higher latitudes. In the tropics, the Ekman anomaly is characterized by positive values for the first half of the year and negative values in the second half of the year. The corresponding geostrophic variability is nearly out of phase with the Ekman contribution in the northern tropics but has a more complicated relationship in the southern tropics. In the subtropics, the phase of the
geostrophic component shifts by 2-3 months relative to the tropics and shows an asymmetry between the southern and northern hemispheres.

The amplitude of the Ekman transport anomalies is generally larger than the geostrophic transport, but due to varying phase relationships between them, the total AMOC variability at any latitude can be either enhanced or diminished relative to the individual components. For example, in the tropics near 8°N, the amplitudes of Ekman and geostrophic anomalies are 15 Sv and 8.5 Sv, respectively, but the AMOC only has an 11 Sv peak to peak fluctuation. On the other hand, AMOC has stronger variability than either the Ekman or the geostrophic transport in subtropical regions, indicating less compensation between them.

3.1.2 Comparison to observations at 26.5°N

To examine if the OFES results are consistent with observations, the climatological seasonal variability of the AMOC from the monthly mean OFES time series and the 26.5°N array are compared. The seasonal cycle of the observed MOC (Fig. 3.2a) exhibits a 6 Sv peak to peak amplitude with a maximum northward flow in July and October and a minimum in March. A similar phase occurs in OFES (Fig. 3.3a) but with smaller amplitude (4 Sv). While the seasonal cycle of OFES shown in Figure 3.3 is derived from 30 years of data (1980-2009), the seasonal cycle derived from the same time period as the in situ observations (2004-2009) gives essentially the same result.

The breakdown of the AMOC seasonal cycle at 26.5°N into its three main components (GS, EK, UMO) for both the observations and OFES is also shown in Figures 3.2 and 3.3. The GS refers to the transport carried by the Florida Current. The
mean strength of the observed GS is 31.5 Sv and its seasonal cycle (Fig. 3.2b) shows an
annual peak-to-peak range of 4 Sv with maximum in July and minimum in November.
Note that the seasonal cycle of the GS shown here is slightly stronger than that in
Kanzow et al. (2010) because of the longer period used in our study. Nevertheless, both
the amplitude and phase displayed here are consistent with the seasonal cycle computed
from the 26-year long time series of the Florida Current (Kanzow et al. 2010). The
Ekman transport varies from 1.2 Sv in February to 4.6 Sv in July which is slightly
different from Kanzow et al. (2010). This is mainly due to the longer time period
included in our estimate of the seasonal cycle at 26.5°N (2004-2011 versus 2004-2008 in
Kanzow et al.), and to the substantial zonal wind stress anomalies that occurred between
Dec. 2009 and Mar. 2010 associated with a strong negative phase of the NAO (McCarthy
et al. 2012). The seasonal change of UMO is about 5 Sv, from -18.9 Sv in Mar. to -13.8
Sv in Oct. The sum of peak to peak variations of the three separate components is 12.5
Sv, much larger than that of the AMOC (6 Sv), indicating that compensation occurs
among the three individual components.

The same decomposition applied to the OFES model (Fig. 3.3) shows that while the
GS in OFES has an annual cycle similar to the observations, both the peak-to-peak
amplitude (1.3 Sv) and the mean strength (22.3 Sv) are much smaller. Like the GS, the
seasonal cycle of UMO in OFES displays similar phase with that in observations, but
with weaker amplitude (2.5 Sv) and weaker annual mean value (-10 Sv). The reason for
this discrepancy is that, in OFES, part of the western boundary current does not flow
through the Straits of Florida; instead it is located east of the Bahamas where it is
included here in the UMO transport. In the OFES model the northward transport in this
"Antilles" Current is about 14 Sv in the top 1100 m. In the real ocean, the majority of the western boundary flow is through the Straits of Florida, and only a 6 Sv northward Antilles Current occurs east of Abaco, Bahamas (Lee et al. 1996; Johns et al. 2008). The total northward western boundary flow in OFES is about 36 Sv and nearly equal to the sum of the observed Florida Current (31.5 Sv) and Antilles Current (6 Sv) transports. The weaker UMO seasonal amplitudes in OFES compared to the observations is attributed to the too weak wind stress curl near the eastern boundary in the OFES model, and a detailed explanation is given in the following section. The phase and amplitude of the Ekman transport in OFES generally agrees with the observed annual cycle, i.e., a peak in July and minimum in March. Their differences can be attributed to the different wind products used (NCEP reanalysis for OFES and Cross-Calibrated Multi-Product (CCMP) winds for the 26.5°N array), and in part to the different time period used in OFES (1980-2010) and the 26.5°N array (2004-2011).

3.1.3 Dynamics of the seasonal cycle

In the following, we will investigate the main physical processes governing the AMOC seasonal cycle at different latitudes. It is well known that the meridional Ekman transport is directly driven by the zonal wind which is closely linked to changes in atmospheric circulation. For instance, the seasonal fluctuations of the tropical Ekman transport are mainly associated with the annual migration of the intertropical convergence zone (ITCZ). These changes in Ekman transport affect the AMOC directly and immediately through changes in the meridional surface layer transport. On the other hand, the variations of the geostrophic transport involve both barotropic and baroclinic
adjustment. The barotropic flow adjusts rapidly to changes in atmospheric forcing and approaches a quasi-equilibrium "topographic Sverdrup" balance after transients such as barotropic basin modes die out (Anderson et al. 1979). The structure of this barotropic response depends crucially on the actual basin topography. On the other hand, the baroclinic flow involves both the dynamics in a flat bottom ocean and the influences of topography. Locally, the baroclinic geostrophic velocity can be estimated with density profile pairs and a level of ‘no-motion’. The basin-wide profile of the baroclinic geostrophic transport is determined by the density difference between the endpoints of the basin (Baehr et al. 2004). Consequently, variations in the AMOC associated with baroclinic flow changes are closely linked to the changes of the density profiles at the basin boundaries.

Western boundary currents (WBCs) provide a good example of how topography affects both barotropic and baroclinic adjustment. WBCs are usually located, at least partly, in relatively shallow regions, such as GS in the Straits of Florida. Their mean strength is determined to first order by the Sverdrup balance established by baroclinic planetary waves, which is not strongly affected by topography (Anderson and Gill 1975; Anderson and Killworth 1977). However, WBC fluctuations on periods much shorter than the time taken for baroclinic adjustment (several years) are largely disconnected from the baroclinic planetary waves. For instance, the seasonal variability of the GS at 26.5°N is largely controlled by the forcing north of the Straits of Florida through boundary wave processes (Anderson and Corry 1985; Atkinson et al. 2010; Czeschel et al. 2012).
To investigate the above dynamics in a simple context, a linear two-layer model with realistic topography is configured to simulate the seasonal cycle of the AMOC. The two-layer model, described in chapter 2, includes both barotropic and baroclinic adjustment induced by wind forcing. As illustrated in Figure 3.4, the seasonal cycle of AMOC and geostrophic transport in the two-layer model generally agrees with those of OFES (compare to Fig. 3.1). In particular, the phases and amplitudes of the seasonal cycle in both tropics and subtropics are well reproduced. The main difference between the two-layer model and OFES is at higher latitudes in the North Atlantic, where the seasonal variability is underestimated. This is probably the result of the damping applied along the northern boundary to prevent artificial boundary wave propagation effects, but could also be due to nonlinear processes associated with the Gulf Stream in OFES. Nevertheless, the similarity in both phases and amplitudes in OFES and the two-layer model demonstrates that the linear two-layer model successfully captures the fundamental features of the seasonal AMOC variation.

3.1.4 Roles of barotropic and baroclinic processes

To better understand the dynamics contained in both OFES and the two-layer model, we explore the roles of barotropic and baroclinic processes in modulating the seasonal variability of the AMOC. Since the same wind forcing is used in both models, their Ekman transports are exactly the same and we will focus on the comparison of meridional geostrophic transport in both models. To perform a comparative analysis between the two models, we define an "upper layer" and "lower layer" geostrophic
velocity in OFES as the respective vertically-averaged flows above and below the depth of the maximum of the AMOC streamfunction, $h_{moc}$, at any given time:

$$v_1(x, y, t) = \frac{1}{h_{moc}} \int_{h_{moc}}^{0} (v(x, y, z, t) - v_{ek}(x, y, z, t)) dz$$

$$v_2(x, y, t) = \frac{1}{H - h_{moc}} \int_{H}^{h_{moc}} v(x, y, z, t) dz$$

where $H$ is the local water depth, $v$ is the full velocity, $v_{ek}$ is Ekman velocity.

We further define a “baroclinic” velocity for the upper layer flow as $v' = v_1 - v_2$. Thus, in OFES, for any region with bottom depths shallower than $h_{moc}$, $v_2 = 0$ and $v' = v_1$. For the two-layer model, $v_1$ and $v_2$ are simply equal to the upper and lower layer velocity and $v'$ is defined as above.

In what follows, we will use the upper layer geostrophic streamfunction accumulated from the eastern boundary:

$$\Psi_1(x, t) = \int_{x}^{x_e} \int_{h_{moc}}^{0} v_1(x, t) dz dx = h_{moc} \int_{x}^{x_e} v_1(x, t) dx$$

as a primary tool in evaluating the AMOC variability. According to the above definitions, we can also break this down into a "baroclinic" component:

$$\Psi_{bc}(x, t) = \int_{x}^{x_e} \int_{h_{moc}}^{0} v'(x, t) dz dx = h_{moc} \int_{x}^{x_e} v'(x, t) dx$$

and a "barotropic" component:

$$\Psi_{bt}(x, t) = \int_{x}^{x_e} \int_{h_{moc}}^{0} v_2(x, t) dz dx = h_{moc} \int_{x}^{x_e} v_2(x, t) dx$$

where the latter represents the part of the upper layer meridional flow that is common to both layers.
As a further aid in interpreting the results, we show in Figure 3.5 the monthly anomalies of the upper layer baroclinic velocity ($v'$) and the associated interface anomaly ($h_2$) from the two-layer model for January, April, July and October. In the following sections, we select three representative latitudes to diagnose the AMOC variability in detail: two in the northern and southern hemisphere extratropics (26.5°N and 34.5°S), and one in the northern tropics (6°N), which are denoted in Figure 3.5a.

3.2 AMOC seasonal cycle at 26.5°N

3.2.1 Seasonal variability of the upper layer streamfunction

Figures 3.6a and 3.7a show the seasonal anomalies of the upper layer cumulative streamfunction $\Psi_1$ at 26.5°N in OFES and the two-layer model, respectively. Both models show a similar basin-wide pattern with predominantly negative anomalies in the first half of the year (Jan - June) and positive anomalies in the second half of the year (July - Nov). Interspersed within these broad zonal patterns are local maxima and minima that are tilted westward (which is more clearly evident in the OFES model), indicating westward propagation. The OFES model pattern is generally noisier than the two-layer model, which can be attributed to the fact that it is interannually forced, rather than forced by a repeating seasonal cycle as in the two-layer model, and also because it is an eddy resolving model and therefore contains random mesoscale features that can leave a residual imprint on the climatological annual cycle. Otherwise the two patterns are quite similar.

Before discussing these results further, we refer the reader to Figure 3.8 which helps to better explain the nature of these seasonal $\Psi_1$ patterns. Figure 3.8a shows the time
series of local upper layer transport (i.e., $v_1(x, t)$ multiplied by $h_{mol}$ and dx) from the last 5 years of the two-layer model simulation at 26.5°N, while Figure 3.8b shows the associated pattern for the cumulative streamfunction $\Psi_1$. Near the eastern boundary, the local meridional transport has a strong seasonal variation with minimum in April and maximum in October (Fig. 3.8a). This variability is forced by a strong annual cycle of wind stress curl at the eastern boundary through the same mechanism described previously by Kanzow et al. (2010), that is, by seasonal uplift and depression of the thermocline near the coast that drives a seasonally varying eastern boundary flow.

Emanating from the western edge of this boundary zone are meridional velocity anomalies (Fig. 3.8a) associated with annual baroclinic Rossby waves that propagate westward at a first baroclinic mode wave speed (the only baroclinic mode supported by the two-layer model). At 26.5°N these waves interact with the Canary Islands just west of the eastern boundary, which locally disturbs the westward phase propagation between about 17-19°W. This is a very localized effect that is not present at neighboring latitudes, where the Rossby wave crests and troughs propagate smoothly away from the eastern boundary, and it has no consequential effect on the basinwide transport.

As the Rossby waves propagate across the ocean interior, their signal is affected by the local wind stress curl forcing along the wave paths and interactions with topography, so that their amplitude is modified, and also they are weakly damped by both lateral and bottom friction. The cumulative streamfunction shown in Figure 3.8b reflects, in the east, the initial contribution from the annually reversing eastern boundary flow, and then it is modulated toward the west by zonal integration across the crests and troughs of the propagating Rossby wave pattern.
As the waves approach the Bahamas (77°W), their meridional velocity signal is amplified by interaction with the Bahamas boundary, which leads to increased amplitudes near 75°W and steadily decreasing amplitudes from there to the Bahamas, with a nearly stationary phase across this boundary region (Fig. 3.8a). This behavior is consistent with expectations from linear Rossby wave reflection at the western boundary, due to the superposition of the long waves and reflected short Rossby waves (Longuet-Higgins 1964; Pedlosky 1987). Local wind stress curl forcing near the Bahamas can also contribute to the meridional transport anomalies here, but this effect is relatively weak compared to the wave reflection/interaction process. The net effect of this wave reflection process is to strongly reduce the amplitude of the pressure anomaly and hence the cumulative streamfunction anomaly ($\Psi_1$) at the Bahamas boundary, due to compensation of the zonally integrated flow across the rest of the basin by an opposing meridional flow near the boundary. This has been clearly demonstrated in observations, as well as in a theoretical/modeling context, by Kanzow et al. (2009). The cumulative streamfunction anomaly at the Bahamas boundary is typically reduced by about a factor of three relative to its seasonal amplitude a few degrees into the interior. As shown by Kanzow et al. (2009), local boundary waves are involved in this process as well as short Rossby waves, and therefore it should be described in general as a western boundary wave reflection/interaction process.

A further, but relatively small, modification of the cumulative streamfunction occurs across the Straits of Florida due to the seasonal cycle of the GS transport, leading to the total AMOC seasonal cycle shown at the western boundary in Figure 3.8b. (Note that the pattern of $\Psi_1$ shown in Fig. 3.7a is exactly the same as that of any of the individual years
of Fig. 3.8b.) The local meridional transport and $\Psi_1$ over multiple years in OFES (not shown) has a similar structure to that shown in Figure 3.8, but is more variable due to the interannual forcing.

Returning to Figures 3.6 and 3.7, we show in the left panels of these figures the role of different regions in determining the seasonal variability of the zonally integrated meridional geostrophic transport. These curves show the seasonal cycle near the eastern basin boundary, western point of the interior ocean, and western basin boundary. A point twice the Rossby deformation radius away from the eastern basin boundary is selected as the eastern point, while the Bahamas Islands are chosen as the western point of the interior ocean. From the eastern boundary to about 17°W, the geostrophic transport sets up a seasonal cycle with maximum in September-November and minimum in March-May. This feature is consistent in both numerical models (blue line in Fig. 3.6 and Fig. 3.7) and is forced by the wind stress curl near the eastern boundary (Fig. 3.7d). West of 17°W, the seasonal variability is modified until it reaches a relatively stable amplitude near 77°W. The fluctuations of the amplitudes have a spatial scale about 15° in longitude for both models. As noted above, this westward propagating pattern is induced by the meridional velocity associated with westward propagating long Rossby waves (Fig. 3.8). Although the wave like structure leads to substantial changes of the local meridional transport in the interior, when integrated across the full width of the basin, including the effects of wave reflection at the western boundary, the cumulative streamfunction anomaly at the western edge of the interior ocean (the Bahamas) is not greatly changed from that at the offshore edge of the eastern boundary layer. West of 77°W, the zonally integrated transport is further modified due to the impact of the Gulf Stream. The
differences between the green line and black line in Figures 3.6a and 3.7a indicates the seasonal cycle of the Gulf Stream (red line).

It is worthwhile to note that while the exact phase of the wave structures across the interior (and the number of maxima and minima in the cumulative streamfunction) depends on the first mode baroclinic wave speed, the zonally-integrated streamfunction anomaly at the western edge of the interior (the Bahamas) is largely insensitive to this. We have tested this in the two-layer model using different choices of the stratification parameter ($g'$), which controls the first mode wave speed, and find no significant differences in the cumulative streamfunction anomaly at the Bahamas when varying this wave speed over a factor of three. This is again because of the strong compensating effect produced by the wave reflection process: this largely nullifies the meridional transport anomaly associated with the waves impinging on the Bahamas, irrespective of the exact phase of the impinging wave.

### 3.2.2 Barotropic and baroclinic contributions to the upper layer streamfunction

In Figures 3.6 and 3.7, we show in addition to $\Psi_1$ the contributions of the barotropic ($\Psi_{bt}$) and baroclinic ($\Psi_{bc}$) components of the upper layer streamfunction in both models. In contrast to the upper meridional geostrophic transport ($\Psi_1$), $\Psi_{bt}$ has weaker seasonal variations of approximately ±2 Sv amplitude across the interior, and $\Psi_{bt}$ makes a negligible contribution to $\Psi_1$ when integrated across the full basin. The phase of the $\Psi_{bt}$ anomalies across the interior is different from $\Psi_1$, and shows maximum positive values in fall and negative values in winter. Unlike the Rossby wave anomalies that dominate the interior pattern of $\Psi_1$, $\Psi_{bt}$ is associated with the seasonal spin up and spin
down of a large scale barotropic gyre in the North Atlantic that is driven by the large scale seasonal WSC forcing. As shown in Figures 3.9a and 3.9b, the barotropic gyre is largely closed over the deep part of ocean and does not extend up onto the continental shelves, so that there is no significant seasonal variability at the eastern boundary, nor at the western interior point or at the western boundary (Fig. 3.6c and Fig. 3.7c). As a result, the barotropic adjustment does not contribute to the seasonal cycle of the meridional geostrophic transport at 26.5°N.

The overall pattern of $\Psi_1$ is very similar to $\Psi_{bc}$, indicating that the oscillations in Figure 3.6a and Figure 3.7a are produced mainly by baroclinic processes. Near the eastern boundary, the minimum and maximum of the $\Psi_{bc}$ (blue line Fig. 3.6b and Fig. 3.7b) are consistent with the uplift and depression of the thermocline along the eastern boundary (Fig. 3.5b, Fig. 3.5d). The interior thermocline anomalies shown in Figure 3.5 are associated with the westward propagating Rossby wave patterns present in $\Psi_{bc}$. The speeds detected from a random transformation of the hovmoller diagram of meridional velocity in OFES and the two-layer model are about -0.04 m/s and -0.042 m/s, respectively. They are consistent with the typical value for a first mode baroclinic Rossby wave at 26.5°N. The annual cycle of the GS is also seen in the anomalous thermocline changes along the western boundary in July and October (Fig. 3.5c, Fig. 3.5d). The thermocline anomalies along both boundaries are meridionally coherent within about 15°, a scale consistent with that detected in Figure 3.1. Therefore, the dynamics evaluated at 26.5°N is representative for the northern subtropical region.
3.2.3 Comparison of Gulf Stream, Ekman, and mid-ocean transports

In accordance to the in situ observations by the 26.5°N array, we split the MOC from the two-layer model into GS, EK, and UMO components and compare them to OFES (Fig. 3.3). Note that the two-layer model does not include the thermohaline circulation, so the annual mean value of each component from OFES is added to the two-layer model result. As shown in Figure 3.3, both the amplitude and phase for the seasonal varying MOC in the two models agree very well. Since the same winds are used in both models, their Ekman transports exhibit exactly the same annual cycle. Peak GS transport occurs in July in both models, while the seasonal minimum in the two-layer model occurs in October, a month earlier than the November minimum in OFES. This difference could be induced by neglected nonlinear advective effects in the two-layer model; also the simple stratification parameter in the two-layer model might not exactly capture the correct phase speed of the coastal baroclinic waves which are believed to play a role in the seasonal cycle of the GS (Anderson and Corry 1985). Nevertheless, the essential characteristics of the seasonal GS are well captured by the two-layer model, suggesting that the seasonal cycle of the GS in OFES is mainly driven by linear processes. This is consistent with the study of Anderson and Corry (1985) who also used a linear model and reproduced most of the observed GS seasonality. For the UMO, the two-layer model is able to produce the minimum in spring and maximum in fall. Although the two-layer model has slightly lower minimum in spring, its overall cycle is consistent with OFES and the in situ observations.
3.2.4 Analogy with a simple forced Rossby wave model

The dynamics underlying this UMO response can be further investigated using a simple linear forced Rossby wave model. On time scales longer than the inertial period, the oceanic response to wind stress curl forcing can be expressed by

$$\frac{\partial P_n}{\partial t} - \beta c_n^2 f^{-2} \frac{\partial P_n}{\partial x} = -c_n^2 f^{-1} F_n \nabla \times \tau$$

Where $P_n$ is pressure for the $n$th vertical mode, $t$ is time, $x$ is the distance eastward, $f$ is the Coriolis parameter, $\beta$ is the planetary vorticity gradient, $c_n$ is the $n$th mode long gravity wave speed, $\tau$ is wind stress, and $F_n$ is the projection of the forcing onto vertical modes (Sturges et al. 1998):

$$F_n = \frac{1}{D_{mix}} \int_{-D_{mix}}^{0} \varphi_n(z)dz / \int_{-D}^{0} \varphi_n(z)^2 dz.$$  

Here, $D$ is water depth, $D_{mix}$ is the surface mixed layer depth, and $\varphi_n(z)$ is the nth vertical mode eigenfunction, which is calculated from the OFES hydrographic data along 26.5°N. $\varphi_n$ and $c_n$ are chosen from a representative longitude (60°W) for the calculation. Only the first baroclinic mode is used here, i.e. $n=1$. The equation is integrated from zero initial conditions using the climatological seasonal cycle of wind stress curl anomaly at 26.5°N extracted from the OFES 1980-2009 monthly fields. An equilibrium seasonal cycle of $P_1$ is reached after about 20 years of integration.

The meridional geostrophic transport above $h_{moc}$ can be calculated according to:

$$T(x) = \int_{h_{moc}}^{0} v(x,z)dz = \int_{h_{moc}}^{0} \frac{1}{\rho_f} \frac{\partial P_1}{\partial x} \varphi_1(z)dz$$
and the streamfunction computed from $T(x)$ is shown Figure 11a. Its spatial pattern and phase are overall consistent with that in OFES and the two-layer model (Fig. 3.6b, Fig. 3.7b). Near the eastern boundary the seasonality is similar to that in the two-layer model and OFES, in terms of both phase and amplitude. This suggests that the forced Rossby wave model captures the essential physics accounting for the upper layer transport variations near the eastern boundary. However, the seasonal cycle at the western edge of the interior in the Rossby wave model has weaker amplitude than that in OFES and the two-layer model. The reason is that, unlike the two-layer model or OFES, the Rossby wave model cannot account for any wave reflection/interaction processes at the western boundary. Therefore, the zonally integrated meridional transport in the Rossby wave model is dependent on the particular phasing of the waves as they strike the Bahamas, and its seasonal cycle is sensitive to the wave speed and basin width. In the case shown (Fig. 3.10), the phasing of the wave arrivals at the western boundary leads to a relatively weak basinwide transport anomaly, but this could easily be much larger (or smaller) if the phase speed of the waves was slightly changed. The wave reflection in the OFES and the two-layer model largely reduces the effect of the waves impinging onto the western boundary, resulting in a seasonal cycle of UMO in the two-layer model that has similar phase and amplitude as that of the meridional transport near the eastern boundary in the Rossby wave model (Fig. 3.10b). This further confirms that the uplift (depression) of the thermocline forced by wind stress curl in the east has a dominant effect on the seasonality of UMO.
3.2.5 Sensitivity to wind forcing

Chidichimo et al. (2010) and Kanzow et al. (2010) found that the observed seasonal cycle of interior flow has 6.7 Sv peak to peak amplitude and the pressure (density) anomalies at the eastern boundary leads to a seasonal cycle of 5.2 Sv. With the same type of Rossby wave model used here, Kanzow et al. (2010) successfully reproduced a seasonal cycle with comparable amplitude and phase to the observations. The seasonal cycle obtained by our study, however, has much weaker amplitude than that in Kanzow et al. (2010). The main reason for the different amplitudes in our study and theirs is that different wind fields are used. Kanzow et al. (2010) used wind stresses derived from the Scatterometer Climatology of Ocean Winds (SCOW) which correctly preserves the WSC near the land/ocean boundaries (Risien and Chelton 2008), whereas both OFES and the two-layer model use the NCEP wind field. The SCOW winds have a seasonal WSC variation at the eastern boundary that is about 2-3 times stronger than that in NCEP. This causes a much larger local response of the density field at the eastern boundary and a much larger seasonal AMOC variation. To verify this, the SCOW winds are also used to force the linear Rossby wave model and the two-layer model, and their corresponding seasonal cycles are shown for comparison in Figure 3.10b. In contrast to the NCEP dataset, the SCOW winds produce a seasonal cycle with more comparable amplitude to the observed AMOC seasonal variability in both the Rossby wave model and the two-layer model. Nearly all of this difference is due to the differences in the SCOW and NCEP winds near the eastern boundary. Therefore, the weaker boundary wind stress curl in the NCEP
dataset is the main reason that OFES as well as the two-layer model underestimate the observed MOC seasonal variability at 26.5°N (Fig. 3.3).

3.3 AMOC seasonal cycle at 6°N

6°N is selected to study the seasonal variations in the tropical Atlantic Ocean where both the Ekman and geostrophic transport have large seasonal fluctuations. As shown in Figure 3.1, the Ekman transport has a seasonal cycle of 14 Sv peak to peak amplitude with a maximum in January and minimum in August. The annual peak to peak amplitude for the geostrophic transport is 9 Sv and its minimum is in May and maximum in October. Similar seasonality can be found in the two-layer model results, suggesting that the variations in both Ekman and geostrophic transport at 6°N are largely controlled by the wind forcing. In fact, the seasonal wind stress forcing in this region is controlled mainly by the annual migration of the intertropical convergence zone (ITCZ), which not only changes the zonal wind stress but also modifies the wind stress curl across the basin interior. The variable wind stress curl modifies the interior flows which feed the western boundary current, the North Brazil Current (NBC) (Johns et al. 1998). Similar to 26.5°N, the seasonal cycle near the eastern basin boundary, western point of the interior ocean, and western basin boundary are selected in both OFES and two-layer model. A point twice the Rossby deformation radius away from the eastern basin boundary is selected as the eastern point, while the location of the maximum (minimum) of the interior streamfunction anomaly is chosen as the western point of the interior ocean. The latter choice corresponds roughly to the outer edge of the western boundary current in each
simulation, since, unlike 26.5°N, there is no geographic feature (i.e., the Bahamas) that clearly separates the interior domain from the western boundary domain.

### 3.3.1 Seasonal variability of the upper layer streamfunction

Figure 3.11 displays the seasonal anomaly of $\Psi_1$ in OFES. Near the eastern boundary there is no significant seasonal variability (blue line), because the wind stress curl near the eastern boundary is relatively weak (Fig. 3.12c). (Notably, the wind stress curl shown here is very similar to the SCOW data.) From the eastern boundary to 30°W, $\Psi_1$ has a maximum in February and a minimum in November, while from 30°W to 45°W, the seasonal cycle is modified and has semiannual variation. From 45°W to 50°W, the seasonal fluctuations are further changed by the retroflexion of the NBC and give rise to an annual cycle maximum in June and minimum in September. The NBC retroflexion also induces large interannual variations so that the seasonal cycles near 48°W are not robust. The main part of the NBC is located between 50°W and 52°W, and at its inshore edge (at ~52°W), the seasonal variability of the meridional transport quickly and completely shifts to a cycle with minimum in April and maximum in October, which remains nearly constant from there to the western boundary. This implies that the NBC plays a fundamental role in shaping the seasonal cycle of the cumulative geostrophic transport across the basin at 6°N.

In the two-layer model (Fig. 3.12), a very similar annual cycle occurs for the cumulative geostrophic transport across the full section, but there are significant differences in the interior. East of 45°W the two-layer model does not reproduce the seasonal variations in OFES. This is because the local meridional transport in OFES is strongly affected by the
nonlinear retroflection of the NBC and the large-scale interior meandering pattern of the North Equatorial Current, which is not represented in the linear two-layer model (e.g., Fig. 3.5). However, the seasonal cycle for the cumulative interior flow (green line in Fig. 3.11a and Fig. 3.12a) is consistent in both models, and westward of this point they remain very similar, leading to a nearly identical seasonal variability in the zonally integrated flow (black lines in Fig. 3.11a and Fig. 3.12a). As shown in Figure 3.11b and Figure 3.12b, \( \Psi_{bc} \) has similar spatial structure and temporal variations to that of the upper layer geostrophic transport (\( \Psi_1 \)). \( \Psi_{bc} \) has much weaker variability (not shown), indicating again that the seasonal fluctuations of the upper geostrophic transport are mostly controlled by baroclinic processes.

### 3.3.2 Role of the North Brazil Current

The seasonal variability in both \( \Psi_1 \) and \( \Psi_{bc} \) is a well-known feature of the tropics and is a consequence of the rapid adjustment of the tropical ocean by baroclinic planetary waves (Philander and Pacanowski 1980). The interior geostrophic flow is forced by the WSC associated with the ITCZ migration. It takes several months for the changes in the interior ocean to be communicated into the WBC. In February (July) anomalous northward (southward) interior geostrophic flow occurs east of 30°W (Fig. 3.12a, b), but the NBC does not reach its corresponding seasonal maximum (minimum) until May (September), respectively (red line in Fig. 3.11a and Fig. 3.12a). The seasonally changing circulation patterns associated with this are shown in Figure 3.5, where 6°N lies near the boundary between the oscillating basin-scale patterns to the north and south. A maximum cyclonic (anticyclonic) circulation anomaly occurs to the north of 6°N during boreal
spring (fall), with opposite circulation anomalies to the south, which correspond to the spin-up and spin-down of the tropical and equatorial upper ocean gyres through the action of the baroclinic planetary waves. Figure 3.12d shows the NBC transport cycle in OFES, two-layer model and that calculated from a stationary Sverdrup balance. The NBC in OFES has minimum in spring and maximum in fall, a cycle very similar to that of the basinwide upper geostrophic transport. Due to the lack of a thermohaline component in the two-layer model, it does not reproduce the mean NBC transport, but the essential features of the seasonal cycle are well captured by the linear wind-driven two-layer model. The difference between the NBC in the Sverdrup cycle and two-layer model reflects the delayed response to the remote wind stress curl due to the propagation of the baroclinic waves. Particularly, the spring minimum is absent in the Sverdrup cycle. The lag of the NBC with respect to the interior leads to an “excess” NBC transport in fall and a corresponding “deficit” in spring, that determines the overall seasonal cycle of the upper geostrophic transport. Therefore, the AMOC seasonal cycle at 6°N is in fact mainly a consequence of the lagged adjustment of the NBC to the interior WSC forcing. This incomplete compensation between the boundary current and interior ocean due to oceanic adjustment is the dominating mechanism for the geostrophic part of the AMOC throughout the tropics.

3.4 Seasonal AMOC cycle at 34.5°S

34.5°S is selected to study the seasonal variability in the southern subtropical ocean. At this latitude the Ekman transport has a minimum in February and maximum in June with peak to peak amplitude of 6Sv, and the geostrophic transport has minimum in
January and maximum in June with peak to peak amplitude of 2.5 Sv (Fig. 3.1). The nearly out of phase Ekman and geostrophic transports, and larger amplitude in Ekman transport, leads to the dominance of Ekman transport in shaping the seasonal AMOC at 34.5°S.

As shown in Figure 3.13a, the seasonal anomaly of $\Psi_1$ in OFES shows an annual cycle of $\pm$1.5 Sv amplitude at the edge of the eastern boundary (blue line), which is twice the Rossby deformation radius away from the eastern boundary. Between 5°E and 15°E, larger fluctuations are associated with Agulhas rings passing northward across the section. These rings not only modify the local meridional flow but also generate substantial interannual variations, so that the derived annual cycle here is not robust. West of 0°E, the impact of the Agulhas ring is not obvious and the seasonal cycle of $\Psi_1$ presents a structure that is very similar to that at the edge of the eastern boundary zone, and this remains stable across most of the interior basin. Westward of 40°W this signal becomes significantly amplified by additional meridional flow in the western part of the basin. At the continental shelf where the water depth is $h_{moc}$ (green dash line), $\Psi_1$ features a minimum in September and maximum in March. Including the WBC (the Brazil Current, hereafter BC), however (red line in Fig. 3.13a), the annual cycle is fundamentally changed and leads to a cycle with minimum in February and maximum in June when integrated across the full section.

A similar seasonal cycle of the meridional geostrophic transport is simulated by the two-layer model (Fig. 3.14). Near the eastern boundary, the seasonal cycle in the two-layer model generally agrees with that in the OFES although its maximum occurs in February rather than December. The Agulhas rings near the eastern boundary are not
reproduced by the two-layer model, but they do not affect the seasonal cycle of the zonally integrated transport west of 0°E. What is captured by both OFES and the two-layer model is the relatively uniform seasonal variability in the interior ocean and the rapid shift across the BC to a phase that is almost opposite to that in the interior ocean (Fig. 3.13a and Fig. 3.14a). It is interesting to note also that the interior modulations of $\Psi_1$ by the annual baroclinic Rossby waves are much less pronounced in both OFES and the two-layer model at 34.5°S than at 26.5°N. This is due to different structure of the WSC across the basin. As shown in Figure 3.7d, the annual cycle of the WSC at 26.5°N has broad coherent pattern across the whole basin so that the Rossby waves generated at the eastern boundary are simply modulated by the annual forcing as they propagate westward and maintain their energy. However, the WSC at 34.5°S has a weak seasonal cycle and varying phase across the basin (Fig. 3.14d), such that the local forcing tends to damp the propagating Rossby waves and reduces their amplitudes in the central part of the basin. The same type of behavior occurs in the simple forced Rossby wave model for the South Atlantic (not shown).

A separation of the geostrophic current into barotropic ($\Psi_{bt}$) and baroclinic ($\Psi_{bc}$) components indicates that $\Psi_1$ is controlled by both barotropic and baroclinic processes. Near the eastern boundary, the seasonal fluctuations in $\Psi_{bc}$ are very similar to that in $\Psi_1$ (blue lines in Fig. 3.13a and Fig. 3.14a). This suggests that it is the uplift and depression of the thermocline that controls the eastern boundary changes, dynamics similar to that at 26.5°N. The vertical motions of the thermocline are forced by the wind stress curl near the eastern boundary (Fig. 3.14d). As shown in Figure 3.5, the interface anomaly along the eastern boundary is coherent between 25°S and 35°S, indicating that
the seasonal cycle shown here is regionally forced over about 10° meridional extent. Note that, the wind stress curl anomaly in SCOW has similar phase and amplitude to that in NCEP (not shown).

In the interior ocean, $\Psi_{bc}$ is affected by the westward propagating Rossby waves, analogous to that at 26.5°N, but as noted above with lesser amplitude. The contribution of the barotropic streamfunction $\Psi_{bt}$ to the upper layer streamfunction $\Psi_1$ is relatively greater at 34.5°S than at 26.5°N, so that $\Psi_1$ reflects attributes of both $\Psi_{bt}$ and $\Psi_{bc}$. The seasonally-varying barotropic circulation is more complicated than in the northern hemisphere and shows two gyre-scale patterns (Figs. 3.9c, d). South of 30°S, the anomalous ocean circulation is dominated by an anti-clockwise (clockwise) gyre in February (July), while north of 30°S a clockwise (anti-clockwise) gyre is evident in February (July). These two different barotropic gyre structures are generated by opposite seasonal cycles of the wind stress curl south and north of 35°S (not shown).

Similar to the results at 26.5°N, however, the net impact of the barotropic circulation on the seasonal AMOC variability is small. At the continental shelf (green dash line in Fig. 3.13 and Fig. 3.14), the barotropic circulation is largely compensated so that $\Psi_{bc}$ dominates the annual cycle in $\Psi_1$ when zonally integrated across the basin (black line in Fig. 3.13c, Fig. 3.14c). Especially, the rapid transition of the annual cycle at the edge of the continental shelf is mostly included in $\Psi_{bc}$, demonstrating that it is the shallow part of the BC in depths ≤1000 m that controls the seasonal AMOC cycle. As shown in Figure 3.5, the seasonal anomalies of the BC transport have a meridional scale of about 15°, or perhaps greater (Figs. 3.5a,c), and are probably associated with both local and remote coastal wind forcing that cause an uplift of the coastal thermocline in
austral winter and a depression of the coastal thermocline in austral summer. Part of the BC annual cycle is also modulated by the Rossby wave pattern that produces more localized, and opposite signed, thermocline depth anomalies just offshore of the western boundary layer. This process is different from 26.5°N because of the blocking effect of the Bahamian Islands.

We note that the barotropic circulation also contributes to the total seasonal changes of the BC. As shown in Figure 3.9c and Figure 3.9d, the intensified barotropic streamlines near the western boundary indicate that the southward BC becomes stronger in February (July) and weaker in July (February), for the regions south (north) of 30°S. This is consistent with the seasonal changes of the BC at 34.5°S in both OFES and the two-layer model (red line in Fig. 3.13a and Fig. 3.14a), implying that the seasonal spin-up (spin-down) of the barotropic circulation leads to acceleration (deceleration) of the BC as indicated above. However, the seasonality of the BC induced by the barotropic circulation is mostly compensated and does not significantly contribute to the geostrophic transport across the basin.

3.5 The basinside wide geostrophic transport

The geostrophic transport across the basin is ultimately determined by the pressure difference between the basin boundaries. In a simplified two-layer ocean, like the simple model used in this study, the seasonal changes in the zonally integrated baroclinic streamfunction, $\Psi_{bc}$, which dominates the seasonal AMOC variation at all latitudes, is related to the change in interface depth on the boundaries. The relative importance of each boundary, at any given latitude, can be evaluated by the ratio of their annual amplitude. Figure 3.15 displays the latitudinal distribution of this ratio estimated from the
two-layer model, together with the annual amplitudes of basinwide geostrophic transport. North of 10°N, the ratio is larger than one, indicating that the eastern boundary has a larger contribution to the seasonal AMOC cycle. In the equatorial region, however, the ratio is much less than one, pointing out the dominant role of the western boundary. In the southern subtropical region, both boundaries tend to have comparable contributions, reflected by ratios that range from 0.5 to 1.5. The meridional distance over which these coastal forcing regions are coherent are typically about 10-15° along either boundary, and these are roughly the same scales over which the AMOC seasonal variation is meridionally coherent.

The dynamical processes evaluated in this study point out the important factors to correctly simulate the seasonal cycle of MOC, and this might shed light on model-based MOC hindcast studies. As noted previously, the wind stress curl at the boundaries in NCEP appears to be unrealistically low, at least at 26.5°N. To assess the sensitivity of the meridional geostrophic transport to different wind forcing, we ran the two-layer model using the SCOW dataset, which is designed to better preserve wind stress patterns near the ocean boundaries. In the southern hemisphere the geostrophic AMOC amplitudes are generally similar to those forced by the NCEP dataset, and they are generally larger than the NCEP run in the northern hemisphere (Fig. 3.15). The ratio of eastern to western boundary contributions to the basinwide geostrophic MOC seasonal cycles is also somewhat different in the NCEP and SCOW runs. Therefore, deficiencies in the wind forcing near the basin boundaries, such as appears to occur in the NCEP reanalysis winds used to drive the OFES model, can lead to inaccuracies in the simulation of the magnitude of the AMOC seasonal cycle.
Figure 3.1. The seasonal anomaly field derived from the monthly OFES result (1980-2009) for (a) AMOC; (b) Geostrophic; and (c) Ekman components. Contour interval is 2 Sv. The 3°S-3°N equatorial band is excluded from (b) and (c).
Figure 3.2. Seasonal cycle (black solid lines) of (a) AMOC; (b) Gulf stream (GS); (c) Ekman (EK); and (d) upper mid-ocean (UMO), as obtained from monthly averages of the 26.5°N time series between April 2004 and April 2011. The gray envelopes represent the standard error of each month.
Figure 3.3. As in Figure 3.2, but for the climatological annual cycle derived from the OFES model (1980-2009). Dashed lines are calculated from the two-layer model simulation, where the annual mean from OFES is added to the two-layer result. (The Ekman contribution is the same for both models.)
Figure 3.4. Seasonal cycle of a) AMOC and b) Geostrophic transport from two-layer model. Unit: Sv.
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Figure 3.7. a), b) and c) are similar to Figure 3.6, but from the two-layer model at 26.5°N. The two-layer model has an exactly repeating annual cycle so there is no error for each month. d) wind stress curl calculated from the wind forcing used in the two-layer model along 26.5°N. Unit: $10^{-7} \text{N/m}^3$. 
Figure 3.8. (a) Upper layer local meridional transport and (b) upper layer cumulative streamfunction for the last five years of the two-layer model simulation at 26.5°N. The local meridional transport is calculated from the meridional geostrophic velocity weighted by $h_{moc}$ and grid distance ($dx$). The upper layer streamfunction is the zonal accumulation of the local meridional transport from the eastern boundary. Unit: Sv.
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Figure 3.10. a) Streamfunction for the baroclinic geostrophic flow simulated by the Rossby Wave model at 26.5°N is shown in the right. Unit: Sv. The seasonal cycle near the eastern (blue) and western (green) boundary are shown at the left, and their locations are marked by blue and green dash lines, respectively. The selected locations near the eastern and western boundary are the same as those in the two-layer model (see text). b) Seasonal cycle of the UMO from the two-layer model forced by NCEP winds (black dash line) and from the two-layer model forced by SCOW winds (blue dash line). The seasonality of eastern boundary from the Rossby wave model forced by NCEP winds and SCOW winds are shown in black solid line and blue solid line, respectively. The comparison between two-layer model and Rossby wave model results illustrates that the seasonal cycle in the two-layer model is mostly contributed by the eastern boundary.
Figure 3.11. As in Figure 3.6, but for (a) upper layer meridional geostrophic streamfunction ($\Psi_1$); and (b) baroclinic streamfunction ($\Psi_{bc}$) from the OFES model (1980-2009) at 6°N. The dash lines mark the locations for the offshore edge of the eastern boundary layer (blue), western point of the interior ocean (green), and the western basin boundary (black). The seasonal cycle at the three points in each streamfunction are also shown at left in their corresponding colors. Red line denotes the North Brazil Current.

Unit: Sv.
Figure 3.12. a) and b): as in Figure 3.11, but for the two-layer model. c) wind stress curl calculated from the wind forcing used in the two-layer model at 6°N. Unit: $10^{-7} \text{ N/m}^{-3}$. d) Seasonal anomaly of the North Brazil Current in OFES (thick solid) and two-layer model (red). Dash line is the boundary current transport derived from the stationary Sverdrup balance.
Figure 3.13. As in Figure 3.6, but for 34.5°S. The red line in a) is the seasonal cycle of the Brazil Current, taken as the difference of the streamfunction at basin boundary (black dash line) and western interior point (the continental shelf where water depth is $h_{moc}$, green dash line).
Figure 3.14. a), b) and c) similar to Figure 3.13, but for the two-layer model at 34.5°S. d) wind stress curl calculated from the wind forcing used in the two-layer model at 34.5°S. Unit: $10^{-7} \, N/m^3$. 
Figure 3.15. a) Latitudinal distribution of the annual amplitude of the basinwide
geostrophic AMOC from the two-layer model forced by NCEP (black) and SCOW(red);
b) Ratio of the annual amplitude of the interface anomaly at the eastern boundary to that
at the western boundary. Results from the two-layer model forced by NCEP and SCOW
are shown in black and red, respectively. The annual amplitude at each latitude represents
the difference between the maximum and minimum values of the annual cycle.
Chapter 4

Wind-forced AMOC interannual variability at 26.5°N

Although many mechanisms have been proposed to explain the AMOC variability in model simulations, it remains to be demonstrated whether these dynamics can explain the directly observed AMOC changes. The AMOC interannual fluctuations at 26.5°N were described by McCarthy et al. (2012) who reported that the AMOC significantly decreased in 2009/2010. They further pointed out that the decline was caused by both a temporarily reversed Ekman transport and strengthening of the southward mid-ocean geostrophic flow. However, it is still unclear what factors control these substantial interannual variations. In this chapter, we analyze the AMOC interannual variability observed by the 26.5°N array and use a simple two-layer model to understand the corresponding dynamics.

The two-layer model used in this section is described in chapter 2. For most of this section we will focus on the results of a linear version of this model, in which nonlinear advective terms in the momentum equations are neglected. This is done to better isolate the directly forced response of the AMOC, although a nonlinear version of the model is also run and compared to the linear model. The two-layer model is spun up from rest for 20 years using the 1988 daily CCMP winds as a repeated annual cycle, and then forced by daily CCMP winds between 1988 and 2011.
4.1 Interannual variability of the AMOC

The zonally-integrated meridional overturning streamfunction observed by the 26.5°N array is shown in Kanzow et al. (2010) and McCarthy et al. (2012). It has been split into three parts in the vertical: a net northward flow is located in the top 1100 m and 5000 m to bottom, respectively; southward flow lies between 1100 and 5000 m. The top 1100 m contains the warm branch of the AMOC ($T_{AMOC}$) transporting 17.6 Sv water toward the north. It is a combination of 31.6 Sv of northward Gulf Stream ($T_{GS}$) flow through the Straits of Florida shallower than 780 m, 3.2 Sv of northward Ekman transport ($T_{EK}$) in the top 100 m and 17.2 Sv of southward upper mid-ocean ($T_{UMO}$) transport in the top 1100 m. The UMO includes the northward Antarctic intermediate water (AAIW) between 800-1100 m, the northward Antilles current east of Abaco (Johns et al., 2008), and the southward gyre recirculation in the top 800 m. The southward flow between 1100 m and 5000 m is the North Atlantic Deep Water (NADW) which consists of 12.1 Sv of Upper NADW (UNADW) between 1100 m and 3000 m and 7.4 Sv of Lower NADW (LNADW) between 3000 m and 5000 m. The lowest part (5000-bottom) is the 2.0 Sv of northward Antarctic Bottom Water (AABW) transport.

The variability of the upper branch of the AMOC is split into its three components: $T_{AMOC}=T_{GS}+T_{EK}+T_{UMO}$. The time series of $T_{AMOC}$ and its individual components between April 2004 and April 2011 are shown in Figure 4.1. Both $T_{AMOC}$ and its components show fluctuations on time scales ranging from subseasonal to interannual. The standard deviation of $T_{AMOC}$, $T_{GS}$, $T_{EK}$ and $T_{UMO}$ are 4.9 Sv, 3.1 Sv, 3.5 Sv and 3.5 Sv, respectively, over the whole time series (Apr. 2004-Apr. 2011). They are very similar to
the respective values of 4.8 Sv, 2.9 Sv, 3.5 Sv and 3.2 Sv for the time period between 2004 and 2008 reported by Kanzow et al. (2010).

The interannual $T_{AMOC}$ anomalies and its three components are illustrated in Figure 4.2, where an 18 month (1.5 year) low-pass filter has been applied to the raw time series. Hereafter we will discuss exclusively these low-pass filtered fluctuations of $T_{AMOC}$ and its components, unless otherwise stated. The low-pass filtered $T_{AMOC}$ gradually decreases by about 2.0 Sv from Oct. 2005 to Oct. 2008, after which it drops abruptly to a minimum in Dec. 2009. Between Oct. 2008 and Dec. 2009, $T_{AMOC}$ declines by approximately 6.0 Sv. From early 2010, $T_{AMOC}$ begins to recover and returns to its average strength by Oct. 2010. Both the low pass filtered $T_{EK}$ and $T_{GS}$ remain relatively steady from Apr. 2004 to Oct. 2008, and then both are slightly intensified before a decline in 2009 where $T_{EK}$ decreases by 3.0 Sv and $T_{GS}$ decreases by 1.5 Sv. Their minima occur in the 2009-2010 winter. Fluctuations of the low-pass filtered $T_{UMO}$ between Apr. 2004 and Apr. 2008 are similar to those of $T_{AMOC}$. However, $T_{UMO}$ starts to decrease (become more southward) around April 2008, about half year earlier than $T_{AMOC}$. The minimum of $T_{UMO}$ anomaly occurs in Aug. 2009 which is also several months earlier than that of $T_{AMOC}$ and $T_{EK}$. The different amplitudes and timing of $T_{UMO}$, $T_{EK}$ and $T_{GS}$ fluctuations indicate that all three components contribute to the interannual anomaly of $T_{AMOC}$, but $T_{UMO}$ has the dominant contribution, especially during the 2009/2010 negative event in $T_{AMOC}$.

$T_{UMO}$ is geostrophic and, based on the thermal wind theory, its strength and variability are determined by the density gradient between the eastern and western boundaries. The contribution from each boundary can be estimated by recalculation of $T_{UMO}$ with the other boundary density profile fixed to its time mean value. Kanzow et al.
(2010) used this approach to point out the dominating contribution of the eastern boundary to the seasonal variations of $T_{AMOC}$. Here we apply the same concept to isolate the contributions of each boundary to the interannual variability of $T_{UMO}$. The time series of $T_{UMO}$ determined by the western (eastern) boundary density variability is called $T_{UMO-wb}$ ($T_{UMO-eb}$), whose low-pass filtered time series are shown in Figure 4.2b. It is obvious that the interannual variability of $T_{UMO}$ is dominated by $T_{UMO-wb}$, implying that the density variations at the western boundary are almost entirely responsible for the interannual anomaly of $T_{UMO}$. This is consistent with the finding by McCarthy et al. (2012) that $T_{UMO}$ is highly correlated with the changes in the thermocline depth at the western boundary.

To examine if the observed fluctuations are representative of the interannual variability at 26.5°N, we compare the 7 years of data from the 26.5°N array with the available measurements over a longer period. $T_{EK}$ between 1988 and 2011 is calculated from the CCMP wind product. As shown in Figure 4.3a, $T_{EK}$ has typical fluctuations between $+1$ Sv and $-1$ Sv, but has its largest negative anomaly in the 2009-2010 winter (Fig. 4.3a). This suggests that the dramatic change in 2009/2010 is unique in the past two decades. The long term time series of $T_{GS}$ has been monitored since 1982 by the undersea telephone cables which are maintained by the Atlantic Oceanographic and Meteorological Laboratory (AOML). To analyze the full period between 1982 and 2011, all gaps of 3-months duration or fewer are linearly interpolated. A 20-month gap between 1998 and 2000 was also filled with the climatological annual cycle derived from the whole period. The low-pass filtered time series (Fig. 4.3b) reveals that $T_{GS}$ has substantial fluctuations, especially during the 1990s. The interannual variability of $T_{GS}$ before 2004
are well documented in previous papers (e.g. Baringer and Larsen, 2001; DiNezio et al., 2009; Meinen et al., 2010). It is interesting to note that, after 2004, when the 26.5°N array was deployed, $T_{GS}$ has remained remarkably steady compared to its historical variability. Due to a lack of long-term measurements of $T_{UMO}$, the representativeness of its variability over longer time scales, as well as that of $T_{AMOC}$, is unclear. However, the significant changes in both $T_{EK}$ and $T_{GS}$ imply that there might have been substantial interannual variations in $T_{AMOC}$ prior to 2004.

4.2 Mechanisms of the interannual variability

In this section, we explore the physics accounting for the interannual variability of $T_{AMOC}$ and its components. The Ekman transport at 26.5°N is driven by the easterly zonal wind stress. As shown by DiNezio et al. (2009), a substantial part of the interannual wind stress variability along 26.5°N is related to the North Atlantic Oscillation (NAO). During high NAO phases, atmospheric pressure over the mid-latitude North Atlantic increases, tending to cause increased trade winds in the subtropics and an enhanced northerly Ekman transport. The correlation between the 18-month low-pass filtered $T_{EK}$ and NAO index from 1988 to 2011, $r=0.6$, which is significant at a 95% confidence level (Fig. 4.3a). In particular, the most substantial change of $T_{EK}$ captured by the 26.5°N array coincides with the negative phase of the NAO in the winter of 2009/2010. Similar weakened $T_{EK}$ occurs in 1996-1998 when the NAO shifts to its negative phase. It should be noted that $T_{EK}$ at 26.5°N does not always closely follow the NAO index. This is because the spatial pattern associated with individual NAO events is not necessarily identical (Hilmer and Jung, 2000; Cassou et al., 2004; Luo et al., 2010).
The GS in the Straits of Florida is predominantly a geostrophic current, so that the pressure gradient across the Strait should be closely related with $T_{GS}$. Lacking *in situ* measurement of pressure profiles across the current, sea level data from the tide gauges in the Bahamas and Miami are used as a proxy for the pressure data. The correlation between low-pass filtered $T_{GS}$ and the sea level difference between Bahamas and Miami is 0.73 (Fig. 4.3b), which is about the sum of the correlation with individual sea level changes (Miami, -0.45; Bahamas, 0.32). Thus, the sea level changes on both sides of the strait contribute to the fluctuations of $T_{GS}$, and they combine nearly linearly to produce the total correlation of the transport with the sea level difference. To verify this result, the regional correlation between $T_{GS}$ and the low-pass filtered Aviso altimetry sea level anomaly (SLA) is shown in Figure 4.4a. Similar to those found between $T_{GS}$ and tide gauge data, negative and positive correlations are located on the western and eastern sides of the Straits of Florida, respectively. In addition, the negative correlation extends well along the coast north of the Strait up to Cape Hatteras, as well as into the Gulf of Mexico, implying that the coastal sea level changes south of Cape Hatteras are broadly coherent in the alongshore direction. A regression of $T_{GS}$ onto interannual wind anomalies over the western part of the basin (Fig. 4.4b) shows a significant correlation (0.85) with alongshore wind stress along the southeast US coast. The most significant correlation occurs in the region between the Straits of Florida and Cape Hatteras, although the correlation pattern is broad and extends well to the east and northeast of the Straits. The relation of $T_{GS}$ to these local alongshore stress anomalies is as expected: increased northward wind stress causes enhanced offshore Ekman transport, leading to decreased coastal sea level and a corresponding geostrophic increase in $T_{GS}$. This response is
consistent with the negative correlation between $T_{GS}$ and SLA along the US east coast (Fig. 4.4a). In addition, the correlation between $T_{GS}$ and alongshore wind rapidly decays when $T_{GS}$ lags the wind stress for 3 months or longer. This implies that $T_{GS}$ responds nearly instantaneously to the interannual fluctuations of alongshore wind forcing, and that remote influences from winds north of the Straits are transmitted rapidly southward by topographic waves. It is worth noting that the generation of interannual $T_{GS}$ by these dynamics is consistent with many previous studies that have attributed seasonal and short-term variations of $T_{GS}$ to local wind stress forcing and upstream wave propagation (e.g. Anderson and Corry, 1985; Lee and Williams, 1988; Schott et al. 1988; Czeschel et al., 2012).

Another feature in Figure 4.4a is the positive correlation between the GS transport and the SLA at the Bahamas site. Since the Bahamas site is adjacent to the open ocean, the correlations found here reflect the effects of the interior ocean changes on the GS transport. To illustrate how the changes in SLA at the Bahamas are related to those in the interior ocean and to the GS transport, we show in Figure 4.5 the time series of the low-pass filtered SLA along 26.5°N. The interannual changes of the GS transport are obviously associated in part with the SLA at the Bahamas. In particular, the weakened GS in 1996 and 2000 coincides with the negative SLA at the Bahamas site; the intensified GS in 1997 and 2002 similarly corresponds to the positive SLA, although the SLA is slightly lagged. Some of the SLA anomalies can be traced back to the interior ocean, suggesting that westward propagating Rossby waves or long-lived eddies play a role in modulating the interannual SLA near the Bahamas. DiNezio et al. (2009) showed that about 50% of the GS transport variability at 3–12 yr periods can be explained by the
low-frequency wind stress curl (WSC) variations in the ocean interior, and that the lagged response of the GS to those fluctuations is consistent with westward propagation at observed first-mode Rossby wave speeds. They also pointed out that the largest correlation between low-passed GS transport and WSC is located in the western sub-basin between 40°W to 75°W. This is consistent with the more evident westward propagating signal west of 45°W in the Aviso SLA data. Thus, westward propagating Rossby waves in the interior appear to play an important role in modulating the interannual changes of the GS transport. It should be emphasized that the dynamical processes controlling the SLA at the Florida coast (Miami) and at the Bahamas both contribute to the interannual changes of GS, so that the SLA difference between Bahamas and Miami is most correlated with the GS transport. For example, the GS has substantial changes in the 1990s, a period when the sea level at both Miami and the Bahamas undergo opposite anomalies (Fig. 4.3b; Fig. 4.5). On the other hand, the decline of the GS during 2009 seems to be mostly associated with increased coastal sea level in Miami.

As shown in Figure 4.2b, the observed interannual changes of $T_{UMO}$ are mainly induced by the density changes at the western boundary. To reveal the vertical structure of the interannual changes of $T_{UMO}$, Empirical Orthogonal Functions (EOF) analysis is performed on the 18-month low-pass filtered mid-ocean transport profile. The first three modes explain 98% of the $T_{UMO}$ variance (Fig. 4.6). The first mode accounts for 70% of the total variance and has a first baroclinic mode structure in the vertical. The temporal component corresponding to the first mode has a variation very similar to that of $T_{UMO}$ (Fig. 4.6b). The second mode, which has the structure of second baroclinic mode, explains 20% of the variance. The third mode represents mainly the external transport
used to satisfy the zero mass transport constraint. It explains 8% of the variability and has a nearly barotropic structure in the vertical. On interannual time scales, the energy contained in the second and third modes are much less than in the first mode. Given the dominant contribution of western boundary density changes to the interannual fluctuations of $T_{UMO}$, it is hypothesized that the density profiles at the western edge are modified by the first baroclinic mode signal, and that this primarily results in the anomalous $T_{UMO}$. The origins of the baroclinic signal at the western edge could be local WSC forcing, remote effects from westward propagating Rossby waves or eddies, or meridionally communicated anomalies. To assess the effects of the local forcing, we analyzed the WSC near the western edge. The WSC is calculated using the daily CCMP winds and averaged within 1 degree in both latitude and longitude near the western boundary (26.5°N, 76.7°W). It is found that there is no close correspondence between the WSC and the observed interannual changes of $T_{UMO}$. Their correlation is 0.2, not significant at the 90% level. Therefore, the baroclinic signal at the western edge of the ocean interior is mainly affected by remote influences, either zonally or meridionally propagating anomalies. In the following we will focus on the zonally propagating waves because numerical studies suggest that they are the dominant source of the AMOC variations in the subtropics (Hirschi et al., 2007; Cabanes et al., 2008).

### 4.3 Verification of the mechanisms

To verify the proposed mechanisms in the previous section, the two-layer model described in section 2.2 is used to investigate the meridional transport anomalies along 26.5°N. While the AMOC can be affected by many factors, such as changes in deep
water formation associated with buoyancy forcing, here we focus explicitly on the role of
the surface wind stress forcing. The two-layer model does not include the buoyancy
forced overturning component, but includes the basic dynamics to simulate the barotropic
and first baroclinic mode oceanic adjustment to the surface wind forcing. The results of
the linear version of the model represent the changes in wind-driven circulation induced
by local wind forcing, coastally-trapped waves and ocean interior anomalies transmitted
to the western basin boundary by planetary waves, while the non-linear model also
includes intrinsic (non-forced) variability. In the following we first compare the results of
the two-layer model with observations, and then explore in more detail the underlying
dynamics of the AMOC variations in the model.

4.3.1 Comparisons between two-layer model and observations

The streamfunction at 26.5°N in the two-layer model is calculated by

\[ \Psi(x, z, t) = \int_x^{x_e} \int_0^z v(x, z, t) dz dx \]

where \( x_e \) is the eastern boundary and \( v(x, z, t) \) is the meridional
velocity along 26.5°N. The upper layer streamfunction is represented by the notation:

\[ \varphi_1(x, t) = \Psi(x, z, t)|_{z=h_1} \]

In a two-layer model context, the variability of \( \varphi_1 \) corresponds to the variation of
the upper branch of the AMOC. The thickness of the upper layer in the model, at 1000 m,
is also similar to the mean depth of the upper branch of the AMOC (~1100 m) in the
26.5°N array data.

The geostrophic part of \( \varphi_1(x, t) \) is calculated by

\[ \varphi_{1geos}(x, t) = \int_x^{x_e} \int_{h_1}^0 v_{1geos}(x, t) dz dx \]
where \( v_1^{geos}(x, t) \) is the upper layer geostrophic velocity, calculated from the sea surface height gradient.

To further understand the dynamics controlling the upper layer geostrophic transport, \( v_1^{geos}(x, t) \) is separated into “barotropic” \( (v_{bt}^{geos}(x, t)) \) and “baroclinic” \( (v_{bc}^{geos}(x, t)) \) components (Zhao and Johns, 2014):

\[
v_{bt}^{geos}(x, t) = \frac{1}{(H_1 + H_2)} (v_1^{geos}(x, t) * H_1 + v_2^{geos}(x, t) * H_2).
\]

\[
v_{bc}^{geos}(x, t) = v_1^{geos}(x, t) - v_{bt}^{geos}(x, t).
\]

Correspondingly, we can also split \( \varphi_1^{geos}(x, t) \) into barotropic \( (\varphi_{bt}^{geos}(x, t)) \) and baroclinic \( (\varphi_{bc}^{geos}(x, t)) \) parts.

The fluctuations of \( T_{AMOC} \) in the two-layer model are defined by the fluctuations in \( \varphi_1(x_w, t) \), where \( x_w \) is at the Florida coast. The \( T_{GS} \) in the two-layer model is defined as \( T_{GS} = \varphi_1(x_w, t) - \varphi_1(x_{bahamas}, t) \), where \( x_{bahamas} \) is the longitude corresponding to the Bahamas island. \( T_{UMO} \) is defined as \( \varphi_1^{geos}(x_{bahamas}, t) \).

Figure 4.7 shows the 18-month low-pass filtered \( T_{AMOC} \) anomaly and its components simulated by the linear version of the two-layer model, in comparison with the observational result for the 26.5°N array. \( T_{EK} \) in the two-layer model is exactly the same as that in the observations, because the same wind product is used in both.

The \( T_{GS} \) anomaly reproduced by the two-layer model has comparable interannual variability to that in the cable measurement, although the simulated \( T_{GS} \) anomaly has slightly weaker amplitude. In particular, the significant changes after 1996 and the behavior during 2004-2011, when the 26.5°N array was deployed, are successfully reproduced by the two-layer model. The major differences occur during the early 1990s when the model fails to reproduce some of the observed \( T_{GS} \) oscillations. The larger
differences at this time could be due in part to missing variability in the wind field prior to 1988 (since the two-layer model is spun up using 1988 daily winds), and also possibly to improved wind fields after the mid-1990's when the global scatterometry coverage increased. The correlation between model $T_{GS}$ and observations is 0.66, significant at the 95% level. As mentioned above, $T_{GS}$ is controlled by the pressure gradient across the Straits of Florida. Consistent with the AVISO SLA data, the correlation between $T_{GS}$ and SSH in the two-layer model shows negative and positive correlations located in the western and eastern sides of the Strait of Florida, respectively (Fig. 4.4c). This indicates that the important role of sea surface height (SSH) on both sides of the Florida Straits in modulating the interannual $T_{GS}$ is well captured by the two-layer model. The correlation pattern between the model $T_{GS}$ and the wind stress field (not shown) is essentially identical to that shown for the observations in Figure 4.4b. Thus, the model results support our conclusion that the coastal sea level variations along the southeast US coast, that are linked to the geostrophic $T_{GS}$ changes, are forced by regional winds in the vicinity of the Straits and areas to the northeast of the Straits.

Since the SSH contains both barotropic and baroclinic signals, the baroclinic effects can be evaluated by the correlation between the interface anomaly and $T_{GS}$ in the two-layer model. As shown in Figure 4.4d, the interface and $T_{GS}$ are positively (negatively) correlated along the US coast (offshore). This pattern is similar to and out of phase with that of the SSH, implying that $T_{GS}$ variability is mainly controlled by baroclinic processes. To better understand these baroclinic dynamics, we examine the interface anomaly and the corresponding geostrophic baroclinic velocity ($v_{bc}^{geos}$) in February 2009 when $T_{GS}$ has its largest positive anomaly during the 26.5°N array observation period. As
shown in Figure 4.8, the thermocline is uplifted along the Florida peninsula and slightly depressed near the Bahamas, driving a northward $T_{GS}$ anomaly. In addition, the shallower thermocline at the continental shelf can be seen to extend northward to about 40°N. The deeper thermocline near the Bahamas is associated with a large scale anomaly in the western basin of the Atlantic. These features are consistent with the general pattern of the correlations between thermocline and $T_{GS}$ (Fig. 4.4d), suggesting that the dynamics evaluated here are representative for the subtropical region.

$T_{UMO}$ simulated by the two-layer model shows similar variations to the in situ observations, which includes an increase in 2004/2005 and a downward trend starting in October 2005 (Fig. 4.7). The two-layer model suggests that the measured increase in 2004/2005 followed a decline in 2004. The $T_{UMO}$ decrease between 2005 and 2009 simulated by the model reaches 4.5Sv, accounting for 90% of the observed decrease in the 26.5°N array data. The minimum in the two-layer model occurs in March 2009, several months earlier than the in-situ observations, indicating that the two-layer model is not able to accurately capture the timing of the large $T_{UMO}$ drop in 2009. Nevertheless, the correlation of $T_{UMO}$ in the two-layer model and 26.5°N array is 0.67, significant at the 95% statistical level, suggesting that the two-layer model is generally successful in reproducing the observed interannual changes of $T_{UMO}$. The model also shows significant $T_{UMO}$ fluctuations before 2004. Similar to the downward trend between 2005 and 2009, obvious declines take place in 1996/1997 and 2001/2002. This implies that the remarkable drop in $T_{UMO}$ transport during 2009 observed by the 26.5°N array might not be unique.
Summing $T_{EK}$, $T_{GS}$ and $T_{UMO}$ components together, the $T_{AMOC}$ anomaly in the two-layer model has similar interannual variations to the in situ observations. Due to the error in timing of the modeled $T_{UMO}$ negative peak in 2009, which occurs earlier and does not last as long as the observed $T_{UMO}$ anomaly, the $T_{AMOC}$ minimum in 2009 is weaker in the two-layer model. Nevertheless, the temporal fluctuations of the simulated $T_{AMOC}$ agree well with that in the 26.5°N array. Their statistical correlation is 0.82, significant at the 95% level, demonstrating that the two-layer model does a fairly good job in reproducing $T_{AMOC}$ at 26.5°N.

### 4.3.2 Dynamics of the UMO transport

To further explore the dynamics accounting for the interannual $T_{UMO}$ in the two-layer model, the upper layer geostrophic streamfunction along 26.5°N ($\varphi_1^{geo}$) and its barotropic ($\varphi_{bt}^{geo}$) and baroclinic ($\varphi_{bc}^{geo}$) parts are shown in Figure 4.9, Figure 4.10 and Figure 4.11, respectively. In $\varphi_1^{geo}$ there are substantial fluctuations in the interior ocean which are linked to westward propagating planetary wave features (Fig. 4.9). In contrast, $\varphi_{bt}^{geo}$ has much weaker variations of approximately ±1 Sv amplitude across the interior, and its spatial pattern does not show clear westward propagation. In addition, $\varphi_{bt}^{geo}$ does not significantly contribute to the interannual variability of $\varphi_1^{geo}$ when integrated across the full interior basin (left, Fig. 4.10). This is because $\varphi_{bt}^{geo}$ mainly reflects the fast (time scales of weeks) spin-up and spin-down a large scale barotropic gyre circulation by barotropic Rossby waves forced by the interannual WSC anomalies in the interior (not shown). At 26.5°N this barotropic gyre circulation is contained over deep f/h contours (f
= planetary vorticity, h = ocean depth) within the deep ocean east of the Bahamas, so that \( \phi_{bt}^{geos} \) is compensated after basin-wide zonal integration.

\( \phi_{bc}^{geos} \), however, shares similar temporal and spatial structure with \( \phi_1^{geos} \), indicating that baroclinic processes play a dominant role in modulating the interannual changes of \( T_{UMO} \). This is consistent with the conclusion diagnosed from the in situ observations that the interannual variability of \( T_{UMO} \) is dominated by the first baroclinic mode.

\( \phi_{bc}^{geos} \) is determined by the interface anomaly (\( h_2 \)) field which is modulated by the wind induced baroclinic Rossby waves. For example, the zonal gradient of \( h_2 \) at 26.5°N in February 2009 (Fig. 4.8) drives southward anomalous flow east of 30°W, weak northward flow between 40°W and 50°W, and southward flow again west of 50°W. These anomalous flows lead to the corresponding spatial pattern in \( \phi_{bc}^{geos} \) (Fig. 4.11): a negative peak near 35°W, gradually weakened negative anomalies west of 40°W, and intensified negative anomalies west of 50°W extending to the Bahamas. This complicated spatial structure is associated with the oceanic adjustment to surface WSC forcing, in which the planetary waves play an essential role. Figure 4.12 shows the Hovmöller diagram for the 18-month low-pass filtered \( h_2 \) along 26.5°N, which reflects the westward propagating baroclinic planetary waves. The \( h_2 \) anomaly shown here is generally similar to and out of phase with the Aviso SLA, although the satellite data contain more variability on shorter spatial scales (Fig. 4.5), implying that the two-layer model can reproduce the baroclinic waves recorded in the SLA data. Since the model is only forced by the surface wind and is linear, so that no instabilities can occur, the baroclinic planetary waves in Figure 4.12 are generated solely by the WSC. In addition, these waves are modified by the local forcing on their way towards the western boundary, so that they
can become intensified, weakened, or even extinguished before approaching the Bahamas. Their amplitudes at any point are related to the integrated WSC anomaly along the wave characteristics. These propagating waves modulate the $h_2$ field and set up the baroclinic pressure gradient that drives anomalous meridional flow. It should be emphasized that although the interior ocean has substantial interannual changes, the zonally integrated meridional transport, i.e. $T_{UMO}$, is ultimately determined by the thermocline gradient between the basin boundaries. The relative contribution from each boundary (i.e., Bahamas or Africa) can be assessed by recalculating $T_{UMO}$ with the thermocline at the other boundary fixed to time-mean value, the same method applied to the 26.5°N array data (section 4.1). $T_{UMO}$ and the corresponding transport associated with the western boundary ($T_{UMO-wb}$) and eastern boundary ($T_{UMO-eb}$) are illustrated in Figure 4.12 (left panel). The interannual changes of $T_{UMO-wb}$ generally agree with $T_{UMO}$, especially for the large amplitude fluctuations. Their statistical correlation is 0.83, much larger than that between $T_{UMO-eb}$ and $T_{UMO}$ (0.2). This demonstrates that the interannual changes of $T_{UMO}$ are mostly affected by the western boundary thermocline changes, which is consistent with the dominant contribution of the western boundary (Bahamas) to $T_{UMO}$ variability in the 26.5°N array data. This highlights the importance of the density profile near the Bahamas in recovering $T_{UMO}$.

The density changes near the Bahamas are affected by both remote forcing carried by westward propagating signals and local forcing. The propagating waves impinge on the western boundary and induce boundary waves which communicate the energy equatorward, leading to decreased amplitude towards the western boundary (Kanzow et al., 2009). The weakened variability near the Bahamas can be seen in Figure 4.12,
especially for the waves reaching the western boundary in 2002/2003, 2007-2008, and 2010. For linear long waves incident on a western boundary, Kanzow et al. (2009) showed that the ratio of the total wave amplitude on the boundary to that of the incoming wave is proportional to $\frac{\beta \Delta y}{f}$, where $f$ is the Coriolis parameter, $\beta$ is the meridional gradient of $f$, and $\Delta y$ is the meridional scale of the incoming wave. At 26.5°N the long planetary waves in the two-layer model have typical $\Delta y$ of 1000km, so that the expected amplitude reduction is about a factor of 3. This estimate is consistent with the fraction of reduced variability in SSH data and dynamic height data from altimetry and the 26.5°N array moorings (Kanzow et al., 2009). Note that $\Delta y$ can be occasionally larger than 1000 km, in which case the amplitude of the incoming wave is, thus, not greatly decreased. This is illustrated by the incident wave carrying a depressed thermocline in 1997-1998. The $\Delta y$ associated with this wave is about 2000 km in the two-layer model, and the reduction factor is only about 1.5. As a result, this wave successfully reaches the western boundary and generates a large negative anomaly of $T_{umo}$. It should be noted that the above theoretical estimate of the reduction factor is only based on the dynamics associated with wave reflection and interaction at a vertical western boundary. In the two-layer model, the specific amplitude of the thermocline changes at the western boundary is also modified by other factors, like local WSC forcing and topography. Nevertheless, the basic dynamics governing the interannual changes of $T_{umo}$ involves these wind induced baroclinic planetary waves. As suggested by the two-layer model, the downward trend between 2005 and 2009 observed by the 26.5°N array is attributed to the baroclinic waves carrying uplifted and depressed thermocline signals around 2005 and 2009, respectively.
4.3.3 Nonlinear effects

Up to this point we have considered only the linear two-layer model, as a way to clearly identify the forced response of the AMOC to wind stress variability. It is useful to also consider what changes may occur in the simulated response if nonlinear effects are included. Therefore, we conducted three additional experiments (NL-A, NL-B and NL-C) with the nonlinear version of the two-layer model. NL-A has the same spin up process and wind forcing with the linear experiment. NL-B and NL-C do not have a spin-up phase, and are instead initialized by the model fields from the NL-A run on Dec 31st, 1994 and Dec 31st, 2001, respectively. NL-B and NL-C are designed to test the sensitivity to initials states. The inclusion of the non-linear term can have several significant effects; first, it allows for instabilities and associated mesoscale variability to develop (to the extent that these processes can be resolved by the 1/4° resolution and simple stratification of the model), and second, it introduces an advection term in the dispersion relationship of the planetary waves which can modify their propagation.

In Figure 4.13, the interannual anomalies of the simulated $T_{GS}$, $T_{UMO}$ and $T_{AMOC}$ in the nonlinear experiments are shown and compared against the linear run and the observations. In general, the nonlinear simulations all show similar behavior to the linear simulation, and all of the major events are common to both. The r.m.s. differences for the $T_{AMOC}$ relative to the linear simulation are 0.7 Sv, 1.0 Sv, and 1.1 Sv for NL-A, NL-B, and NL-C, respectively (Table 4.1). Although the maximum differences can reach up to 2.3 Sv, the mean differences in $T_{AMOC}$ over the period 1988-2012 are less than 1.1 Sv. Most of the differences in $T_{AMOC}$ are caused by differences in $T_{UMO}$. In particular, the two nonlinear experiments NL-B and NL-C, which are initialized with higher values of the
$T_{AMOC}$ and $T_{UMO}$ than the linear experiment, tend to retain these positive biases for a number of years before becoming more similar to the linear experiment. Also, the nonlinear experiments all exhibit lower values of the $T_{AMOC}$ and $T_{UMO}$ during 2002-2005, but with similar oscillations. However, there is no evidence of any phase shifting in the peaks and troughs of the $T_{UMO}$ or $T_{AMOC}$ that would indicate a systematic effect of advection on the arrival of planetary wave signals at the western boundary. The differences due to non-linear effects therefore appear to be mainly random or related to initialization effects rather than having a systematic nature.

It is interesting to note that the 2009/2010 $T_{AMOC}$ anomaly in each of the nonlinear experiments has larger amplitude and shows improvement in reproducing the observed 2009/2010 signal. This improvement is mostly contributed by $T_{UMO}$, where the negative anomaly lasts longer than in the linear experiment. However, $T_{UMO}$ in these experiments still does not exactly match the observations. To further explore the $T_{UMO}$ differences between the linear and nonlinear experiments, we show in Figure 4.14 the upper layer $\varphi_{1}^{g ios}$ of the NL-A model and the time series of $T_{UMO}$ from NL-A, the linear model, and observations. The detailed structure of $\varphi_{1}^{g ios}$ in the nonlinear model is somewhat different from that in the linear model (compare to Fig. 4.11). In particular, the structures of the positive interior $\varphi_{1}^{g ios}$ anomalies in 1993-1996 and 2001-2002 show some significant differences, with the former event being much stronger in the nonlinear experiment and the latter being weaker, presumably due to the effects of instabilities or wave-wave interactions on these forced wave signals. However, the main features of the two simulations are consistent. It is interesting to note in Figures 4.11 and 4.14 that the negative $\varphi_{1}^{g ios}$ anomaly that strikes the western boundary in 2009-2010 displays a much
broader temporal scale when away from the western boundary than it does once it arrives at the boundary. In the linear model, and less so in the nonlinear models, this signal penetrates fully to the western boundary only near the beginning of this event and produces the relatively intense but short-lived $T_{UMO}$ decrease in early 2009. We speculate that if this entire wave signal had been more effectively transmitted to the western boundary, the duration of this event would have been longer in the simulations, lasting until at least early 2010, more similar to that of the observed $T_{UMO}$ variability. Therefore it appears that the actual $T_{UMO}$ variability is quite sensitive to the details of the interaction of these interior waves with the western boundary, which can be different between events. The interaction between the planetary waves and the western boundary might also trigger some mesoscale processes which are certainly not correctly simulated by the linear model, nor probably by the nonlinear model at this resolution (0.25 degree). We do not undertake a further sensitivity study of these processes here, but this topic seems worthy of future study with more realistic models.
Table 4.1. Basic statistics representing the difference between nonlinear and linear experiments for $T_{AMOC}$ and its components. All statistics are derived from 18 month low-pass filtered time series. For nonlinear models NL-B and NL-C, the statistics exclude the first year of the simulations to reduce the initialization effects.

<table>
<thead>
<tr>
<th></th>
<th>RMS difference (Sv)</th>
<th>Max difference (Sv)</th>
<th>Mean difference (Sv)</th>
</tr>
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<tbody>
<tr>
<td>$T_{GS}$</td>
<td>NL-A</td>
<td>NL-B</td>
<td>NL-C</td>
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<tr>
<td></td>
<td>0.3</td>
<td>0.4</td>
<td>0.3</td>
</tr>
<tr>
<td>$T_{UMO}$</td>
<td>0.5</td>
<td>0.9</td>
<td>1.0</td>
</tr>
<tr>
<td>$T_{AMOC}$</td>
<td>0.7</td>
<td>1.0</td>
<td>1.1</td>
</tr>
</tbody>
</table>
Figure 4.1. The thin lines (colors) represent the 10-day low-pass filtered time series of $T_{AMOC}$ (red), $T_{GS}$ (blue), $T_{EK}$ (black) and $T_{UMO}$ (magenta) for the period between April 2004 and April 2011. Horizontal lines (colors) are the mean values for each component. The bold lines (black or yellow for $T_{EK}$) are the 90-day low-pass filtered time series of each transport component. Unit: Sv.
Figure 4.2. The 18-month Butterworth low-pass filtered time series of $T_{AMOC}$ (red), $T_{GS}$ (blue), $T_{EK}$ (black) and $T_{UMO}$ (magenta), $T_{UMO-eb}$ (green) and $T_{UMO-wb}$ (blue) for the period between April 2004 and April 2011. $T_{UMO-eb}$ and $T_{UMO-wb}$ are calculated by fixing the density profile at the western boundary and the eastern boundary, respectively, to their mean values. The temporal mean values are removed to obtain the interannual anomaly. Unit: Sv.
Figure 4.3. a) 18-month low pass filtered $T_{E K}$ (black) and NAO index (color bar) between 1988 and 2011. $T_{E K}$ is derived from the CCMP wind at 26.5°N. The NAO index is obtained from National Oceanic and Atmospheric Administration’s Climate Prediction Center (available online at http://www.cpc.noaa.gov/data/teledoc/nao.shtml).

b) 18-month low-pass filtered Florida current transport ($T_{G S}$, solid blue) and normalized sea level anomaly (SLA) derived from the tide gauge stations at Bahamas Settlement Point, Bahamas (black) and Virginia Key, Miami (green, 1996-2011) and their difference (red). The tide gauge data are obtained from University of Hawaii Sea Level Center (http://uhslc.soest.hawaii.edu/).
Figure 4.4. a) The correlation between 18month low-pass filtered cable measurements of Florida current transport ($T_{gs}$) and Aviso SLA. b) Regression of $T_{gs}$ onto low-pass filtered surface CCMP wind (vectors) and the correlations between $T_{gs}$ and surface CCMP winds (contour). The black line is the 95% confidential level. c) Correlation between $T_{gs}$ and SLA in the linear two-layer model. d) Correlation calculated from $T_{gs}$ and interface (thermocline) anomaly in the linear two-layer model.
Figure 4.5. Left: 18-month low-pass filtered $T_{gs}$ (blue) from the cable measurements and normalized AVISO SLA at Miami (green) and Bahamas (black). Right: Low-pass filtered Aviso SLA along 26.5°N. Unit: cm. The blue dash line marks the location of the Bahamas Island.
Figure 4.6. a) First three modes of the Empirical Orthogonal Function (EOF) calculation with the observed velocity profiles for $T_{UMO}$ at 26.5°N. Unit: Sv/m. The percentage of variance explained by each mode is labeled. b) Low-pass filtered time series of $T_{UMO}$ (magenta) and the corresponding contributions from each mode (black for mode 1, blue for mode 2, and red for mode 3).
Figure 4.7. Low-pass filtered time series of the anomalous $T_{AMOC}$ (red), $T_{GS}$ (blue), $T_{EK}$ (black) and $T_{UMO}$ (magenta). The dash lines are for the in situ observations between April 2004 and April 2011. The solid lines are the linear two-layer model results between 1988 and 2011. Since the two-layer model does not include the thermohaline part, their anomalies are compared. For better visualization, the mean level of $T_{GS}$, $T_{AMOC}$ and $T_{UMO}$ are shifted by 4Sv each. Unit: Sv.
Figure 4.8. The interface anomaly (color shading, unit: m) and corresponding baroclinic geostrophic velocity ($v^{geos}_{bc}$, black vectors, unit: cm/s) in February 2009 from linear two-layer model. Magenta dash line denotes the 26.5°N latitude.
Figure 4.9. Left: low-pass filtered time series of $T_{umo}$ anomaly in the linear two-layer model (solid) and \textit{in situ} observations (dash). Right: low-pass filtered $\varphi_1^{geos}$. The Bahamas is on the left edge of the plot. Unit: Sv.
Figure 4.10. Left: low-pass filtered time series of $T_{UMO}$ anomaly in the linear two-layer model (solid) and in situ observations (dash). The black line represents the meridional geostrophic transport in the upper layer due to the barotropic velocity ($v_{bt}^{geo}$). Right: low-pass filtered $\varphi_{bt}^{geo}$ in the upper layer. Unit: Sv.
Figure 4.11. Left: low-pass filtered time series of $T_{UMO}$ anomaly in the linear two-layer model (solid) and *in situ* observation (dash). The black line represents the meridional geostrophic transport in the upper layer due to the baroclinic velocity ($v_{bc}^{eos}$). Right: low-pass filtered $\phi_{bc}^{eos}$ in the upper layer. Unit: Sv.
Figure 4.12. Left: low-pass filtered time series of $T_{U_MO}$ anomaly (Magenta), $T_{U_MO-wb}$ (blue) and $T_{U_MO-eb}$ (green) in the linear two-layer model. Unit: Sv. Right: low-pass filtered interface anomaly along 26.5°N in the two-layer model. Unit: m.
Figure 4.13. Low-pass filtered time series of the anomalous $T_{AMOC}$ (red), $T_{GS}$ (blue), $T_{EK}$ (black) and $T_{UMO}$ (magenta) for the in situ observations (dash lines, between April 2004 and April 2011) and linear two-layer model results (thick solid lines, between 1988 and 2011). The thin solid black, green and light blue lines are results from the NL-A, NL-B, NL-C experiments, respectively. Unit: Sv.
Figure 4.14. Left: low-pass filtered time series of $T_{UMO}$ anomaly in linear two-layer model (magenta solid) and in situ observations (magenta dash). Result for nonlinear experiment (NL-A) is shown in black. Right: low-pass filtered $\varphi_{1}^{geos}$ simulated by the NL-A experiment. Unit: Sv.
Chapter 5

Basinwide response of the AMOC to interannual wind forcing

This chapter expands on chapter 4, which studied the fundamental dynamics of the observed AMOC interannual variability at 26.5°N, to explore the wind-driven AMOC interannual variability throughout the Atlantic basin. Both the OFES and two-layer model are used. In this chapter, the same winds that force the OFES model (i.e. the NCEP/NCAR reanalysis) are used to force the two-layer model. In this chapter, the two-layer model described in chapter 2 is linearized by removal of nonlinear advective terms in the momentum equations. This is done to examine whether the simplest dynamics can account for the AMOC variability. We also conduct another experiment using the full primitive equation two-layer model (chapter 2), its results will be discussed in chapter 6. Therefore, if not explicitly stated, all results in this chapter are from linear two-layer model. The model is spun up from rest for 20 years using the NCEP/NCAR 1950-2010 climatological winds as a repeated annual cycle, and then forced by monthly NCEP/NCAR winds between 1950 and 2010. There is no thermohaline forcing in the two-layer model, and therefore its time-varying AMOC is driven only by winds, and it is also missing the large scale thermohaline circulation present in the OFES model.
5.1 AMOC interannual variability in OFES

The overturning circulation simulated by OFES can be described by the meridional streamfunction: \( \Psi(z, y, t) = \int_{x_w}^{x_e} \int_0^{z} v(x, y, z, t) \, dz \, dx \) where \( x_w, x_e \) are the western and eastern boundaries, and \( v(x, y, z, t) \) is the meridional velocity. The time-mean structure of \( \Psi(z, y, t) \) in OFES between 1950 and 2010 is shown in Figure 2.1. The maximum values of \( \Psi(z, y, t) \) are defined as the AMOC strength, so that the AMOC refers to the basinwide northward meridional transport in the upper ocean. Since the meridional transport in the upper ocean includes both Ekman and geostrophic current, the AMOC anomalies are also separated into Ekman and geostrophic transport, respectively. The Ekman transport is derived from the zonal wind stress:

\[
\Psi_{ek}(y, t) = \int_{x_w}^{x_e} \frac{-1}{\rho f} \tau_x(x, y, t) \, dx
\]

where \( \tau_x, \rho, f, x_w \) and \( x_e \) are the zonal wind stress, reference density, Coriolis parameter, and western and eastern boundaries, respectively. The geostrophic contribution to the AMOC is estimated by removal of Ekman transport from the total meridional transport above \( h_{moc} \), where \( h_{moc} \) is the depth where the maximum \( \Psi \) occurs.

In the following, our focus is on the low-frequency variability of AMOC, so that a 2-year low-pass filter is applied to the monthly time series of AMOC. Low-frequency changes were also examined by removal of a climatological seasonal cycle created from the 61 year monthly time series; its results are essentially the same with those from the 2-year low-pass filtered data. We therefore only discuss the low-pass filtered results in this study. The low-pass filtered AMOC anomalies throughout the basin are shown in Figure 5.1 and the corresponding Ekman and geostrophic components are shown in Figure 5.2.
and 5.3, respectively. The AMOC anomalies have typical year-to-year variability of $O (2–4 \text{ Sv})$ in most latitudes and they are coherent between $15^\circ \text{S}$ and $30^\circ \text{N}$. Similarly, significant changes also occur in both the Ekman and the geostrophic transports, suggesting that both contribute to the AMOC low-frequency changes. These low-frequency variations apparently have different time scales which can be categorized as decadal-interdecadal and interannual variability. A 10-year low-pass filter is applied to the AMOC, Ekman and geostrophic transports to isolate the decadal-interdecadal variability and 2-10 year band filter is used to highlight the interannual variability.

The AMOC decadal-interdecadal anomalies between 1950 and 2010 exhibit a gradual upward trend south of $40^\circ \text{N}$ and decadal oscillations north of $45^\circ \text{N}$ (Fig. 5.1b). This behavior is consistent with the simulated AMOC decadal changes in the late 20th century using various numerical models, such as Biastoch et al., (2008) and Yeager and Danabasoglu (2014). The decadal-interdecadal changes in the Ekman transport have largest amplitude in the tropics but are much weaker in the subtropics and high-latitudes (Fig. 5.2b). In contrast, the geostrophic transport shares similar type of variability with the AMOC except in the tropics where the basinwide anomalies are disconnected in the geostrophic transport (Fig. 5.3b). In fact, the 10-year low-pass filtered Ekman and geostrophic transport are generally out of phase in the tropics (Fig. 5.2b, Fig. 5.3b), meaning that the decadal-interdecadal changes of the tropical geostrophic transport are compensational variability to the surface wind. This indicates that the AMOC decadal-interdecadal variability is contributed by both Ekman and geostrophic transports in the tropics and dominated by the geostrophic component outside of the tropics. In addition, the anomalies in the AMOC and geostrophic transport after 1960 tend to originate from
the north Atlantic subpolar region and slowly propagate southward. This is in line with other numerical studies that the AMOC decadal changes are mainly produced by high-latitude buoyancy flux forcing (Biastoch et al., 2008; Danabasoglu 2008). It is interesting to note that a weakening trend in the AMOC and geostrophic transport take place over the whole north Atlantic after the late 1990s, which is consistent with the weakening trend detected in the 26.5°N array (Smeed et al. 2014).

The AMOC interannual anomalies (Fig. 5.1c) are characterized by coherent structure across the tropical and subtropical regions, particularly between 5°S and 30°N, and they have comparable amplitude to the decadal changes. The coherence is quantified by the correlations between 26.5°N and different latitudes, and the magnitude of the interannual variability is measured by standard deviation. As shown in Figure 5.4, the significant correlation is distributed between equator and 35°N and the magnitude in this region does not change too much, suggesting that the AMOC variability at 26.5°N is generally representative for the northern tropics and subtropics. The AMOC variability on different time scales is further decomposed into Ekman and geostrophic transports (Fig.5.2c and Fig.5.3c). Because of the small Coriolis parameter in the low latitudes, the magnitude of the Ekman and geostrophic variations is much larger in the tropics. Outside of the tropics, their interannual variability are comparable, indicating that they both contribute to the AMOC interannual changes(Fig. 5.4).

The Ekman transport reflects the changes of surface zonal wind so that the westerly or easterly anomalies associated with different atmospheric modes in the Atlantic, such as the North Atlantic Oscillation (NAO) and Atlantic Meridional Mode (Hurrell et al. 2001; Chiang and Vimont 2004), are primarily responsible for modulating the Ekman transport.
Consequently, the meridional coherence of the Ekman transport is mostly determined by the meridional scale of the atmospheric modes. For example, the Ekman transport along 26.5°N is positively correlated with those between 10°N and 35°N and negatively correlated with those north of 40°N, generally consistent with the spatial pattern of the NAO (Fig. 5.4).

The geostrophic transport, however, depends on the density field so that it contains the signals induced by oceanic variability. The meridional coherence in the geostrophic transport generally follows the AMOC, but tend to have larger correlations north of 40°N, implying that the high latitudes have remote influence on 26.5°N. Also, there is cross-equatorial coherence in the AMOC, but it does not occur in either Ekman or geostrophic transport individually. This is probably because the Ekman and geostrophic transport have similar magnitudes and are compensational to each other in the tropics, and the AMOC is determined by their their superposition. The remainder of this study will focus on diagnosing the dynamics for the geostrophic transport with an aim to understand how the ocean variability leads to the AMOC interannual anomalies.

As mentioned above the AMOC includes the northward transport in the upper ocean, similarly, its geostrophic component is determined by all the geostrophic current in the upper ocean between the eastern and western basin boundaries. To identify the critical regions dominating the geostrophic transport, the zonally cumulative geostrophic streamfunction above $h_{moc}$ is computed: $\varphi_g(x,y,t) = \int_{x_e}^{x_e} \int_{h_{moc}}^{0} v_g(x,y,z,t)dzdx$ where $x_e$ is the eastern boundary. The geostrophic meridional velocity ($v_g$) is derived by removing the Ekman velocity from the total meridional velocity. The correlation between $\varphi_g(x,y,t)$ and basinwide geostrophic transport on the interannual time scale (2-10 years)
is shown in Figure 5.5. The correlation is 1.0 at the western basin boundary because the value of \( \varphi_g(x, y, t) \) at the western boundary (i.e. \( \varphi_g(x, y, t)|_{x=x_w} \)) is the basinwide geostrophic transport. In the tropics the cumulative transport (\( \varphi_g \)) starts to be significantly correlated with the basinwide geostrophic transport in the interior ocean, demonstrating that the upper geostrophic current over the broad open ocean considerably contributes to the zonally integrated transport. For the regions outside of the tropics, the correlations between \( \varphi_g \) and basinwide geostrophic transport are generally low over much of the basin but rapidly reach high values near the western boundary. This clearly indicates that most variance in the geostrophic transport is contributed by anomalous flows close to the western boundary. Since the geostrophic transport is largely determined by the pressure gradient between the basin boundaries, the significant correlations near the western boundary also means that the pressure anomalies near the western boundary are mostly responsible for the geostrophic interannual variability. The western boundary pressure changes can be remotely influenced by two dynamical processes: zonally propagating waves and eddies impinging onto the western boundary and meridionally propagating boundary waves or topographic waves. Both mechanisms are assessed in the following.

The section along 26.5°N is selected to comprehensively understand the dynamics for the AMOC interannual changes, with an aim to shed light on the interpretation of the observed signal by the 26.5°N array. Figure 5.6 illustrates the 2-10 year band filtered zonally geostrophic cumulative streamfunction above \( h_{moc} \) (i.e. \( \varphi_g \)) along 26.5°N. The most apparent feature is the alternating positive and negative anomalies in the broad interior ocean, demonstrating anomalous northward and southward integral transport
starting from the open ocean. Importantly, many of the anomaly belts are tilted westward, indicating westward propagation. This feature results from the fact that the zonally propagating baroclinic Rossby waves modify the pressure profiles in the interior, which in turn determine the zonally integrated transport (\( \phi_g \)). The amplitudes of \( \phi_g \) anomalies are amplified near the western boundary, especially west of 70°W, but are rapidly decreased from about 75°W towards the Bahamas boundary (77°W). This behavior is induced by the interaction between Rossby waves or eddies and the Bahamas boundary. The incoming waves or eddies impinging onto the western boundary lead to wave reflection at the boundary. The superposition of incoming and reflected signals tend to reduce the amplitude of the pressure anomaly and hence the cumulative streamfunction anomaly (\( \phi_g \)) at the Bahamas boundary. This has been clearly demonstrated in observations, as well as in theoretical/modeling context, by Kanzow et al. (2009) and Zhao and Johns (2014b). Westward of the Bahamas boundary, \( \phi_g \) is further modified cross the Straits of the Florida by the anomalous Gulf Stream (GS) transport, leading to the interannual fluctuations of the basinwide geostrophic transport shown in the left panel of Figure 5.6. Note that, the interannual variability of the GS transport is also modulated by the baroclinic Rossby waves generated in the interior ocean (DiNezio et al. 2009; Zhao and Johns 2014a). This is also confirmed by the fact that some of the anomalies in the Straits of Florida can be clearly traced back towards the interior ocean. It is worthwhile to mention that not all of the westward proagating features in the open ocean have a significant impact on the zonally integrated meridional transport (i.e. basinwide geostrophic transport), due to the fact that the signals carried by the Rossby waves can be
attenuated by additional forcing along the wave paths and their interactions with
topography.

Meridionally propagating boundary or topographic waves can also modulate the
basinwide pressure gradient and hence the geostrophic transport. To highlight these
processes the lead (lag) correlation between the geostrophic transport at 26.5°N and other
latitudes is computed (Fig. 5.7a). The geostrophic transports between northern tropics and
45°N are significantly correlated with the 26.5°N at zero lead (lag), meaning that the
geostrophic transport across the subtropics are broadly coherent. This coherence can also
be seen in Figure 5.3c. On the other hand, the significant correlation of the geostrophic
transport between 26.5°N and north of 45°N occurs when the higher latitudes lead the
26.5°N by 6-10 months. This time scale is slower than the time needed for internal
boundary Kelvin wave propagation (1-2 months) (Kawase 1987; Johnson and Marshall
2002), but much faster than the time associated with the advection by mean background
flow (several years) (Gerdes and Köberle, 1995; Getzlaff et al. 2005; Zhang 2010).
Döscher et al. (1994) found that a delayed subtropical response to subpolar buoyancy
changes occurred after several months to 1 year, a time scale that agrees well with the 6-10
months detected here. They attributed the delay behavior to remote signals near the
western boundary carried by coastally trapped or topographic waves propagating towards
the equator and interacting with topographic features and stratification. Therefore, the
time lags observed in OFES suggest that meridionally propagating coastally trapped or
topographic waves connect the subpolar and the subtropical region and play a vital role in
modulating the interannual changes at 26.5°N. It should be noted that when we removed
the climatological seasonal cycle from the original time series and recalculated the lead-
lag correlation, significant correlations with high latitudes leading the 26.5°N by 6-10 months were also found.

5.2 Dynamics of the AMOC interannual variability

5.2.1 The AMOC at 26.5°N

In the above analysis two dynamical processes modulating the AMOC interannual variability in OFES are proposed and evaluated. In this section, a simple two-layer model described in chapter 2 is configured to verify the proposed mechanisms. As shown in the diagnostic studies by Biastoch et al., (2008) and Yeager and Danabasoglu (2014), the AMOC interannual changes are mostly generated by the momentum flux, therefore the two-layer model forced by the surface wind stress forcing is capable to capture the AMOC interannual variability. Lacking thermohaline forcing, the two-layer model does not include the buoyancy forced overturning component but it has the basic dynamics to simulate oceanic adjustment to the wind forcing. In a two-layer model context, the AMOC is calculated from the meridional velocity in the upper layer and the geostrophic transport is computed from the upper layer geostrophic velocity. The initial mean thickness of the upper layer in the model is about 1000 m, which is similar to the mean depth of the upper branch of the AMOC (h_moc) in OFES. Essentially, the AMOC anomalies in the two-layer model represent to the upper layer anomalous meridional transport induced by the oceanic adjustment to the surface wind forcing.

The low-pass filtered AMOC and geostrophic anomalies simulated by the two-layer model are illustrated in Figure 5.8 and Figure 5.9. Similar to the OFES results, both the AMOC and geostrophic transport have substantial low-frequency fluctuations throughout
the Atlantic basin (Fig. 5.8a and Fig. 5.9a). The anomalies occur on different time scales and many of them are characterized by meridionally coherent structure across the tropics and subtropics. The 10-year low-pass filtered anomalies suggest that the two-layer model does simulate a certain amount of decadal-interdecadal variability for the AMOC, but its pattern is different from the OFES model (Fig. 5.8b and Fig. 5.1b). A detailed comparison of the geostrophic component between the two-layer model and OFES shows that their decadal-interdecadal variability share similarity in the tropics but differ substantially outside of the tropics. This indicates that the decadal-interdecadal changes of the tropical geostrophic transport s in the OFES are mostly generated by the surface wind which is the only forcing in the two-layer model. On the other hand, the interannual changes in the two-layer model (Fig. 5.8c and Fig. 5.9c) show similar pattern with those in OFES (Fig. 5.1c and Fig. 5.3c), especially in terms of the latitudinal coherence and the southward propagation of the subpolar signals. As discussed in section 5.2.2, the AMOC and geostrophic transport in both models are significantly correlated and have comparable magnitudes in most latitudes. This indicates the primary role of surface wind in modulating the AMOC interannual changes.

Based on the above comparisons, it is reasonable to assume that the anomalies simulated by the two-layer model provide an estimate of the wind-driven response in OFES. In addition, Biastoch et al., (2008) and Yeager and Danabasoglu (2014) concluded that the AMOC variability is almost a linear superposition of buoyancy-driven and wind-driven changes. Thus, we try to infer the buoyancy forced signals in OFES by removal of two-layer model results, which are taken as an approximation of the wind-driven variability in OFES. Figure 5.10 shows the residual geostrophic decadal-interdecadal
variability in OFES after removing the 10-year low-pass filtered two-layer model geostrophic results. The remaining signals in OFES show a coherent structure throughout the basin. This pattern is quite similar to the purely buoyancy forced anomalies shown in Biastoch et al. (2008) and Yeager and Danabasoglu (2014), further demonstrating the importance of buoyancy flux in driving the AMOC decadal-interdecadal variability.

As discussed in section 5.1, the westward propagating Rossby wave is critical to modulate the interannual variability of the geostrophic transport and AMOC in OFES. This dynamical process is also included in the two-layer model. Figure 5.11 displays the upper layer geostrophic accumulated streamfunction \( \varphi_{1g} \), which is computed from the meridional geostrophic velocity in the upper layer of the two-layer model. Similar to the OFES results, numerous anomalies in \( \varphi_{1g} \) can be found over a large fraction of the ocean, reflecting that the meridional transport between the eastern boundary and the open ocean undergoes dramatic year-to-year changes. Note, the OFES model pattern (Fig.5.6) can reach larger amplitudes and is generally noisier than the two-layer model, probably because the OFES model is an eddy resolving three-dimension model and therefore contains random mesoscale features or instability that can significantly modify the local pressure profiles and hence the values of \( \varphi_{1g} \) at some longitudes. It should be emphasized, however, that both the OFES and two-layer model are characterized by the westward titled features, implying that they are modulated by the westward propagation of the Rossby waves.

To check whether there is meridional propagating signal in the two-layer model, the lead (lag) correlation of the interannual changes for the geostrophic component between different latitudes and 26.5°N is exhibited in Figure 5.7b, the interannual fluctuations
along 26.5°N is significantly correlated with those between north equator and 40°N. The anomalies in the subpolar North Atlantic are leading the 26.5°N by 6-12 months. It should be emphasized that the overall pattern of the lead (lag) correlation in the two-layer model is remarkably similarly to that in the OFES (Fig.5.7a), although the 2-layer model generally shows higher maximum correlations. If the leading signal from the subpolar region in OFES is caused by the coastally trapped or topographic waves, the two-layer model results suggest that these waves can be mostly captured by the first baroclinic mode dynamics.

To further demonstrate the remote forcing from the high-latitude on the AMOC at 26.5°N, a sensitivity experiment is designed with the two-layer model. The control run refers to the simulation forced by 1950-2010 monthly NCEP/NCAR reanalysis data whose results are discussed above. The sensitivity experiment is named "subpolar run", in which the surface wind forcing north of 45°N is the same as the control run; the wind forcing south of 45°N, however, is from a repeated annual cycle. In other words, there is no interannually varying wind forcing south of 45°N. The interannual anomalies found south of 45°N, if any, represent the remote influences from north of 45°N. Note that the interannual variability in the control run during the last 10 years of the spin-up period, when the wind forcing is a repeated annual cycle, is found to be negligible. Therefore, the intrinsic variability induced by the repeated annual cycle is ignored in the subpolar run. As shown in Figure 5.12b, substantial interannual anomalies appear south of 45°N in the subpolar run and they are apparently linked to the anomalies in the high latitudes, indicating that wind-forced signals in the subpolar region can meridionally propagate into the subtropics. To identify the pathway for the remote-forcing signals, several snapshots
of the thermocline anomalies are illustrated in Figure 5.13. In January 1958, thermocline depth anomalies with magnitude of 40m are generated in the subpolar region. Meanwhile, the negative thermocline anomalies around 45°N start to move towards lower latitudes along the continental shelf. By May 1960 the negative thermocline depth anomalies have arrived 26.5°N and significantly modify the AMOC (Fig.5.12b). Two months later, these anomalies enter the Gulf of Mexico and further approach the equator (Fig. 5.13c). Once they reach the equator, they are communicated into the eastern boundary by Kelvin waves and continue to move along the eastern boundary into Northern and Southern Hemisphere (Fig.5.13d), consistent with the pathways found by Johnson and Marshall (2002). Between 45°N and 26.5°N, the signals mostly move along the continental shelf, indicating that they are communicated by coastally-trapped or topographic waves. The time scale for these waves to travel from 45°N to 26.5°N is about 4 months, consistent with the time lag in their lead-lag correlations (Fig.5.7) and yielding a propagation speed about 0.3-0.5m/s. This speed is one order smaller than the first baroclinic mode Kelvin wave (2-5m/s) in this region.

As a result of the propagation of these thermocline depth anomalies, the AMOC has negative anomalies at this time (Fig. 5.12b), suggesting that this oceanic process substantially contributes to the AMOC variability. To quantitatively evaluate the AMOC anomalies induced by the remote forcing in the high latitudes, the AMOC variance in the subpolar run are computed and compared with control run (Fig.5.12). North of 45°N the wind forcing is identical in both experiments so that their AMOC shares the same variability. South of 45°N the AMOC amplitudes in the subpolar run gradually decay with latitude, suggesting that the propagating signals are dissipated along the way. For
the latitudes between 15°N and 45°N (i.e. the subtropics), the subpolar run generally accounts for about half of the AMOC anomalies in the control run. In addition, the correlation of the AMOC anomalies in both experiments is above 0.5, significant at the 95% level, in this area. For the tropics and southern hemisphere, the subpolar forcing has much weaker influence. Consequently, the most significant response to the high-latitude forcing occurs in the subtropics.

5.2.2 Basinwide AMOC variability

The mechanisms evaluated in the 26.5°N section are also applicable for other latitudes. Figure 5.14 compares the AMOC anomalies from OFES model and the two-layer model (control run) at 40°N, 26.5°N and 6°N. The AMOC interannual changes in both models have similar amplitudes and in-phase year-to-year variations. Their correlations at the three latitudes are 0.74 (40°N), 0.77 (26.5°N) and 0.83 (6°N), respectively, all are significant at the 95% confidential level. This further confirms that the essential dynamics included in the simple two-layer model are responsible for the AMOC interannual changes in the OFES model.

For the entire Atlantic basin, the similar AMOC interannual variability in both OFES and the two-layer model is evaluated in terms of correlation and standard deviation (Root mean square, RMS). As illustrated in Figure 5.15, the correlation of the AMOC interannual anomalies between the two models is generally larger than 0.4 over the whole basin, all significant at the 95% level. The RMS of the AMOC interannual fluctuations in OFES ranges from 0.6Sv to 1.2Sv. The largest variability is located near 35°N due to nonlinear variability and meandering of the GS. The RMS in the two-layer model
generally follows the latitudinal distribution of the OFES, but with slightly weaker amplitude and much smoother shape. The two-layer model does not have the RMS peak near 35°N, probably because the linear model is not able to fully capture the variability of the GS. It should be noted that both OFES and the two-layer model are forced by the same momentum flux, which means that they have the same wind-driven Ekman transport. This will, to certain extent, contribute to the similar evolution of the AMOC in both models. To further quantify the indirect effect of the wind forcing, the Ekman transport is removed from the AMOC to isolate the geostrophic transport contribution. The correlation of the geostrophic transport has a different pattern from the AMOC (Fig.5.15). The significant correlation is mostly confined within the tropics and the subtropics. The amplitude for the geostrophic interannual variance is, however, still comparable between the two models. This behavior can be attributed to the fact the two-layer model uses a spatially uniform reduced gravity parameter so that the propagation speed of the baroclinic Rossby wave is not accurately simulated in the higher latitudes or that the linear two-layer model is not able to well simulate the advective effect of the Gulf Stream (North Atlantic Current). As shown in Figure 5.6 and Figure 5.11, the timing of positive (negative) peaks in both OFES and the two-layer is determined by the westward propagating Rossby wave. If wave speed is different in the two models, their geostrophic transport as well as the AMOC time series tend to be mismatch but still keep comparable amplitude. Nevertheless, the similar geostrophic transport over tropics and subtropics in both models prove that the dynamical processes evaluated in the previous section are responsible for the AMOC interannual changes.
Figure 5.1. The AMOC low-frequency changes in OFES are illustrated by 2-year low-pass filtered anomalies (a). The decadal-interdecadal variability is shown by 10-year low-pass filtered AMOC (b). 2-10 year band filtered data exhibits the interannual variability (c). Unit: Sv.
Figure 5.2. Similar to Figure 5.1, but for the basinwide Ekman transport throughout the Atlantic basin. Unit: Sv.
Figure 5.3. Similar to Figure 5.1, but for the basinwide geostrophic transport throughout the Atlantic basin. Unit: Sv.
Figure 5.4 Upper panel: The correlation between 26.5N and other latitudes for the AMOC (black), basinwide geostrophic transport (blue) and Ekman transport (red) in OFES after 2-10 year band-pass filter.
Lower panel: The interannual variability is measured by the standard deviation for the AMOC (black), basinwide geostrophic transport (blue) and Ekman transport (red) in OFES after 2-10 year band-pass filter. Unit: Sv.
Figure 5.5. The correlation between the basinwide geostrophic transport and the zonally accumulated geostrophic streamfunction above $h_{\text{moc}}$ in OFES $\varphi_g(x,y,t)$ and on the interannual time scale (2-10 years). The equatorial band (3°S-3°N) is masked.
Figure 5.6. Left: The basinwide geostrophic interannual variability at 26.5°N. Right: The 2-10 years band filtered zonally cumulative geostrophic streamfunction over $h_{moc}$ (i.e. $\varphi_g$) along 26.5°N. Unit: Sv.
Figure 5.7. The lead (lag) correlation between the geostrophic transport at 26.5°N and other latitudes in OFES (a) and two-layer model (b). The black lines denote the 95% t-test significance level.
Figure 5.8. a): two-year low-pass filtered AMOC anomalies simulated by the two-layer model. Their decadal-interdecadal and interannual variability are shown in b) and c), respectively. Unit: Sv.
Figure 5.9, Similar to Figure 5.7, but for the geostrophic transport in the two-layer model. Unit: Sv.
Figure 5.10. The 10-year low-pass filtered geostrophic transport in the OFES with the geostrophic transport simulated by the two-layer model removed. Unit: Sv.
Figure 5.11. Left: The 2-10 year band filtered basinwide geostrophic transport from the OFES (black) and two-layer model (red) at 26.5°N. The geostrophic transport in the two-layer model is computed from the geostrophic meridional velocity in the upper layer. There is no mean thermohaline component in the two-layer model, so only the anomalies are shown here.
Right: the band filtered upper layer zonally accumulated geostrophic streamfunction along 26.5°N in the two-layer model. Unit: Sv.
Figure 5.12. a), standard deviation of the AMOC anomalies in the control run (red) and subpolar run (black). b), the AMOC interannual anomalies over the whole basin from the subpolar run. c), correlation of the AMOC interannual anomalies between the control run and subpolar run.
Figure 5.13. The thermocline depth anomalies in the subpolar run in January 1985 (a), May 1985 (b), July 1985 (c) and January 1986 (d). Negative values mean deeper thermocline. The magenta line denotes the 45°N where south of that no interannual wind forcing is applied.
Figure 5.14. The AMOC interannual variability along 40°N (a), 26.5°N (b) and 6°N (c) from OFES(black) and two-layer model(red). Unit: Sv. Note that no thermohaline component is included in the two-layer model, so that only the anomalies are compared here.
Figure 5.15. Upper panel: The correlation of AMOC (black) and basinwide geostrophic transport (blue) between OFES and the control run in two-layer model. Lower panel: The AMOC interannual variability is measured by the standard deviation in two-layer model (red) and OFES (black). The basinwide geostrophic transport is shown in blue (OFES) and magneta (two-layer model). Unit: Sv.
Chapter 6

Summary and Discussion

Understanding the AMOC variability is essential to assess the global climate state and predict future climate variability. Despite that the long-term climate change is mostly linked to the AMOC variability induced by buoyancy forcing, there is evidence that the wind-forced AMOC interannual variability also leads to some important climatic consequences. For instance, the heat storage changes associated with the 2009-2010 AMOC anomaly may have caused a feedback on the NAO and winter European weather (Taws et al. 2011; Cunnningham et al. 2013 and Bryden et al. 2014). Therefore, it is important to understand how the wind-driven variability contributes to AMOC changes, and what the time scales and processes are, so that longer term AMOC variability related to, e.g., buoyancy forcing, can be better detected. In this work, the wind-driven AMOC variability on different time scales is quantified in both in-situ observations and numerical models. In the following we summarize the main results of this dissertation, which are contained in Chapters 3, 4, and 5.

6.1 The AMOC seasonal cycle

In chapter 3, the AMOC seasonal cycle over the Atlantic bains is derived from an eddy-resolving ocean model simulation. The seasonal variability is largest in the northern tropical Atlantic region where the peak to peak amplitude is 10 Sv with maximum in boreal winter and minimum in boreal autumn. The variations in subtropical regions are smaller with peak to peak amplitudes of 4-6 Sv and varying phases. By splitting the
AMOC variability into Ekman and geostrophic transport components, the contribution of each to the total AMOC variability can be examined. It is found that both contributions are characterized by substantial variations throughout the year. Their seasonal cycles are almost out of phase in the tropical Atlantic region, and the Ekman transport has larger amplitude, so that the seasonality of the AMOC is dominated by the Ekman component. In subtropical regions, however, their magnitudes are comparable and their phases vary with location so that both of them determine the seasonal cycle of the AMOC. The mechanisms governing the geostrophic component depend on how the barotropic and baroclinic circulation modes develop across the basin on seasonal time scales.

To explore the dynamics responsible for the seasonal cycle of AMOC and its components, a linear two-layer model forced by climatological NCEP winds is utilized. The similarity of the AMOC seasonal cycle between the two-layer model and OFES, which is forced both by wind and buoyancy fluxes, indicates that the variability of the AMOC on seasonal time scales is mainly associated with the wind forcing, and that the two-layer model includes the fundamental dynamics to reproduce this variability. It is shown that the depth dependent geostrophic transport is modulated by the oceanic adjustment to wind forcing, in which coastal boundary currents, baroclinic Rossby waves in the ocean interior, and large-scale barotropic flow adjustment play a vital role.

In order to identify the dominating mechanism in different regions, three latitudes, 26.5°N, 6°N, and 34.5°S, are selected to represent the northern subtropical, equatorial and southern subtropical ocean. At each latitude, the zonally accumulated streamfunction of the meridional geostrophic flow above $h_{moc}$ indicates that the geostrophic seasonal cycle is shaped either by the WBC or determined by both the WBC and interior transports.
Based on thermal wind theory, the interior flow in the deep ocean is determined by the density difference across the basin. The factors affecting the interior transport are the anomalies carried by westward propagating Rossby waves toward the western boundary and the fluctuations forced by the local wind stress curl. These physics are incorporated into a linear Rossby wave model which successfully accounts for the seasonal cycle of the upper mid-ocean transport at 26.5°N.

One important result in chapter 3 is the role of the western boundary current (WBC) in determining the basinwide geostrophic transport and AMOC seasonal cycle. In the tropical region, the WBC is controlled by the time dependent Sverdrup balance established by tropical baroclinic Rossby waves. The lagged response of the WBC to the interior wind forcing causes a seasonal imbalance of the net upper layer flow across the basin, which determines the annual cycle of the AMOC. On the other hand, in the extratropics the seasonal cycle of the WBC is not dominantly controlled by the planetary waves; instead it is determined mostly by local and remote forcing communicated along the western boundary. Local and remote wind forcing along the eastern boundary can also contribute significantly to the seasonal AMOC cycle, which, for example, is found to be the dominant mechanism at 26.5°N.

The overall results of this chapter point to the dominant role of eastern and western boundary currents in the extratropics in determining the basin-wide geostrophic AMOC seasonal cycle. While both baroclinic planetary waves and barotropic gyres set up by the interior WSC forcing strongly affect the total upper layer geostrophic circulation at any location, they do not contribute significantly to the seasonal AMOC cycle because of internal compensation when integrated over the full width of the basin. We find further
that the eastern boundary contribution to the AMOC cycle is generally dominant in the northern hemisphere, while this is not true in the southern hemisphere. The dominance of the eastern boundary contribution that has been demonstrated by the 26.5°N array is therefore not a result that holds over the whole basin. The seasonal AMOC variability in the tropics is fundamentally different, in that the interior upper-ocean geostrophic flow driven by the basin-wide Sverdrup forcing is an integral part of the response. Its lagged compensation by the WBC, set by the relatively fast adjustment time scale of tropical baroclinic Rossby waves, determines the basin-wide geostrophic AMOC seasonal cycle in the tropics. Here the eastern boundary plays a much more limited role, and it is primarily the WBC response to the interior wind forcing, rather than local or regional forcing effects near the boundaries, that dominates the geostrophic contribution to the AMOC seasonal cycle.

6.2 Observed AMOC interannual variability at 26.5°N

Chapter 4 analyzes the interannual variability of the AMOC at 26.5°N using in situ observations. $T_{AMOC}$ is found to have a downward trend starting in 2005 and reaches a minimum in 2009/2010. A separation of the $T_{AMOC}$ into $T_{GS}$, $T_{EK}$ and $T_{UMO}$ components suggests that its significant change is dominated by $T_{UMO}$ and also partly contributed by $T_{GS}$ and $T_{EK}$. In addition, the dynamics for each component are diagnosed from the in situ observations. In particular, the Ekman transport across the section is related to the NAO. Especially, the dramatic negative anomaly in 2009/2010 winter is caused by westerly wind anomalies associated with the NAO negative phase. $T_{GS}$ is controlled by the pressure (sea level) gradient across the Straits of Florida. The coherent SLA along the
Florida peninsula, and the entire southeastern US coast to the Gulf of Mexico, suggests that these fluctuations are controlled by coastal forcing processes including propagation of coastally-trapped waves. The SLA near the Bahamas is linked to the fluctuations in the interior ocean. The interannual variation of $T_{UMO}$ has a first baroclinic mode structure and is determined by the thermocline changes near the Bahamas (western side of the interior ocean).

It is found that $T_{AMOC}$ and its components simulated in a linear wind-driven two-layer model agree very well with the available in situ observations. Especially, the downward trend between 2005 and 2009 in both $T_{AMOC}$ and $T_{UMO}$ is mostly recovered, and is attributed to baroclinic planetary waves carrying uplifted and depressed thermocline signals around 2005 and 2009, respectively. Since the two-layer model is only forced by the winds, the good agreement between numerical results and observations demonstrates that a large part of the observed interannual changes at 26.5°N is induced by the surface wind forcing. The two-layer model also suggests that $T_{AMOC}$ had substantial changes prior to 2004 and that different components of $T_{AMOC}$ dominated these fluctuations at different times. Particularly, $T_{GS}$ played a bigger role in the past than during the 26.5°N array observational period (2004-2011). We also note that there are other simulated events in $T_{UMO}$ that are comparable to the observed 2009/2010 decline, such as the rapid $T_{UMO}$ decrease between 1995 and 1997. The Ekman anomaly in the 2009/2010 winter, however, appears to be unique to the past two decades. In addition, the latitudinal distribution of the $T_{AMOC}$ anomaly simulated by the two-layer model (not shown) indicates that these interannual anomalies are generally coherent between 10°N and 35°N, suggesting that the interannual variations recorded by the 26.5°N array is
representative for the subtropics in the North Atlantic. Finally, it is worth noting that the
dynamics shown in this section to be responsible for the interannual fluctuations of $T_{UMO}$
and $T_{AMOC}$ are different from the seasonal cycle which is generated largely by wind-
driven upwelling near the eastern boundary (chapter 3).

The significant influence of the surface wind forcing on $T_{AMOC}$ identified in the simple
two-layer model is consistent with results derived from more complex OGCM
simulations. Böning et al. (2006) and Biastoch et al. (2008) showed that the interannual
fluctuations of $T_{AMOC}$ in the subtropical region are dominated by wind driven anomalies
rather than buoyancy forced changes. In their experiments to isolate the AMOC signal
associated with buoyancy forcing, the amplitude associated with the deep water
formation in higher latitudes is only $O(1–2 \text{ Sv})$ at $40^\circ\text{N}$, and even weaker at $26.5^\circ\text{N}$(about 1 Sv). However, the wind-driven variations are more than twice that amplitude.
Here, our two-layer model also suggests that the interannual fluctuations of $T_{AMOC}$ are
about 3–5 Sv, much stronger than the estimated buoyancy forced changes in the OGCM
simulations.

Another mechanism that could explain some of the $T_{AMOC}$ interannual changes is
intrinsic ocean variability. Hirschi et al. (2012) estimated the $T_{AMOC}$ interannual changes
in two eddy-permitting ($0.25^\circ$) OGCM runs with the same forcing but different initial
conditions, and Thomas and Zhai (2013) investigated the $T_{AMOC}$ anomalies with an eddy-
resolving ($0.1^\circ$) OGCM forced by climatological forcing. In these studies the low-
frequency variability of $T_{AMOC}$ at $26.5^\circ\text{N}$ due to internal processes has r.m.s values of
about 0.3 Sv and 1.0-1.5 Sv, respectively. Larger variations of up to 3-4 Sv can
occasionally occur; for example, Thomas and Zhai (2013) found an AMOC reduction
with comparable amplitude to the 2009/2010 event in the 26.5°N array. While it is possible that intrinsic variability could trigger such extreme events as 2009/2010, the results of our study point to a dominantly wind-forced response as the explanation for that event. The r.m.s differences in the \( T_{AMOC} \) low-frequency variability between our linear and nonlinear simulations range from 0.7 to 1.1 Sv, more comparable to the eddy-resolving study of Thomas and Zhai (2013) than the eddy-permitting experiments of Hirschi et al. (2012). Nevertheless, it could be anticipated that the models with eddy-permitting resolution, such as ours and Hirschi et al. (2012), tend to underestimate the intrinsic variability, so that their results more reflect the forced response. Comparisons of the forced versus intrinsic \( T_{AMOC} \) variability on interannual time scales in realistically forced, high resolution (eddy-resolving) models, including also higher vertical resolution than used here, would be valuable to better establish the range of internal variability in relation to the forced response.

6.3 Basinwide AMOC responses to interannual wind forcing

The dynamics of the basinwide AMOC responses to interannual wind forcing are investigated in Chapter 5. Detailed analysis of the AMOC low-frequency changes in OFES model suggest that both the coastal/topographic waves and interior baroclinic Rossby waves are important factors modulating the interannual variability in the OFES. The signals carries by these two processes reach the western basin boundary and modify the pressure profile there and hence the basinwide geostrophic transport. A simple two-layer model forced only by the surface wind is used to reproduce the AMOC interannual variability in OFES. Both the remote forcing from the interior ocean and high-latitudes
are evaluated within the framework of two-layer model. The wind-forced anomalies in
the interior ocean propagate westward by means of Rossby wave and uplift (depress) the
thermocline depth at the basin boundary. On the other hand, the anomalies generated in
the high latitudes are communicated into the lower latitudes along the continental shelf
through coastally trapped or topographic waves. The superposed signals from both
pathways determine how the subtropical AMOC responds to the surface wind forcing.
With all the above dynamics incorporated, the two-layer model is able to basically
capture the AMOC interannual changes in the OFES.

It is worth noting that the pathways for the dynamic response to the higher latitudes
found in this paper are fundamentally different from previous studies which argued that
the AMOC meirdional propagation is either through boundary Kelvin wave or mean
advection along the NADW interior pathway (Kawase 1987; Johnson and Marshall 2002;
Getzlaff et al. 2005; Zhang 2010). There is no deep water formation in the two-layer
model, so that the mean abyssal circulation in the interior ocean, which is required to
advect the anomalies, does not exist. In the two-layer model and OFES, the AMOC in the
subtropics lags the subpolar by several months to 1 year, a time scale mainly established
by coastally trapped or topographic waves. The boundary Kelvin wave, however, takes
one or two months to reach the equator (Johnson and Marshall 2002), much faster than
the topographic wave. In addition, although the OFES model uses Arakawa B grid, eddy-
resolving resolution (1/10°), and the two-layer model use Arakawa C grid, 0.25°
resolution, the time scales for the subtropical lagged response in both models are
consistent, suggesting that the mechanisms found here are robust across different model
configurations.
Another point to be emphasized is that all the comparisons in chapter 5 and the above discussions in this section are based on the linear two-layer model. We also performed another experiment using the full primitive equation (nonlinear) model as described in chapter 2. The spin-up and forcing fields of the nonlinear model are exactly the same as the linear model. The differences of the 2-year low-pass, 10-year low pass and 2-10 year band pass filtered anomalies between the linear and nonlinear model are shown in Figure 6.1. It is clear that their differences are very small compared to the magnitudes of the AMOC variability in the linear model, suggesting that the effects of the nonlinear term in the two-layer model can largely be ignored for the time scales longer than 2 years. This further supports our conclusion that the simple dynamics in the two-layer model can mostly explain the AMOC interannual variability in the eddy-resolving OFES.

Recently, substantial assimilation and hindcast modeling efforts are focused on reaching a consensus on the mean state and variability of the AMOC over the past few decades throughout the Atlantic basin. However, the mean states of the AMOC are different among the models owing to different subgrid scale parameterizations and parameter choices as well as to differences in vertical and horizontal grid resolutions in the ocean models (Danabasoglu et al. 2014). In this study, the AMOC interannual changes reproduced by the simple two-layer model are similar to those in OFES, indicating that the AMOC interannual anomalies can be mostly understood by barotropic and first mode wind-driven dynamics. In fact, the dynamics in the simple two-layer model are definitely also included in the ocean general circulation model, it is therefore possible to conclude the AMOC interannual changes across different numerical models.
6.4 Sensitivity to wind forcing

One of the main goals in this study is to quantify the roles of surface wind forcing in driving the AMOC variability. Throughout the study, the two-layer model is forced by two types of wind forcing. The CCMP wind product is used in chapter 4 and produces similar AMOC interannual variability to that found in the *in-situ* observations. Chapter 5 utilizes the NCEP dataset and shows a favorable comparison with the AMOC low-frequency variability in OFES. The two sets of experiments enable us to test the sensitivity of the AMOC variability to different wind products. Figure 6.2 displays the 2-year low-pass filter, 10-year low-pass filter and 2-10 year band filter AMOC anomalies from the two-layer model forced by CCMP winds. Note that the CCMP stops at 2011 and the ECMWF-intrim wind is used from 2012 to 2014. The AMOC has energetic low-frequency changes throughout the basin (Fig. 6.2a). Several minima with comparable magnitude with the 2009-2010 event also occur before 2004. Interestingly, there is another minimum near the 2012-2013 winter occupying in the tropics and northern subtropics. The 10-year low pass filtered anomalies suggest that the CCMP winds produce decadal variability in the AMOC (Fig. 6.2b), but it exhibits a different pattern from that driven by the NCEP winds (Fig. 5.8b). For example, there is positive anomaly between 1995 and 2003 south of equator in the NCEP run, but the CCMP run produces a negative anomaly. The decadal trend after 2004 also seems different in the NCEP and CCMP runs. This is perhaps induced by different decadal trends in various wind datasets. The AMOC interannual changes driven by the CCMP winds is characterized by coherent variability across different latitudes (Fig. 6.2c), a feature also occurs in the NCEP run (Fig. 5.8c). Detailed comparison indicates that the magnitude of individual anomalies in both
runs is different, probably because the monthly mean of NCEP winds is used but the CCMP uses daily fields. More comprehensive comparisons, including the thermocline depth anomalies, between the two experiments are necessary to identify the fundamental differences caused by the two wind products. Nevertheless, the different behavior of the AMOC variability, especially the decadal changes, driven by different wind datasets suggest that the sensitivity to wind forcing should be carefully considered when drawing conclusions about AMOC longer-term trends.
Figure 6.1. The difference between linear and nonlinear two-layer model experiments in terms of AMOC anomalies after 2-year low-pass filter (a), 10-year low-pass filter (b) and 2-10 year band filter (c). Unit: Sv.
Figure 6.2. a): 2-year low-pass filtered AMOC anomalies simulated by the two-layer model forced by CCMP. Their decadal (10-year low-pass filtered) and interannual variability (2-10 year band filtered) are shown in b) and c), respectively. Unit: Sv.
Chapter 7

Future Work

This study mostly focuses on the upper warm branch of the AMOC, i.e the meridional transport above the top 1000m. It’s deeper and cold counterpart is called the North Atlantic Deep Water (NADW). The NADW is traditionally considered to be confined within the deep western boundary current (DWBC). In situ observations from the RAPID-MOCHA-WBTS observing system at 26.5°N find, however, that the DWBC carries much more volume transport than the NADW and that there is a substantial northward recirculation offshore of the narrow DWBC. As a result, it is difficult to simply use the DWBC as an index to assess the southward NADW flux and its variability. In order to attribute the variability of the DWBC and NADW transport and explore their possible connection, the abyssal ocean circulation structure should be taken into account. Our following work will try to utilizes OFES model results to assess the abyssal volume flux anomalies in different regions across the basin, including the DWBC, volume transport in the western sub-basin, and across the entire basin. In addition, the wind-driven two-layer model will be used to investigate dynamics that may control the variability in the abyssal volume transport.
Bibliography


Cheng, W., John C. H. Chiang, and Dongxiao Zhang, 2013: Atlantic Meridional Overturning Circulation (AMOC) in CMIP5 models: RCP and historical Simulations. J. Climate, 26, 7187–7197. doi: http://dx.doi.org/10.1175/JCLI-D-12-00496.1


