2016-11-08

Eddy-driven Seasonal Variability of the Florida Current Transport

Ricardo M. Domingues
ricardormd@gmail.com

Follow this and additional works at: https://scholarlyrepository.miami.edu/oa_theses

Recommended Citation
https://scholarlyrepository.miami.edu/oa_theses/630

This Embargoed is brought to you for free and open access by the Electronic Theses and Dissertations at Scholarly Repository. It has been accepted for inclusion in Open Access Theses by an authorized administrator of Scholarly Repository. For more information, please contact repository.library@miami.edu.
UNIVERSITY OF MIAMI

EDDY-DRIVEN SEASONAL VARIABILITY OF THE FLORIDA CURRENT TRANSPORT

By
Ricardo Marques Domingues
A THESIS

Submitted to the Faculty
of the University of Miami
in partial fulfillment of the requirements for
the degree of Master of Science

Coral Gables, Florida, USA
December 2016
UNIVERSITY OF MIAMI

A thesis submitted in partial fulfillment of
the requirements for the degree of
Master of Science

EDDY-DRIVEN SEASONAL VARIABILITY OF THE FLORIDA
CURRENT TRANSPORT

Ricardo Marques Domingues

Approved:

William Johns, Ph.D.
Professor, Department of Ocean
Sciences

Mohamed Iskandarani, Ph.D.
Associate Professor, Department of
Ocean Sciences

Christopher Meinen, Ph.D.
Physical Oceanographer
NOAA Atlantic Oceanographic
and Meteorological Laboratory
Miami, Florida

Guillermo Prado, Ph.D.
Dean of Graduate School
In this study, the role of westward propagating signals in driving year-to-year changes in the seasonal variability of the Florida Current (FC) transport is investigated based on controlled realistic numerical simulations carried out using the Regional Ocean Modeling System (ROMS). Different sets of idealized numerical experiments are performed with and without background flows associated with the FC to assess the different mechanisms involved, and include experiments initialized: (1) with single eddies of different sizes and intensities in the ocean interior without the background flow (SENS-E0x); (2) with single eddies at different latitudinal locations with the background flow; and (3) with eddy-full configurations in the ocean interior both with and without background flows.

The main finding from this study is that westward propagating signals can cause transient seasonal variability in the FC transport by means of both direct and indirect forcing mechanisms, in which the forced response is characterized by seasonal variability associated with variable annual phase. In the direct forcing mechanism, westward propagating signals cause a two stage response in the Florida Straits, in which the first stage is characterized by the development of barotropic velocity anomalies linked with the eddy-induced Rossby wave field, and the second stage is linked with the development of baroclinic wall-jets that propagate through Northwest Providence Channel. In the indirect forcing mechanism, westward propagating signals originating in the open ocean perturb the eddy
field offshore of the Gulf Stream in a manner analogous to the butterfly effect, which can drive the Gulf Stream variability to evolve into different state. The perturbed Gulf Stream variability is then linked to the Florida Straits through baroclinic coastaly trapped signals that travel along the east U.S. coast. Results indicate that the in the real ocean, the indirect forcing mechanism may play a dominant role in linking the open ocean variability from westward propagating signals to the changes in the FC transport, while the direct response mechanism may play a secondary role. The FC response driven by westward propagating signals simulated by this study had an amplitude of 2 Sv and is essentially linked with transient seasonal variability. Results reported here based on numerical simulations confirm findings from previous studies based on observations, which reported that year-to-year changes in the FC seasonality are largely linked with elevated levels of background transient variability due to westward propagating signals originating in the open ocean. In addition, the mechanisms reported here can be potentially linked to year-to-year changes in the seasonality of the Meridional Overturning Circulation (MOC) and Meridional Heat Transport (MHT), given that the FC corresponds to an important component of the MOC and MHT.
“I dedicate this research to my loving wife, Luana Gonçalves Domingues”

Ricardo Marques Domingues
Acknowledgments

In the course of my life and career, I have had the luck to meet people that always helped me keep moving forward to achieve bigger and bigger goals in my life. This was also the case during these past few years, as I received the help and support from a large number of people, without whom I would not be able to obtain this Master of Science degree.

First, I would like to thank my advisor, Dr. William Johns, for having me as part of your team, and for always being accessible, easygoing, and mindful with me. I am very grateful for your support and guidance during these past few years and during the development of this research. I have great admiration for you, and it has been a great honor and a pleasure to be your student. I would also like to thank my committee members Dr. Christopher Meinen, and Dr. Mohamed Iskandarani for also taking a critical part on this very important step of my career. I am also very grateful for your guidance during these past year and I feel very honored to have you as part of my committee.

I would like to thank my colleagues at the NOAA Atlantic Oceanographic and Meteorological Laboratory, from whom I am constantly learning and assimilating new knowledge. It is a great honor to have the opportunity to work among the most remarkable scientists and engineers in the field. I am specially thankful to Gustavo Goni and Molly Baringer, who have always supported my pursue for an advanced degree, and for all the great opportunities and recognition that I receive at NOAA/AOML. I am also thankful to Matthieu Le Henaff and George Halliwell for providing the North Atlantic HYCOM fields used to initialize some of the simulations developed in this study.

I would also like to thank my colleagues at RSMAS for all these great years, for the inspiring conversations at the wetlab, and for the overall support with courses and with the research. I am especially thankful for Gustavo Mastrorocco Marques for helping me setting up ROMS for my application.
Finally, and more importantly, I am extremely thankful to my loving wife Luana Gonçalves Domingues for always being my life partner, for taking a critical part in making my dreams come true, for always being there for me when I needed, for taking care of me, and for giving me great support during this part few years as I focused on getting a higher educational degree. I also thank the support of my parents Ana Maria de Lima Marques, Sergio Luis Rodrigues Domingues, and Ana Rita Domingues, which always had a very important role in helping me define my path in life.

I acknowledge support from the University of Miami, from the Cooperative Institute of Marine and Atmospheric Studies, and from the NOAA Atlantic Oceanographic and Meteorological Laboratory. The Florida Current cable data used in this study are made freely available on the Atlantic Oceanographic and Meteorological Laboratory web page www.aoml.noaa.gov/phod/floridacurrent/. The altimetry products were produced by Ssalto/Duacs, distributed by AVISO, and supported by the CNES (available at http://www.aviso.oceanobs.com).
Contents

List of Figures .......................................................... viii
List of Tables ........................................................... xix

Chapter One - Background and Scientific Methods ........................................ 1
  1.1 Introduction ........................................................... 1
  1.2 Scientific Background .............................................. 4
    1.2.1 Variability of the Florida Current transport .............. 4
    1.2.2 Westward propagating signals in the North Atlantic .... 10
    1.2.3 Eddy-related variability at North Atlantic’s western boundary . 16
    1.2.4 Eddy-wall mechanisms ..................................... 20
  1.3 Goal and Research Questions .................................... 23
  1.4 Methods .............................................................. 24
    1.4.1 Regional Ocean Modeling System - ROMS ................. 24
    1.4.2 Numerical experiments ..................................... 25
    1.4.3 The feature model approach ................................ 29

Chapter Two - Case Studies of Eddy-Western Boundary Interaction .................. 35
  2.1 No Background Flow Case Study .................................. 35
    2.1.1 Experiment setup ............................................ 36
    2.1.2 Results ....................................................... 38
Chapter Three - On the Impact of Realistic Westward Propagating Signals Based on Satellite-Altimetry Data

3.1 Experiment setup ................................. 103
3.2 Results ........................................... 105
3.3 Discussion ....................................... 120

Chapter Four - Conclusion ....................... 126

References ......................................... 136
List of Figures

1.1 Location of the Florida straits. Highlighted are the locations of the telephone cables used to estimate the Florida Current volume transport, and the location of relevant landmarks in the region of study, such as the Northwest Providence Channel (NWPC). Thin black lines show bathymetric contours every 200 m for depths between 100 m and 1000 m, while thin gray lines show bathymetric contours every 20 m for depths shallower than 100 m.

1.2 (a) Florida Current transport time-series (raw, light gray line) derived from measurements of voltage differences across the straits of Florida using telephone cables. Overlaid is the climatological annual cycle for the Florida Current during 1983-2014 (red line) and the transient component (blue line). (b) Wavelet transform for the Florida Current transport. The thick white lines highlight peaks in the spectral power that are significant at the 95% confidence level, while the black line marks the cone of influence for the analysis. The magenta lines delimits the 73-525 days frequency band. (c) Seasonal variability of the Florida Current transport during 1983-2014.
1.3 (a) Time-series of the transient component of the FC transport (black line, left axis), of the AMOC (red line, left axis), and of MHT (blue line, right axis). Linear regression analysis between the transient component of the FC transport and the (b) AMOC, and the (c) MHT.

1.4 Dispersion relationship for first mode baroclinic Rossby waves setting Rd equal to 47 km at 27°N, and using equal zonal and meridional wavenumbers (k=l, blue line), and also setting the meridional wavenumber to zero (l=0, red line).

1.5 SHA Root Mean Square along 27°N based on: raw (blue line), annual cycle (red line), high-frequency component (< 73 days, gray line), low-frequency component (> 525 days, green line), 73-525 days frequency band (black line).

1.6 SHA field for March 1st, 2000, filtered at the 73-525 days frequency band. Overlaid is the SHA-derived surface geostrophic velocity vectors.

1.7 (a) SHAr Hovmoller (longitude-time) diagram along 27°N. (b) Normalized power spectrum density from SHAs along 27°N for each longitude. (c) Wavelengths associated with SHAr along 27°N estimated using a fast-fourier transform. (d) Histogram distribution of SHAr at 27°N west of 70°W. Red lines in (d) indicate values of the 2.5th and 97.5th percentiles from the SHAr distribution, which have values of ∼15 cm and ∼15 cm, respectively.

1.8 (a) Internal radius of deformation based on climatological conditions the Word Ocean Atlas 2013 calculated using Equations. (b) westward propagating phases speeds for long, linear, first-baroclinic Rossby waves.
1.9  (a) Model domain used for numerical experiments developed in this study using ROMS. The grid has an horizontal resolution of approximately $1/25$ degree with 540 X 390 points. (b) Grid configurations shown in detail at the Florida Straits. The magenta lines show the location of sections across (c) the Florida Straits, and (d) the Northwest Providence Channel, showing the configuration of vertical layers at these locations.

1.10  (a) Temperature, (b) salinity, and (c) density profiles used as background conditions for model initialization, and its derived (d) buoyancy frequency. Baroclinic modes linked with the (e) vertical velocity and density structure, and with the (f) horizontal velocity and pressure field. Modes are solved using a rigid-lid boundary condition. In panel (e), the modified first baroclinic mode for vertical velocity ($\phi_m(z)$) is shown as the thick black line.

1.11  Gaussian function used to apply vertical displacements on isopycnal surfaces, where $D_E$ is the radius of the eddy, and $\Delta Z_{max}$ is the maximum vertical displacement allowed.

1.12  (a) Temperature, (b) salinity, and (c) density structure of an anticyclonic feature used as initial conditions for experiment Prel-E01.

1.13  Example of vertical displacements of isothermal surfaces using SHA data as the horizontal function in equation 1.5d, instead of the original Gaussian function (Figure 1.11).

1.14  (a) Surface velocity field, (b) surface relative vorticity field, and (c) vertical velocity structure of the eddy after full adjustment during spin up phase on diagnostic mode for experiment Prel-E01.

2.1  Surface velocity field from initial conditions used in experiments SENS-E0x following model spin up in ROMS diagnostic mode.
2.2 (a) Longitude-time Hovmoller diagram of SHA along 27°N for experiment SENS-E04. (b) Dispersion relationship for first mode baroclinic Rossby waves setting Rd equal to 47 km at 27°N, and using equal zonal and meridional wavenumbers (k=1, blue line), and also setting setting the meridional wavenumber to zero (l=0, red line). The black dots on panel (b) indicate the dispersion of SHA signals at 27°N for simulation experiments SENS-E0x. 

2.3 Time-series of transport anomalies observed in the Florida Straits for numerical experiments (a) SENS-E01 to SENS-E07, (b) SENS-E08, and (c) SENS-E09. The black circles overlaid on the lines indicate the timing of specific snapshots of model solution that are emphasized in the text.

2.4 Fields of surface velocity (left column), SSH (center), and meridional velocity at the Florida Straits (right column) at different snapshots of model solution for experiment SENS-E04.

2.5 Fields of surface velocity (left column), SSH (center), and meridional velocity at the Florida Straits (right column) at different snapshots of model solution for experiment SENS-E07.

2.6 Fields of surface velocity (left column), SSH (center), and meridional velocity at the Florida Straits (right column) at different snapshots of model solution for experiment SENS-E08.

2.7 Fields of surface velocity (left column), SSH (center), and meridional velocity at the Florida Straits (right column) at different snapshots of model solution for experiment SENS-E09.
2.8 Location of propagation paths (black lines) used to analyze rate of wall-jet propagation for experiments: (a) SENS-E01 to SENS-E06 along North-west Providence Channel; (b) SENS-E07 north of the Bahamas; and (c) SENS-E08 and SENS-E09 along the east U.S. coast. 48

2.9 Longitude-time Hovmoller diagram of surface velocities along the topographic slope through NWPC for experiments (a-f) SENS-E01 to SENS-E06, and north of the Bahamas for experiment (g) SENS-E07. 49

2.10 Latitude-time Hovmoller diagram of surface velocities along the east US topographic slope for experiment (a) SENS-E08, and (b) SENS-E09. Note that time reference is displayed on the x axis, unlike the Longitude-time Hovmoller diagrams displayed in Figure 2.9, which have time reference displayed on the y axis. 50

2.11 Surface velocity field for experiments (a) SENS-E04 at 161 days, (b) SENS-E07 at 196 days, and (c) SENS-E08 at 280 days. Black lines on panels (a), (b), and (c) show contours of $f/H$, while the magenta lines show the location where values of mean barotropic velocity were computed. (d) Relationship between rate of propagation of jet-like structure and the mean barotropic velocity computed at the entrance of NWPC, north of the Bahamas, and at the US coast. 51

2.12 Surface velocity standard deviation for experiments (a) SENS-E04 (anticyclone at 27°N), (b) SENS-E07 (cyclone at 27°N), (c) SENS-E08 (anticyclone at 28.5°N), and (d) SENS-E09 (cyclone at 28.5°N). All experiments shown here were initialized with features associated with a 500 km wavelength and 15 cm absolute SSH signal. 52
2.13 Initial SST conditions from experiment Cntr_E02, which is derived from fields based on a six years average from HYCOM simulations for the North Atlantic Ocean. 

2.14 Average (a) FC, and (b) AC meridional velocity at the Florida Straits and east of the Bahamas, respectively, from control run experiment Cntr-E02. Please note that the bottom topography at the Florida Straits in the model is ∼20 shallower than the real bottom topography due to topographic smoothing to comply with ROMS requirements to reduce pressure gradient errors (please refer to section 1.4.2).

2.15 Initial conditions of surface velocity for (a) the control run experiment Cntr-E02, and for (b-k) experiments from set ModFC-E0x. Note that the velocity scale is saturated at 30 cm s$^{-1}$ to highlight the surface velocity structure associated with the prescribed eddy-like features in the ocean interior.

2.16 FC transport time-series (thin gray line) from the control run experiment Cntr-E02. The black line shows values low pass filtered at 73 days.

2.17 Standard deviation of meridional velocity in the Florida Straits for experiment Cntr-E02: (a) total standard deviation; (b) high-frequency component (<73 days period); and (c) transient seasonal component (73-525 days period).

2.18 Velocity anomalies in the Florida Straits referenced during (a) maximum, and (b) minimum transport anomalies for experiment Cntr-E02.

2.19 Surface velocity fields for selected snapshots of model solution from the control run experiment Cntr-E02, as described in the text.
2.20 Maps of SHA for control run experiment Cntr-E02 at different snapshots of model solution. Emphasis is given to specific positive (magenta circle) or negative (green circles) SHA signals that are advected by the background circulation, that may generate coastally trapped signals as they reach shallow waters north of 35°N.

2.21 (a) Location of grid-points used to evaluate sea-level variability along the east US coast. (b) Latitude-time Hovmoller diagram for SHA along the east coast of US during experiment Cntr-E02. The black box in panel (b) emphasizes features emphasized in the text.

2.22 Snapshots of model solution for SHA at different times for control run experiment Cntr-E02. Green squares indicates the region where lateral meandering at ∼30°N generates positive southward propagating SHA signals that reaches the Florida Straits approximately 7 days after.

2.23 Longitude-time Hovmoller diagram of SHA along 27°N. SHA values are referred to the mean SSH from experiment Cntr-E02.

2.24 (a) FC transport time-series at the Florida Straits from numerical experiments ModFC-E0x. (b) Transport anomalies from ModFC-E0x experiments with respect to the Control run Cntr-E02.

2.25 Time-series of volume transport anomaly at the Florida Straits (left column), and meridional velocity anomalies (central and right column) with respect to the control run experiment Cntr-E02 for experiments ModFC-E04, ModFC-E05, ModFC-E09, and ModFC-E10.
2.26 Standard deviation from surface velocity anomalies referenced to control run experiment Cntr-E02 for experiments (a) ModFC-E04 (anticyclone at 27°N), (b) ModFC-E09 (cyclone at 27°N), (c) ModFC-E05 (anticyclone at 28.5°N), and (d) ModFC-E10 (cyclone at 28.5°N). All experiments shown here were initialized with features associated with a 500 km wavelength and 15 cm absolute SSH signal.

2.27 Surface velocity fields of snapshots of model solution for experiments ModFC-E04 (anticyclone at 27°N), ModFC-E09 (cyclone at 27°N), ModFC-E05 (anticyclone at 28.5°N), and ModFC-E10 (cyclone at 28.5°N).

2.28 Fields of SHA from experiment ModFC-E04 at different times of model solution. The green lines with “x” marker indicates the track followed by traceable SHA signals linked with the original eddy-like features prescribed in this experiments.

2.29 Track of SHA signals that were associated with the original eddy-like features prescribed in ModFC-E0x experiments. The color code indicate individual ModFC experiments as listed in Figure 2.24.

2.30 Fields of SHA from satellite altimetry data for the dates shown in the upper-left side of each panel. Traceable positive (negative) SHA signals are emphasized by the magenta (green) line.

2.31 Location of zonal sections in the domain that are used for the computation of cumulative meridional volume transport from west to east.
2.32 Longitude-time Hovmoller diagram of cumulative transport from west to east for control experiment Cntr-E02 (left column), for experiment ModFC-E04 (central column), and the difference (right column), at four latitudes. The green “x” on the difference plot shows the projection of the traceable SHA signals linked with the prescribed eddy from ModFC-E04 experiment (see. Figure 2.28). ........................................ 79

2.33 Time-series of jet meandering at (a) 34°N, (b) 32°N, (c) 30°N, and (d) 28°N for the different ModFC-E0x, in comparison to the control run experiment Cntr-E02. Lines are color-coded according to the legend shown in Figure 2.24. ......................................................... 81

2.34 (a) Location of grid-points used to evaluate sea-level variability along the east US coast. Latitude-time Hovmoller diagram for SHA along the east coast of US during experiment (b) Cntr-E02, (c) ModFC-E04, (d) ModFC-E05, (e) ModFC-E09, and (f) ModFC-E10. .............................. 83

2.35 Spatial distribution of correlation coefficients between SHA differences and FC transport differences evaluated as: (a) results from ModFC-E04 minus results from Cntr-E02 at 0 lag; (b) results from ModFC-E09 minus results from Cntr-E02 at 0 lag; (c) results from ModFC-E04 minus results from Cntr-E02 at -21 days lag. Correlation coefficients that are not significant at the 95% confidence level are masked. .............................. 84
2.3.6 Comparison of equivalent outputs from SENS-E0x experiments (no background flow) with ModFC-E0x (with background flow). (left column) Transport anomalies in the Florida Straits observed for the SENS-E0x experiments (blue line) and ModFC-E0x (red line). (center column) Fields of surface velocity from experiments SENS-E04, SENS-07, SENS-E08, and SENS-E09. (right column) Fields of surface velocity from experiment ModFC-E04, ModFC-E09, ModFC-E05, and ModFC-E10.

3.1 Initial condition fields of SSH for experiments: (a) Cntr-E01 - zero flow; (b) EdFul-E01 - elongated Rossby waves; (c) EdFul-E02 - eastern boundary conditions based on satellite altimetry data; (d) Cntr-E02. EdFulFC-E0x are initialized using satellite altimetry data for (e) EdFulFC-E01 - 01-06-1995; (f) EdFulFC-E01 - 01-03-1997; (g) EdFulFC-E01 - 01-05-2001; (h) EdFulFC-E01 - 01-03-2003; (i) EdFulFC-E01 - 01-07-2005.

3.2 Snapshots of model solution of SHA for experiment EdFul-E01.

3.3 (a) Time-series of meridional volume transport anomalies at the Florida Straits for experiment EdFul-E01. (b) Longitude-time Hovmoller diagram of SHA for experiment EdFul-E01. (c) Standard deviation of meridional velocity anomalies developed at the Florida Straits for experiment EdFul-E01.

3.4 Cross-section at 25.6°N showing the meridional velocity component for experiment EdFul-E01 at time: (a) initial condition, and (b) 84 days. The green circle indicates the entrance to Northwest Providence Channel.

3.5 Snapshots of model solution of SHA for experiment EdFul-E02.

3.6 Time-series of meridional volume transport anomalies at the Florida Straits for experiment EdFul-E02.
3.7 (a) Surface velocity field, and (b) velocity section at the Florida Straits for experiment EdFul-E02 at simulation time 5.2 years.

3.8 (a) Time-series of volume transport at the Florida Straits from experiment EdFul-E02 plotted as a function of synthetic month. The mean is shown as the thick black line, while the standard deviation above (below) the mean is shown as the thin line. The blue triangles indicate the weekly averages that are statistically equal to zero with 95% confidence.

3.9 Time-series of volume transports at the Florida Straits for experiments (a) EdFul-E01, (b) EdFul-E02, and (c) EdFulFC-E0x. (d) Volume transport anomalies with respect to results from the control run experiment Cntr-E03.

3.10 Snapshots of model solution of SHA for experiment EdFulFC-E01 referenced to the averaged dynamic topography from the control run experiment Cntr-E02.

3.11 Snapshots of surface velocity fields for experiment EdFulFC-E01 in the proximity of the Florida Straits.

3.12 Plots of volume transport oscillations for individual ModFC-E0x experiments as function of months. The red line shows the transport for the first year of simulation, and the blue line shows the transport for the second year of simulation.

3.13 Similar to Figure 3.8, but for results from experiments EdFulFC-E0x.

3.14 (a) Volume transport anomalies from EdFulFC-E0x experiments as a function of month. (b) Same as panel (a), but using the cable-derived observed transient component of the FC transport.
List of Tables

1.1 Parameters defining configuration of vertical layers on the grid used in this study. ................................................................. 27
1.2 Set of experiments proposed in this study. ............................................... 28
2.1 Properties of experiments from set SENS-E0x in terms of simulated wavelength (twice the eddy diameter) and intensity (eddy maximum SSH), including maximum volume transport response developed in the Florida Straits. ................................................................. 37
2.2 Properties of experiments from set ModFC-E0x in terms of simulated wavelength (twice the eddy diameter) and intensity (eddy maximum SSH), initial location of eddy features, and temporal standard deviation of volume transport variability in the Florida Straits for each experiment. ........ 53
2.3 Results from the multi-linear regression analysis performed for the control run and experiments ModFC-E0x. Slope coefficients \(A_{xx}\) labeled as “FL” indicate coefficients from the west or Florida side of the Straits, while coefficients labeled as “BHS” indicate coefficients from the east or Bahamas side of the Straits. Standard deviation of SHA in cm for each side of the Florida Straits is shown in the first two columns. Physical coefficients have units of \(\text{Sv cm}^{-1}\), while normalized coefficients are non-dimensional. The asterisk indicates coefficients that are not significant at 95% confidence level.

2.4 Same as Table 2.3, except that values reported here are computed using time-series of SHA and transport differences from ModFC-E0x with respect to the control run experiment Cntr-E02.

3.1 Properties of experiments from set EdFul-E0x and EdFulFC-E0x.
Chapter One

Background and Scientific Methods

1.1 Introduction

The Florida Current (FC) is the western boundary current closing the subtropical gyre circulation in the North Atlantic Ocean. Unlike other western boundary currents, the FC carries both the return flow associated with the wind-driven gyre and the upper branch of the Atlantic Meridional Overturning Circulation (AMOC). The total transport associated with the FC is 32 Sv ($1\text{ Sv} = 10^6 \text{ m}^3\text{s}^{-1}$), from which about 20 Sv are generally associated with the wind-driven gyre, and about 12 Sv with the North Atlantic MOC [e.g. Stommel, 1958; Atkinson et al., 2010]. In addition to the FC, the western boundary circulation is also composed by the northward flowing Antilles Current (AC) carrying 3-5 Sv that is associated with North Atlantic’s subtropical gyre [Olson et al., 1984; Lee et al., 1990], and by the southward flowing the Deep Western Boundary Current (DWBC) carrying 26 Sv that is associated with the overturning circulation [Bryden et al., 2005; Johns et al., 2008].

Most of the FC flow has its origin in the Loop-Current, which transport Caribbean waters into the Gulf of Mexico through the Yucatan Channel, and then to the Florida Straits, with transport estimates ranging between 24 to 28 Sv [Johns et al., 2002; Shein-
baum et al., 2002]. Other important sources for the FC flow originate in Old Bahama Channel north of Cuba, and in the Northwest Providence Channel within the Bahamas Archipelago, where previous studies suggest a contribution of $\sim 1.5$ Sv [Johns et al., 2002], and of $\sim 1.4$ Sv [e.g. Beal et al., 2008], respectively. The FC continues flowing through the Florida Straits, feeding into the Gulf Stream, which continues flowing along the east coast of United States (US) until it reaches Cape Hatteras ($\sim 35^\circ$N), where it detaches from the coast and starts flowing eastward.

The FC/Gulf Stream is one of most widely observed and studied [e.g Stommel, 1958; Wunsch et al., 1969; Baringer and Larsen, 2001; Beal et al., 2008; Meinen et al., 2010; Rossby et al., 2010] western boundary currents in the world. Particular attention has been dedicated to this current because it carries the upper branch of the MOC, corresponding to an important component of the climate system. In addition, the FC has been also linked with significant changes in sea-level along the east coast of US [Ezer, 2013; Ezer and Atkinson, 2014], giving it an added level of relevance for consideration by decision makers dealing with coastal resilience. Despite these many studies, there remains many questions about what drives the FC variability. Therefore, given its large importance for earth’s climate system, studies aiming to improve the understanding of the different mechanisms and processes driving the variability of this current at different frequencies are of ultimate relevance.

The FC shows large temporal variability over the record. Measurements at the Florida Straits showed that changes in the FC volume transport can be as large as 10 Sv between the highest and lowest values [Schott et al., 1988]. Among the different time-scales linked with the FC variability, particular interest was dedicated to the the annual cycle exhibited by the FC transport, which is generally characterized as having a peak-to-peak amplitude of 4 Sv, with maximum transport in July [Niiler and Richards, 1973]. Using a 16 years record of continuous volume transport measurements at the Florida Straits, Baringer and
Larsen [2001] reported substantial year-to-year changes in the characteristics of the seasonal variability. The detailed analysis presented in Meinen et al. [2010] showed that the year-to-year changes in the FC annual cycle reported by Baringer and Larsen [2001] were in fact the result of elevated levels of background stochastic variability linked with other processes and time-scales.

While there is general agreement that the FC annual cycle is forced by a combination of along-channel winds with wind stress curl upstream (Caribbean) and downstream (Gulf Stream) from the Florida Straits [Schott et al., 1988], little is known about other components driving the stochastic seasonal variability of the FC. Recent findings [Czeschel et al., 2012; Frajka-Williams et al., 2013; Domingues et al., 2016] showed that westward propagating signals coming from the ocean interior could potentially drive a significant source of variability for the FC at seasonal time-scales. Because observations [Molinari et al., 1990] and results from numerical models [Böning and Budich, 1991] indicate that the seasonal changes in FC transport are linked with seasonal variability in the Meridional Heat Transport (MHT) in the Atlantic Ocean at 26°N, understanding the different components of the seasonal variability of this current is important.

This study aims to improve the understanding of the FC seasonal variability by assessing the forcing mechanisms linking westward propagating signals with changes in the FC transport. This thesis is organized as follows: in section 1.2, detailed background scientific information and motivation for this study is provided, including a detailed assessment on the FC variability (section 1.2.1), and on westward propagating signals (section 1.2.2); section 1.3 lists the main goal and the scientific questions that will guide the development of this study; in section 1.4, the approach adopted in this study is described, including the description of the numerical model employed, and a brief description of numerical experiments developed. In Chapter Two, the analysis focuses on understanding the mechanisms linking the open ocean variability to changes in the FC seasonality. To
accomplish this, experiments based on single-eddy like features prescribed in the ocean interior are developed in the presence and absence of background flows associated with the FC, AC and Gulf Stream. Results are assessed and discussed accordingly. In Chapter Three, the analysis focuses on understanding how more realistic conditions based on eddy-full configurations of westward propagating signals can drive year-to-year changes in the seasonality of transport anomalies in the Florida Straits. To accomplish this, experiments are developed using eddy-full configurations using satellite altimetry data. Results are also analyzed and discussed accordingly. In section 3.3, the main conclusions from this study are drawn.

1.2 Scientific Background

1.2.1 Variability of the Florida Current transport

Unlike most other western boundary currents, the FC has been almost continuously monitored since 1982 at the Florida Straits using telephone cables between Florida and the Bahamas (Figure 1.1). The cable-derived FC transport time-series relies on simple, yet robust, electromagnetic principle: seawater is an efficient electrical conductor, which implies that water motion in the oceans are linked with the generation of electromagnetic potentials and currents [Stommel, 1948]. Daily volume transport measurements are made available in this region by measuring voltage differences between the two sides of the straits. Voltage measurements are then calibrated to provide transport estimates using in situ observations from other platforms, such as based on dropsondes [Larsen, 1992]. The FC cable data is made freely available at NOAA’s Atlantic Oceanographic and Meteorological Laboratory website (http://www.aoml.noaa.gov/phod/floridacurrent/).

The FC transport time-series derived from the telephone cables is a continuous record (gray line, Figure 1.2a) that enabled numerous studies on the different spectral character-
istics of this western boundary current [e.g. Baringer and Larsen, 2001]. For example, Meinen et al. [2010] used the FC transport record during 1982-2008 combined with measurements from other in situ instruments to quantify the contribution of different time-scales to the total variability of the FC transport, reporting that: (i) 46% of the variability was in the 1-11 month frequency band, (ii) 24% was in the sub-month band, (iii) 13% was in the 13-42 month band, (iv) 9% was in the annual band, and (v) 8% was linked with periods larger than 42 months. An updated analysis of the FC transport variability is provided here for the period between 1983-2014. It is quantified that: (i) 40% of the variability was in the 1-11 month frequency band, (ii) 37% was in the sub-month band, (iii) 8% was in the 13-42 month band, (iv) 8% was in the annual band, and (v) 7% was linked with periods larger than 42 months. Below, an analysis of the FC variability for slightly different frequency bands is also provided.

During 1983-2014, the FC transport (Figure 1.2a) exhibits an average volume transport of 31.9 Sv (1 Sv = 10^6 m^3 s^{-1}), with temporal standard deviation (STD) variability of 3.4 Sv. The transport also shows a climatological annual cycle characterized by high transports (∼34 Sv) from July to September, and low transports (∼30 Sv) from November to January (black line, Figure 1.2b). The wavelet transform of the FC transport time-series (Figure 1.2c) indicates that the FC includes with variability in different frequency bands, and that there were significant changes in the spectral characteristics of this current throughout the record. Variability significant at the 95% confidence interval with respect to randomly estimated wavelet spectrum using a Monte Carlo approach is observed for: (a) the high-frequency band with periods smaller than 73 days; (b) an intermediate frequency band with semi-annual and annual periods within the 73-525 days range; and (c) the low-frequency band with two-year periodicity. The wavelet diagram also suggests the presence of 5 year periodicity in the FC transport time-series, however, longer records are needed to confirm the statistical significance of this signal. It is quantified here that
the total variability of the FC transport during 1983-2014 is accounted by: (i) 53% due to high-frequency variability (<73 days, STD = 2.5 Sv); (ii) 35% due to changes in transport within the 73-525 day frequency band (STD = 1.9 Sv); and 12% due to low-frequency variability (> 525 days). The transport variability within the 73-525 day frequency band may be further decomposed as: 8% due to the climatological annual cycle of this current; and 27% due to seasonal variability linked with variable annual phase, previously defined as the transient seasonal component [Domingues et al., 2016].

Different studies addressed the seasonal variability of the FC transport, with particular focus on the climatological annual cycle exhibited by transport measurements across the Florida Straits (Figure 1.2b). The FC annual cycle was first described based on 90 dropsonde transects during 1964-1970 as having an amplitude of 4 Sv, with maximum transport in July [Niiler and Richards, 1973]. There is general agreement that the annual cycle of the FC is driven by a combination of along-channel winds with wind stress curl upstream (Caribbean) and downstream (Gulf Stream) from the Florida Straits [Schott et al., 1988]. Changes in along-channel winds drive anomalous eastward Ekman transport, leading to adjustments in the pycnocline and pressure gradients across the straits [Schott et al., 1988; Rousset and Beal, 2011]. Changes in the upstream/downstream wind stress curl are linked with the generation of barotropic waves that can reach the Florida Straits and adjust the flow of this current within one month [Czeschel et al., 2012].

Even though the FC exhibits a statistical annual cycle linked with the wind forcing, the annual cycle accounts for only 8% of the total FC variability, and seasonal changes recorded for a specific year may exhibit highly anomalous patterns (Figure 1.2b). For example, while in 1989 an amplified annual cycle with similar phase was observed, a 90° phase displacement in the annual cycle was recorded in 1986, and even a semi-annual cycle in 1998. Changes in the annual behavior of the FC were first acknowledged by Baringer and Larsen [2001], who reported that the seasonal variability during 1982-1990
Figure 1.1: Location of the Florida straits. Highlighted are the locations of the telephone cables used to estimate the Florida Current volume transport, and the location of relevant landmarks in the region of study, such as the Northwest Providence Channel (NWPC). Thin black lines show bathymetric contours every 200 m for depths between 100 m and 1000 m, while thin gray lines show bathymetric contours every 20 m for depths shallower than 100 m.

(annual cycle) was substantially different from the variability during 1991-1998 (semi-annual cycle). Additional changes in the FC seasonal variability were also reported for the period during 2001-2007, which was characterized by an annual cycle having a weak semi-annual component, and with small changes in phase [Meinen et al., 2010]. Based on these observations, Meinen et al. [2010] suggested that different processes, other than the wind forcing, may drive the observed interannual changes in the apparent FC seasonal
Figure 1.2: (a) Florida Current transport time-series (raw, light gray line) derived from measurements of voltage differences across the straits of Florida using telephone cables. Overlaid is the climatological annual cycle for the Florida Current during 1983-2014 (red line) and the transient component (blue line). (b) Wavelet transform for the Florida Current transport. The thick white lines highlight peaks in the spectral power that are significant at the 95% confidence level, while the black line marks the cone of influence for the analysis. The magenta lines delimits the 73-525 days frequency band. (c) Seasonal variability of the Florida Current transport during 1983-2014.

variability. Indeed, more recent studies showed that the local eddy field [Frajka-Williams et al., 2013] and baroclinic signals coming from the interior [Czeschel et al., 2012] may drive changes in the seasonal variability of the FC.

Interannual changes in the FC seasonal variability are linked with the transient seasonal component (blue line, Figure 1.2a), which contributes 27% to the total FC variability, which is a substantial portion of the total FC variability. To further understand the impacts of this component in the AMOC and MHT, a comparison between the transient component of the FC transport is performed against the equivalent components from the
Figure 1.3: (a) Time-series of the transient component of the FC transport (black line, left axis), of the AMOC (red line, left axis), and of MHT (blue line, right axis). Linear regression analysis between the transient component of the FC transport and the (b) AMOC, and the (c) MHT.

This analysis shows that the transient component of the FC transport exhibits overall good agreement ($r=0.57$) with the filtered time-series of AMOC and MHT. This results suggest that at least 30% of the year-to-year changes in the AMOC and MHT seasonality can be accounted for by the transient component of the FC. These changes can be as large as $\sim 5$ Sv in the AMOC (Figure 1.3b) and 0.4 PW in the MHT (Figure 1.3c), and are consistent with previous values reported by Atkinson et al. [2010]. Therefore, understanding the mechanisms driving year-to-year changes in the FC seasonality is important.

Observations analyzed by Domingues et al. [2016] showed that westward propagating signals coming from the interior largely explain the variability exhibited by the transient seasonal component of the FC transport. However, a better understanding of the response generated by these signals at the western boundary is still required, and this study aims to
fill this gap. In the following section (section 1.2.2), westward propagating signals in the North Atlantic within the 73-525 day frequency band are introduced and characterized.

### 1.2.2 Westward propagating signals in the North Atlantic

In this section, the characteristics of westward propagating signals in the North Atlantic are assessed. The characteristics described here will be used as background information for designing the numerical experiments developed in this study (Section 1.4.1).

Westward propagating signals or waves correspond to solutions for the linearized quasi-geostrophic equations of motion when the latitudinal-dependency of the Coriolis parameter (i.e. the $\beta$ term) is considered. For example, solutions of the type of first-mode baroclinic Rossby waves may be obtained from the linearized equations of motions considering small perturbations in a one-and-a-half layer ocean model. In this model, the propagation of perturbations follows the dispersion relation for first-mode baroclinic Rossby waves:

$$\omega = \frac{-\beta k}{k^2 + l^2 + R_d^2} , \text{ and } C_p = \frac{\omega}{k}$$ (1.1)

where $\beta$ indicates variations of the Coriolis parameter with latitude, $k$ is the zonal wave number, $l$ is the meridional wave number, $R_d$ is the first-mode radius of deformation, and $C_p$ is the zonal phase speed. Figure 1.4 shows the dispersion relation for first mode, baroclinic Rossby waves at $27^\circ$N using mean temperature and salinity profiles from the World Ocean Atlas 2013 [Locarnini et al., 2013; Zweng et al., 2013] for the region between $82^\circ$W and $61^\circ$W, and $22^\circ$N and $37^\circ$N, which corresponds to the area where numerical experiments are developed in this study (see section 1.4.2).

Even though westward propagating signals may be described by relatively simple dynamics, they are largely observed in the real oceans using satellite-derived sea height
Figure 1.4: Dispersion relationship for first mode baroclinic Rossby waves setting $R_d$ equal to 47 km at 27$^\circ$N, and using equal zonal and meridional wavenumbers ($k=l$, blue line), and also setting the meridional wavenumber to zero ($l=0$, red line).

anomaly (SHA) data from altimetry [e.g. Oliveira and Polito, 2013], and using sea surface temperature (SST) data [e.g. Halliwell Jr et al., 1991]. Previous studies using SHA data [Polito and Liu, 2003] showed that westward propagating signals behaving like first-mode baroclinic Rossby waves account for a large component of the SHA variability in the global oceans. Because SHA data provides overlaid signals from first-mode baroclinic Rossby waves and from nonlinear westward propagating mesoscale eddies, the generic term “westward propagating signals” [Oliveira and Polito, 2013] is adopted here.

Polito and Liu [2003] reported that westward propagating signals with semi-annual and annual time-scales contribute most of the SHA variability in the North Atlantic. In the following analysis, westward propagating signals are characterized for the 73-525 days frequency band, which corresponds to the combined semi-annual and annual waves described by their study. A brief quantitative assessment on the role of these signals in the upper-ocean variability in the subtropical region of the North Atlantic is also described. To accomplish this, weekly fields of gridded SHA with 1/4degree horizontal resolution are obtained from AVISO’s (Archiving, Validation and Interpretation of Satel-

Satellite-derived SHA data shows variability in different time and spatial scales in the North Atlantic. At 27°N, the latitude of cable in the Florida Straits, SHA is linked with STD variability of ∼5 cm close to eastern boundary, which increases westward to ∼13 cm at 76°W, and decrease rapidly to ∼7 cm at the western boundary (blue line, Figure 1.5). The rapid decline in the SHA STD values, or in the eddy variability, at the western boundary is further discussed below. Another important component of the SHA variability is due to a well-defined annual cycle during 1993-2013 (not shown here) that is obtained by calculating weekly averages during 1993-2013. This component is generally linked with the steric variability of sea-level [Guinehut et al., 2006]. The STD linked with the SHA annual cycle shows small changes in the zonal direction, with values ranging between ∼3 cm and ∼5 cm (red line, Figure 1.5). It is estimated here that this component contributes 21% of the total SHA variability at 27°N west of 60°W (60°W corresponds to the eastern limit for numerical experiments developed in this study, Section 1.4.1). After removal of the annual cycle, filtered SHA at the 73-525 days frequency band (SHAr) show STD values ranging between ∼2 cm at the eastern boundary and ∼7 cm at 76°W, displaying a rapid decline west of 76°W. The SHA variability at the 73-525 days frequency band accounts for 42% of the total SHA variability west of 60°W. The low-frequency (periods > 525 days) and high-frequency (periods < 73 days) components show STD values of ∼2 cm in the interior (green and gray lines, Figure 1.5). Close to the western boundary, the high-frequency component show values as large as ∼6 cm. The low-frequency and high-frequency components contribute 12% and 25% to the total SHA variability west of 60°W, respectively. From now on, focus is given to the SHA variability in the 73-525 days frequency band, which is dominated by the semi-annual and annual westward propagating signals studied by Polito and Liu [2003], and to the same frequency band linked
Figure 1.5: SHA Root Mean Square along 27°C N based on: raw (blue line), annual cycle (red line), high-frequency component (< 73 days, gray line), low-frequency component (> 525 days, green line), 73-525 days frequency band (black line).

with the transient seasonal variability of the FC transport described above. As reported here, the 73-525 days frequency band is the main component of the SHA variability west of 60°W. This frequency band also includes first mode, linear, baroclinic Rossby waves, which have a cutoff period and wavelength of 153 days and 293 km, respectively, for the shortest westward propagating wave (Figure 1.4).

After removal of the annual cycle, SHAr exhibits an eddy-like spatial signature in the North Atlantic (e.g., SHAr on March 1st, 2000, Figure 1.6), with amplitudes of ± 15 cm at 27°C N, and greater than ± 20 cm north of 35°C N linked with the Gulf Stream extension. Similar eddy-like westward propagating signals with time-scales similar to the one reported here were also previously identified in this region using SST data [Halliwell Jr et al., 1991]. The westward propagation associated with these features is identified by the sloping SHA pattern using a Hovmoller (longitude-time) diagram of SHAr (e.g., at 27°C N, Figure 1.7a), which indicates that signals travel with phase speed of approximately -4.6 ± 1.4 km day⁻¹ at 27°C N. Phase speeds estimated here are consistent with previous values reported by other studies in the North Atlantic based on SHA [Polito and Liu, 2003], and SST [Halliwell Jr et al., 1991] data. Westward phase speeds from standard linear theory
for first baroclinic Rossby waves can be estimated using real ocean data; in the long wave limit, the dispersion relationship and derived zonal phase speed (Equation 2.7) may be simplified in the form of:

\[ C_p = -\beta R_d^2 \], where \( R_d = \frac{c_1}{f} \) (1.2)

In this equation, \( f \) is the Coriolis parameter, and \( c_1 \) is first mode speed. Estimates of \( c_1 \) can be obtained by solving the modal equation numerically:

\[ \phi_n(z) = -c_n^2 \frac{\partial}{\partial z} \left( \frac{1}{N_s^2} \frac{\partial \phi(z)}{\partial z} \right) \] (1.3)

where \( \phi_n(z) \) provides the vertical structure of mode \( n \), \( c_n \) is the modal speed, and \( N_s \) is the buoyancy frequency. Figure 1.8 exhibits values of \( C_p \) and \( R_d \) calculated for the North Atlantic using climatological temperature [Locarnini et al., 2013] and salinity [Zweng et al., 2013] fields from the World Ocean Atlas 2013. Analysis of \( C_p \) indicates that phase speeds predicted by standard linear theory at 27\(^\circ\)N (-3.5 ± 0.4 km day\(^{-1}\)) are approximately 30\% slower than westward propagating signals revealed by satellite altimetry. Indeed, ocean observations generally show westward phase speeds faster than values predicted by standard theory [e.g. Chelton and Schlax, 1996; Killworth et al., 1997; Oliveira and Polito, 2013], which is in part because potential vorticity gradients induced by background baroclinic flows in the real ocean are neglected in the standard linear theory [Killworth et al., 1997]. However, the evaluation of these mechanisms is beyond the scope of this work.

The power spectral density obtained from a fast fourier transform of SHAr time-series show the intensification of the spectral power towards the western boundary (Figure 1.7b). This indicates an increase in the eddy energy towards the western part of the basin. The power spectrum density also shows a rapid decay in the spectral energy west of 76\(^\circ\)W, which coincides with the rapid decline in the SHA STD close to the western boundary de-
Figure 1.6: SHA field for March 1st, 2000, filtered at the 73-525 days frequency band. Overlaid is the SHA-derived surface geostrophic velocity vectors.

scribed above. A similar decline in the SHA STD (or eddy variability) close to the western boundary has been previously observed by Kanzow et al. [2009], who proposed a mechanism to explain this observation: because normal velocities are not admitted through the boundary, along-boundary pressure gradients are not sustainable, causing pressure anomalies to propagate quickly along the boundary of the basin. This mechanism causes a rapid decay of the eddy energy near the western boundary, which is transmitted equatorward at western boundary regions. Improving the understanding of such mechanisms at the western boundary region of the North Atlantic is one of the motivations for this work, as will be detailed below (Section 1.3).

In order to evaluate the wavelengths linked with westward propagating signals within the 73-525 day frequency band in the North Atlantic, a similar analysis using a fast-fourier transform is applied by transforming the SHAr data from spatial coordinates to wavenumber space. This analysis shows that westward propagating signals at 27°N are linked with a broad band spectrum, with dominant wavelengths ranging between 300 km and 700 km (Figure 1.7c). The relative intensity of these signals may be evaluated by their SHA value. Analysis of the distribution of values associated with westward propagating SHA signals at 27°N indicates that 95% of the time, SHA values associated with these signals range
Figure 1.7: (a) SHA Hovmoller (longitude-time) diagram along 27°N. (b) Normalized power spectrum density from SHAs along 27°N for each longitude. (c) Wavelengths associated with SHA along 27°N estimated using a fast-fourier transform. (d) Histogram distribution of SHA at 27°N west of 70°W. Red lines in (d) indicate values of the 2.5th and 97.5th percentiles from the SHA distribution, which have values of ∼15 cm and ∼15 cm, respectively.

between ± 15 cm (Figure 1.7d). These characteristics will be used later by this study to simulate signals originating in the ocean interior in the numerical experiments developed here (see section 1.4.2).

1.2.3 Eddy-related variability at North Atlantic’s western boundary

The influence of westward propagating signals in the variability of the western boundary current system in the subtropical North Atlantic has long been evident in the in situ ob-
Figure 1.8: (a) Internal radius of deformation based on climatological conditions the Word Ocean Atlas 2013 calculated using Equations. (b) westward propagating phases speeds for long, linear, first-baroclinic Rossby waves.

servations in the region. In the Florida Straits, early observations showed that the overall transport variability observed within individual cruises was comparable and even greater than levels of variability associated with the annual cycle shown by this current [Leaman and Molinari, 1987; Leaman et al., 1987]. These studies reported that transport variability at seasonal time-scales was partially associated with meandering of the FC, and with intermittent signals coming from the Northwest Providence Channel.

The Northwest Providence Channel corresponds to the main passage connecting the Florida Straits with the western boundary region east of the Bahamas (Figure 1.1), where strong variability at seasonal time-scales is also generally observed [e.g. Lee et al., 1990, 1996; Bryden et al., 2005; Johns et al., 2008]. Observations from current measurements during April 1987 to June 1988 by a mooring array east of Abaco, Bahamas, exhibited variability dominated by periodic events occurring every 100 days associated with anticyclonic features centered northeast of the Bahamas [Lee et al., 1990]. Further analysis
by Lee et al. [1996] showed that the variability in the upper 800 m east of Abaco was dominated by westward propagating wavelike baroclinic eddies at mean periods of 30 and 100 days, with wavelengths of approximately 230 and 335 km, respectively. Their study also reported that meridional transport anomalies were linked with strong offshore-onshore meandering of the DWBC east of the Bahamas that could potentially mask the wind-driven annual cycle in the region. The core of the DWBC tended to shift offshore as these westward propagating baroclinic signals interacted with the western boundary at the Bahamas escarpment.

The numerical simulations developed by Böning and Budich [1991] on a 1/3 degree grid provided one of the earliest model-based results showing evidence of the influence of westward propagating baroclinic signals in the the western boundary currents in the North Atlantic at time-scales of 80-120 days. The response simulated by their numerical experiments was generally in good agreement with observations, showing that the wind-driven annual cycle of meridional volume transports east of the Bahamas was masked by strong transport fluctuations with time-scales of 100 days associated with baroclinic signals coming from the interior. Another study on the North Atlantic [Czeschel et al., 2012] based on a adjoint model approach showed that westward propagating baroclinic signals coming from the interior corresponded to a relevant component of the FC annual variability. According to their study, the FC annual variability is driven by both: (a) the wind forcing north of the Florida Straits, which causes a fast barotropic adjustment in the current with amplitude of $\sim$1 Sv; and by (b) long baroclinic Rossby waves coming from the interior, which force a $\sim$0.8 Sv adjustment in the FC transport with different annual phase.

Recent studies based on in situ and SHA observations provided additional insight on the eddy-induced variability of the western boundary current system in the North Atlantic. In one study [Frajka-Williams et al., 2013], coincident fluctuations in Antilles Current
and Florida Current transports were found to be associated with anticyclonic features east of the Bahamas. Their results showed that anticyclonic eddies with diameters of 200 km - 300 km and amplitudes of $\pm 15$ cm coming from the interior could potentially drive changes in the seasonality of the western boundary current system in the subtropical North Atlantic ocean, consistent with the concept of transient seasonal variability described for the FC transport introduced by Domingues et al. [2016]. In Domingues et al. [2016], focus was given to year-to-year changes in the seasonality of the FC transport, which were found to be associated with sea-level changes along the coast due to westward propagating signals reaching the western boundary between $\sim 26.5^\circ$N and $42.0^\circ$N. According to their results, the transport in the Florida Straits “feels” integrated changes in sea-level along the coast, accounting for $\sim 50\%$ of the FC transport variability at these time-scales.

While most of the studies described here have addressed the influence of westward propagating baroclinic signals in the western boundary circulation in the North Atlantic, focusing on variability that occurs at seasonal (73-525 days) time-scales, previous studies also showed that these similar baroclinic signals originating in the ocean interior may also drive interannual changes in the FC transport. For instance, based on the analysis of the FC cable record and wind stress curl data over the North Atlantic Ocean, DiNezio et al. [2009] reported a mechanism by which interannual changes in the wind changes caused baroclinic adjustments in the FC circulation at time-scales of 3-12 years. Therefore, even though the semi-annual and annual westward propagating signals dominate the altimetry record in the North Atlantic [Polito and Liu, 2003], low-frequency baroclinic signals can also have an impact in the western boundary variability in the region.

Results from the studies described above provide valuable evidence of the variability induced by westward propagating signals in the western boundary current system in the subtropical North Atlantic. In the next section (Section 1.2.4), some of the mechanisms
driving the eddy/western boundary interaction are assessed from studies based on idealized analytical and numerical simulations.

1.2.4 Eddy-wall mechanisms

Even though westward propagating signals behaving like first baroclinic Rossby waves have been widely observed in real oceans [e.g. Halliwell Jr et al., 1991; Polito and Liu, 2003; Oliveira and Polito, 2013], isolating their response at western boundary regions can be very challenging. This is because these regions are characterized by highly nonlinear dynamics due to the presence of intense background flows and high-levels of eddy kinetic energy (EKE). The inherent nonlinear dynamics in western boundary regions implies that processes occurring at various temporal and spatial scales can interact, making it hard, or nearly impossible in some cases, to isolate signals linked with specific processes. For example, Byrne et al. [1995] tracked westward propagating eddies shed by the Agulhas Current in the South Atlantic, and found that identifying their signal in regions close to the western boundary was complicated by the presence of the Brazil Current. Therefore, previous studies of westward propagating features have relied on simplified dynamics and idealized numerical experiments to investigate the response generated by eddies and other signals at the western boundary, focusing mainly on: (a) the evolution of pressure anomalies (dynamic height) along the boundary; and on (b) the fate of the eddy itself. In this section, the underlying dynamics and main outcome from these earlier experiments are summarized.

The study by Milliff and McWilliams [1994] focused on the evolution of pressure anomalies along the boundary of ocean basins given an initial perturbation in the ocean interior. They used two numerical models based on the shallow water equations and on the quasi-geostrophic equations initialized with a monopole vortex (anticyclone) of 200 km diameter on a rectangular, and flat-bottom ocean basin. Some of the main results
reported by their study include: (1) the westward propagation of the vortex at a rate consistent with the dispersion of Rossby waves; (2) the development of a Rossby wave field in the wake of the vortex; (3) the development of coastally trapped waves when the vortex encountered the boundary; (4) the scattering of signals from coastally trapped waves along the meridional boundary; (5) the completion of vortex-boundary interaction after 120 days of simulation; and (6) residual variability at the western boundary after 120 days caused by westward propagating Rossby waves that developed in the wake of the original vortex. Their experiments provided a mechanism coupling the short timescale motions in the coastal waveguide with longer time-scale motions in the ocean interior. Similar results were also found by Kanzow et al. [2009] using a nonlinear one-and-a-half layer reduced-gravity model in a rectangular, flat-bottom ocean basin. Results by Kanzow et al. [2009] also showed that these boundary-trapped waves cause the observed decline of sea-level variability along the coast, once they provide the mechanism for the fast export of transport anomalies equatorward due to eddies impinging in the western boundary.

With respect to the fate of the eddy once it reaches the western boundary, previous studies by Nof [1988] and Nof [1999] provided a detailed analysis of mechanisms and time-scales involved using a nonlinear, inviscid one-and-a-half layer model. Based on analytical and numerical experiments initialized with a zero potential vorticity anticyclonic eddy \((\partial u / \partial y = f)\), Nof [1999] reported that: (1) as the ring moves into the wall, it gradually leaks its content equatorward along the boundary until it loses all of its mass, like a peeling onion, forming a wall-jet feature along the wall; (2) the “peeling” rate is \(2\beta R_d^2/9\), implying that the eddy migrates towards the wall at a rate that is one-third of the free open ocean rate controlled by the dispersion of Rossby waves; (3) upon contact with the western boundary, anticyclonic eddies tend to stay at a fixed latitude due to two sub-processes in approximate balance, the (a) \(\beta\) equatorward tendency, and (b) the rocket effect, which causes a poleward tendency due to the equatorward leakage of the eddy along at the wall.
According to their study, a typical intense eddy would take $\sim 200$ days to “leak” $70\%$ of its volume. Even though they acknowledged that identifying such mechanisms in the real oceans would be difficult, they suggested that similar processes may occur in various western boundary regions, such as at the Gulf Stream.

An additional case considered by Simmons and Nof [2002] was the interaction of eddies that encountered porous western boundaries regions, such as the islands of the Lesser Antilles in the tropical North Atlantic. To accomplish this, they developed sets of experiments using different boundary configurations in the same model as Nof [1999].

Their results revealed that: (1) throughout the encounter with the porous boundary, eddies remained axisymmetric while being drained by wall jets; (2) while leaking, the eddy diameter ultimately adapted to the gap width, so that it would lose contact with the walls and drift slowly into the interior of the western basin; (3) in the multiple gap problem, all the fluid from the approaching eddy penetrated into the interior of the western basin, in contrast to most of the single gap problems; and that (4) eddies that encountered large islands showed a tendency to break into smaller offsprings. They also showed that even large eddies can pass their volume through narrow passages, estimating that $85\%$ of the content from a anticyclonic eddy with $280$ km diameter would be able to flow through a passage $80$ km wide. Their results indicate that the signal from eddies and other mesoscale features can efficiently cross islands arcs, such as the Bahamas archipelago.

The western boundary region in the subtropical North Atlantic has some characteristics that are similar to configurations used in the experiments described above, such as the combination of a solid wall (steep slope) north of $27^\circ$N, and a porous wall south of $27^\circ$N due to the Bahamas archipelago. Therefore, similar responses may be observed at the North Atlantic’s western boundary, to some extent. While these studies provided insight on the mechanisms linked with eddy-wall interactions based in highly idealized experiments, here we take a further step towards understanding the variability induced
by westward propagating signals at the western boundary region in the subtropical North Atlantic by performing sets of controlled numerical experiments with a more realistic high-resolution topography (Section 1.4.2).

1.3 Goal and Research Questions

The goal of this study is to improve the understanding of the FC transient seasonal variability. To accomplish this, I intend to address the following questions:

(Q1) Can westward propagating signals drive transport variability at seasonal time-scales (73-525 day) in the Florida Straits?

(Q2) What are the mechanisms driving the response at North Atlantic’s western boundary?

(Q3) How is the response different for anticyclonic and cyclonic eddies?

(Q4) What is the vertical structure of the responses induced in the Florida Straits?

(Q5) How do westward propagating signals in different latitude bands contribute to the variability at 27°N in the Florida Straits?

(Q6) How does the background FC flow modulate the response in the western boundary?

(Q7) Can westward propagating signals drive year-to-year changes in the Florida Straits seasonality at 27°N?
1.4 Methods

In this study, I will simulate in detail the response generated by westward propagating signals at the western boundary in the subtropical North Atlantic. To address the questions listed in Section 1.3, this study employs a robust approach combining a sophisticated numerical model that enables realistic oceanic simulations using a realistic bottom topography (Section 1.4.1), with controlled experiments based on idealized initial conditions (Section 1.4.2).

1.4.1 Regional Ocean Modeling System - ROMS

The research developed here is based on numerical experiments using the Regional Ocean Modeling System (ROMS, available at https://www.myroms.org/). ROMS consist of a free-surface, terrain-following ocean model that solves the hydrostatic primitive momentum equations in a staggered Arakawa C-grid [Shchepetkin and McWilliams, 2005].

Because ROMS provides a wide range of modules and configurations, it has been widely used by the scientific community, with applications ranging from studies focused on sediment transport at the coast [Blaas et al., 2007] to studies making long-term climate projections [Ådlandsvik, 2008]. In this study, the following configuration is used: (i) non-linear equation of state; (ii) analytic spherical grid; (iii) centered, fourth order advection of tracers and momentum; (iv) harmonic horizontal mixing of tracers and momentum; (v) KPP vertical mixing [Large et al., 1994]; and the (vi) splines density Jacobian for the pressure gradient computation [Shchepetkin and McWilliams, 2000].

For all experiments, boundary conditions are applied on all boundaries using: Flather boundary condition [Flather, 1976] for the free-surface; Chapman boundary conditions [Chapman, 1985] for the 2-D barotropic velocity; Gradient boundary condition for mixing turbulent kinetic energy; and Mixed Radiation-Nudging boundary conditions [March-
esiello et al., 2001] for baroclinic velocity, and for temperature and salinity. The Mixed Radiation-Nudging boundary condition enables a better control of inflow into the domain by setting nudging time-scales to 1 day at the boundary, while permitting the outflow/radiation of anomalies from inside the domain. Nudging time-scales to boundary values are set to one day, except where otherwise stated for a few experiments.

1.4.2 Numerical experiments

Experiments developed in this study are carried out in a domain within 60°W - 82°W, and 21°N - 37°N (Figure 1.9). The grid used for numerical simulations has an horizontal resolution of approximately 1/25degree, with 540 × 390 points in the longitudinal and latitudinal direction, respectively, and 30 vertical layers. The bottom topography is derived from ETOPO-1 [Amante and Eakins, 2009], which is then smoothed to to control the maximum grid stiffness by setting the Beckman and Haidvogel parameter (rx0) to 0.1, and the Haney parameter (rx1) to 2.7. These configurations of rx0 and rx1 are selected to avoid spurious and unrealistic flows resulting from pressure-gradient errors [Beckman and Haidvogel, 1993; Haney, 1991]. Because this research focuses on the response of the FC driven by westward propagating signals arriving at the western boundary, realistic bottom topography is used at and in the proximity of the east coast of the United States (Figure 1.9a,b). The bathymetry at the ocean interior, on the other hand, is simplified by setting the maximum depth to 5000 m, and by enforcing flat-bottom conditions east of 70°W. Parameters describing the configurations of vertical layers are available in Table 1.1.

Different sets of experiments are developed in this study (Table 1.2) that are designed to directly address the questions listed in Section 1.3. Numerical experiments are developed using initial conditions constructed based on different configurations of westward propagating signals prescribed as small perturbations in the pycnocline depth in the ocean
Figure 1.9: (a) Model domain used for numerical experiments developed in this study using ROMS. The grid has a horizontal resolution of approximately 1/25 degree with 540 X 390 points. (b) Grid configurations shown in detail at the Florida Straits. The magenta lines show the location of sections across (c) the Florida Straits, and (d) the Northwest Providence Channel, showing the configuration of vertical layers at these locations.

interior. Additional information describing the method for including westward propagating signals in initial condition fields are provided in section 1.4.3. All experiments are developed in the absence of wind forcing, in order to isolate the response driven by westward propagating signals in the variability at the western boundary region.

The first few sets of experiments are developed in the absence of the FC flow to enable an initial understanding of the responses generated by westward propagating signals impinging along the east coast of United States. Later, experiments are developed with
Table 1.1: Parameters defining configuration of vertical layers on the grid used in this study.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>N</td>
<td>30</td>
<td>Number of vertical layers</td>
</tr>
<tr>
<td>Vtransform</td>
<td>2</td>
<td>Transformation equation</td>
</tr>
<tr>
<td>Vstretching</td>
<td>4</td>
<td>Vertical stretching function</td>
</tr>
<tr>
<td>Theta_s</td>
<td>4</td>
<td>Surface stretching parameter</td>
</tr>
<tr>
<td>Theta_b</td>
<td>0</td>
<td>Bottom stretching parameter</td>
</tr>
<tr>
<td>Tcline</td>
<td>10</td>
<td>Critical depth [m]</td>
</tr>
</tbody>
</table>

the FC to evaluate the role of the background flow from this current in modulating the responses at the western boundary.

In the set of experiments designated SENS-E0x, a sensitivity analysis is performed using numerical simulations initialized without the FC and with single anticyclonic or cyclonic eddies with different intensity and wavelength configurations. With these configurations, I hope to: assess whether westward propagating eddies can drive variability in the Florida Straits at seasonal time-scales (Q1); identify the mechanisms driving the response at the western boundary (Q2); analyze differences in the response driven by anticyclonic eddies and by cyclonic eddies (Q3); and also to evaluate the vertical structure associated with the different responses at the western boundary (Q4). assess the contribution of eddies impinging in different locations along the boundary for driving variability at the Florida Straits (Q5); More information about this experiment set is provided in section 2.1.

In the set of experiments designated EdFul-E0x, simulations are initialized with constant inflow of westward propagating signals using altimetry data as boundary conditions. Experiments are designed mostly to investigate the impact of westward propagating signals on driving year-to-year changes in the seasonal transport variability in the Florida
Table 1.2: Set of experiments proposed in this study.

<table>
<thead>
<tr>
<th>Experiment</th>
<th>Relationship to Questions</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>Cntr-E01</td>
<td>N/A</td>
<td>Control run initialized with flat isopycnals, no FC, and no eddies</td>
</tr>
<tr>
<td>Prel-E01</td>
<td>Q1</td>
<td>Preliminary experiment initialized with no FC, and single anticyclone at 27°N, 70.5°W</td>
</tr>
<tr>
<td>Prel-E02</td>
<td>Q1</td>
<td>Preliminary experiment initialized with no FC, and train of eddies along 27°N</td>
</tr>
<tr>
<td>Prel-E03</td>
<td>Q1</td>
<td>Preliminary experiment initialized with no FC, and single anticyclone at 29°N, 76°W</td>
</tr>
<tr>
<td>SENS-E0x</td>
<td>Q1, Q2, Q3, Q4</td>
<td>Set of sensitivity experiments initialized with no FC, and single anticyclonic or cyclonic eddies in different latitudes</td>
</tr>
<tr>
<td>EdFul-E0x</td>
<td>Q5, Q7</td>
<td>Set of experiments initialized with no FC, and different configurations of train of waves/eddies using synthetic features and AVISO’s data as boundary conditions</td>
</tr>
<tr>
<td>Cntr-E02</td>
<td>N/A</td>
<td>Control run initialized with background flow from the FC, and no eddies in the interior</td>
</tr>
<tr>
<td>ModFC-E0x</td>
<td>Q6</td>
<td>Similar to experiments SENS-E0x, but with the FC flow</td>
</tr>
<tr>
<td>Cntr-E03</td>
<td>N/A</td>
<td>Similar configurations as Cntr-E02, but used as the control run for EdFulFC-E0x experiments</td>
</tr>
<tr>
<td>EdFulFC-E0x</td>
<td>Q7</td>
<td>Similar to experiments EdFul-E0x, but with the FC flow</td>
</tr>
</tbody>
</table>

Straits (Q7). More information about the setup of these experiments is found in section 3.1;
The set of experiments designated ModFC-E0x will be equivalent to experiments in SENS-E0x, but this time with the background flow of the FC also included. With this analysis, we intend to understand how the background flow of the FC modulates the response at the western boundary (Q6). Section 2.2 contains more information on the setup of this set of experiments. Similarly, the set of experiments designated EdFulFC-E0x is also equivalent to EdFul-E0x experiments, but this time with the FC flow. This set of experiments will enable the evaluation of year-to-year changes in the variability of seasonal oscillations of the FC flow that are induced by signals originating in the ocean interior. The full list of experiments developed in this study is given in Table 1.2.

1.4.3 The feature model approach

This section describes the technique used for introducing different configurations of westward propagating signals in our simulations. Westward propagating signals are prescribed in the initial and/or boundary condition fields as perturbations in the pycnocline depth. To accomplish this, we rely on concepts introduced by the feature model approach [Robinson et al., 1988], which consists in creating oceanographic features of interest using mathematical functions [Calado et al., 2008].

In this study, eddy-like perturbations on isopycnal surfaces are constructed using two functions: (i) a Gaussian function ($G_{xy}$, Equation 1.4) and (ii) a modified first baroclinic mode for vertical velocity ($\phi_{1m}(z)$, black line, Figure 1.10e). The Gaussian function $G_{xy}$ defines the horizontal structure and magnitude of isopycnal displacements, and is calculated according to the following formula:

$$G_{xy}(x) = \Delta z_{\text{max}} \times \exp \left( \left\{ -\frac{2xe}{D_E} \right\}^2 \right), \text{ with } x \in \left[ -\frac{D_E}{2}, \frac{D_E}{2} \right] \quad (1.4)$$

In this equation, the parameter $D_E$ defines the horizontal diameter of the eddy, and $\Delta z_{\text{max}}$
Figure 1.10: (a) Temperature, (b) salinity, and (c) density profiles used as background conditions for model initialization, and its derived (d) buoyancy frequency. Baroclinic modes linked with the (e) vertical velocity and density structure, and with the (f) horizontal velocity and pressure field. Modes are solved using a rigid-lid boundary condition. In panel (e), the modified first baroclinic mode for vertical velocity ($\phi_{1m}(z)$) is shown as the thick black line.

sets the maximum vertical displacement of the pycnocline depth. If $\Delta z_{\text{max}}$ is negative, perturbations on isopycnal surfaces will lead to an anticyclonic eddy, while a positive $\Delta z_{\text{max}}$ will lead to a cyclonic eddy. Figure 1.11 provides a schematic view of $G_{xy}$.

The modified first baroclinic mode for vertical velocity is used to define the vertical structure of the vertical displacement of individual isopycnal surfaces. The original (rigid-lid) first baroclinic mode (gray line, Figure 1.10e) is obtained solving Equation 1.3.
using the mean temperature and salinity profiles (Figure 1.10a,b) for the domain. \( \phi_{1m}(z) \)
is then calculated by adding a linear profile with maximum value of 0.5 at the surface
and minimum value of 0 at \( \sim 1300 \) m, which is the depth associated with the maximum
value observed for this mode. This modification of \( \phi_1(z) \) is performed to enable the devel-
opment of surface velocities and free-surface signals, which would be otherwise absent
if using the original mode based on a rigid-lid approximation. Vertical perturbations on
isopycnal surfaces are ultimately achieved using:

\[
\frac{\partial \rho}{\partial t} = -w \frac{\partial \rho}{\partial z} \quad (1.5a)
\]

\[
\rho' = -(w \Delta t) \frac{\partial \rho}{\partial z} \quad (1.5b)
\]

\[
\rho' = -(\Delta Z) \frac{\partial \rho}{\partial z} \quad (1.5c)
\]

where \( \Delta Z \) is obtained as:

\[
\Delta Z(x, z) = G_{xy}(x) \times \phi_{1m}(z) \quad (1.5d)
\]

Application of equation 1.5d on a circular grid using the Gaussian function \( G_{xy}(x) \) results
in eddy-like perturbations on isopycnal surfaces. Figure 1.12 shows an example of the
temperature, salinity, and density structure of a single anticyclonic eddy prescribed at
27°N and 70.5°W using \( D_E \) equal to 150 km and \( \Delta z_{max} \) equal to 300 m. A similar method
can be used to prescribe perturbations on isopycnal surfaces with using horizontal shapes
(e.g. asymmetrical features), which can be accomplished by replacing the original \( G_{xy}(x) \)
function in equation 1.5d by any other function of interest. For example, the Gaussian
function may be replaced by a function derived from satellite-altimetry
Figure 1.11: Gaussian function used to apply vertical displacements on isopycnal surfaces, where $D_E$ is the radius of the eddy, and $\Delta Z_{\text{max}}$ is the maximum vertical displacement allowed.

Figure 1.12: (a) Temperature, (b) salinity, and (c) density structure of an anticyclonic feature used as initial conditions for experiment Prel-E01.

observations (Figure 1.13). More details about this modified approach is provided in section 3.1.

The approach described above is used to create perturbations in the initial temperature and salinity fields that are configured according to the specific characteristics of experiments developed in this study. After the creation of perturbed initial conditions fields, numerical experiments are started from rest in the model’s diagnostic mode, which can be accomplished in ROMS by activating the flag $TS\_FIXED$. The diagnostic mode implies the evolution of momentum fields (free-surface, baroclinic and barotropic velocity) with tracers fields (temperature and salinity) held constant. This approach enables the
Figure 1.13: Example of vertical displacements of isothermal surfaces using SHA data as the horizontal function in equation 1.5d, instead of the original Gaussian function (Figure 1.11).

Geostrophic adjustment of momentum fields to the perturbations in the pycnocline depths prescribed as initial temperature and salinity conditions; it prevents the partial dispersion of the available potential energy during initialization due to the propagation of internal waves. Experiments are developed in diagnostic mode until the full adjustment of momentum fields to the prescribed perturbations on isopycnal surfaces. The full adjustment can be identified once the domain-averaged kinetic energy levels out, indicating the development of steady-state conditions. Once the geostrophically adjusted momentum and tracers fields are obtained, experiments are allowed to evolve in time using ROMS in prognostic mode, which is accomplished by deactivating the flag TS\_FIXED. Figure 1.14 shows an example of the adjusted momentum fields for preliminary experiment Prel-E01, which was initialized with a single eddy-like isopycnal perturbation in the ocean interior. Snapshots of time-averaged model solution are saved every 7 days of simulation time.
Figure 1.14: (a) Surface velocity field, (b) surface relative vorticity field, and (c) vertical velocity structure of the eddy after full adjustment during spin up phase on diagnostic mode for experiment Prel-E01.
Chapter Two
Case Studies of Eddy-Western Boundary Interaction

In this chapter, specific methods and results from experiments initialized with single eddy-like perturbations are assessed. This includes the experiments from sets SENS-E0x and ModFC-E0x, which aim to evaluate specific mechanisms by which eddy-like westward propagating signals drive variability in the Florida Straits with and without the background flows associated with the FC, AC, and the Gulf Stream, respectively.

As described in section 1.2.2, westward propagating signals with seasonal time-scales in the North Atlantic are characterized as eddy-like features with broad band spectrum dominated by wavelengths ranging between 300 km and 700 km at 27°N (Figure 1.7c). They are associated with STD SHA variability of approximately 7.5 cm (Figure 1.5), and during 95% of the time their sea-level signature has values ranging between -15 cm and 15 cm (Figure 1.7d). These characteristics are used as background parameters for introducing westward propagating signals in my modeling experiments.

2.1 No Background Flow Case Study

Experiments from set SENS-E0x are designed to assess the response in the Florida Straits to single anticyclonic or cyclonic eddies with different intensity and wavelength configurations (Figure 2.1). Two additional experiments are also designed to assess the sensitivity
of the response in the Florida Straits to signals reaching the western boundary north of the Florida Straits at \( \sim 29^\circ \text{N} \).

### 2.1.1 Experiment setup

In the first seven experiments from set SENS-E0x (Table 2.1), eddy-like perturbations with different wavelength and intensity configurations are prescribed east of the Bahamas at 74\(^\circ\)W, with the latitudinal position defined according to the type of feature: anticyclonic eddies are initialized at 27\(^\circ\)N; cyclonic eddies are initialized at 25\(^\circ\)N. This configuration is applied in order to ensure that cyclonic and anticyclonic eddies reach the western boundary in the proximity of Northwest Providence Channel entrance (Figure 1.1), since anticyclonic (cyclonic) eddies have a component of southward (northward) propagation as they move to the west. The two additional experiments designated SENS-E08 and SENS-E09 are initialized with features north of the Florida Straits. Both experiments are developed using eddy-like features associated with wavelength of 500 km and absolute SSH of 15 cm (intensity), except that experiment SENS-E08 is initialized with an anticyclonic eddy and experiment SENS-E09 is initialized with a cyclonic eddy.

For all SENS-E0x experiments, flat-isopycnal conditions are imposed everywhere in the domain by applying the average temperature, salinity, and density profiles (Figure 1.10a,b,c) from the World Ocean Atlas 2013 [Locarnini et al., 2013; Zweng et al., 2013]. Flat-isopycnal conditions are initially prescribed everywhere in the domain to ensure stationary initial conditions, removing background baroclinic flows that are intrinsic to real-ocean climatological fields (e.g due to horizontal density gradients).

Eddy-like perturbations are prescribed in the initial conditions fields of each experiment using the method described in section 1.4.3, applying the original Gaussian function from equation 1.5d. After the creation of initial condition fields, experiments are carried out with ROMS using the procedure also described in section 1.4.3. Experiments are in-
Table 2.1: Properties of experiments from set SENS-E0x in terms of simulated wavelength (twice the eddy diameter) and intensity (eddy maximum SSH), including maximum volume transport response developed in the Florida Straits.

<table>
<thead>
<tr>
<th>Exp. Name</th>
<th>Wavelength</th>
<th>Intensity [SSH]</th>
<th>Initial location</th>
<th>FL Straits resp.</th>
</tr>
</thead>
<tbody>
<tr>
<td>SENS-E01</td>
<td>500 km</td>
<td>7.5 cm</td>
<td>27°N, 74°W</td>
<td>0.53 Sv</td>
</tr>
<tr>
<td>SENS-E02</td>
<td>300 km</td>
<td>15.0 cm</td>
<td>27°N, 74°W</td>
<td>1.04 Sv</td>
</tr>
<tr>
<td>SENS-E03</td>
<td>400 km</td>
<td>15.0 cm</td>
<td>27°N, 74°W</td>
<td>1.12 Sv</td>
</tr>
<tr>
<td>SENS-E04</td>
<td>500 km</td>
<td>15.0 cm</td>
<td>27°N, 74°W</td>
<td>1.08 Sv</td>
</tr>
<tr>
<td>SENS-E05</td>
<td>600 km</td>
<td>15.0 cm</td>
<td>27°N, 74°W</td>
<td>0.94 Sv</td>
</tr>
<tr>
<td>SENS-E06</td>
<td>700 km</td>
<td>15.0 cm</td>
<td>27°N, 74°W</td>
<td>0.87 Sv</td>
</tr>
<tr>
<td>SENS-E07</td>
<td>500 km</td>
<td>-15.0 cm</td>
<td>25.5°N, 74°W</td>
<td>-0.74 Sv</td>
</tr>
<tr>
<td>SENS-E08</td>
<td>500 km</td>
<td>15.0 cm</td>
<td>29.5°N, 75°W</td>
<td>1.08 Sv</td>
</tr>
<tr>
<td>SENS-E09</td>
<td>500 km</td>
<td>-15.0 cm</td>
<td>27°N, 75°W</td>
<td>1.08 Sv</td>
</tr>
</tbody>
</table>

integrated for up to 330 days to allow the westward propagation of initial perturbations, and the interaction of these signals with the western boundary.

Boundary conditions for SENS-E0x experiments are configured for all boundaries (east, west, north, south) as follows: flat-isopycnal conditions consistent with temperature and salinity fields from the interior of the domain are imposed everywhere along the boundary; zero 2-D barotropic and 3-D baroclinic fields are also imposed everywhere along the boundary. Boundaries are configured according to options described in section 1.4.1. The Mixed Radiation-Nudging boundary conditions combined with values imposed along the boundary implies the absence of active flows into the domain, while still allowing anomalies to exit the domain through radiation conditions.

The SENS-E0x experiments simulate eddies with different sizes corresponding to the range of the dominant wavelengths (~300-700 km) observed in the filtered altimetry record for the North Atlantic Ocean at 27°N (Figure 1.7c). Prescribed anticyclonic (cyclonic) eddies are assumed to represent the wave crest (through) associated with a Rossby wave field. Therefore, to simulate the dominant wavelengths observed in the
altimetry record, eddies with diameters equal to half of the desired wavelength are employed. The analysis of these preliminary experiments is intended to provide insight for addressing questions Q1, Q2, Q3, and Q4 raised by this study (see section 1.3).

2.1.2 Results

In this section, results from experiments forced by single eddy-like reaching the western boundary at $\sim 26.5^\circ$N (SENS-E01 to SENS-E07) are analyzed first, followed by the analysis of experiments forced by features north of the Florida Straits. The initialized eddies for each of the experiments in this set are shown in Figure 2.1.

The experiments are initially characterized by the westward propagation of the prescribed eddy-like features in the ocean interior. For example, the Longitude-time Hovmoller diagram for experiment SENS-E04 (Figure 2.2a) indicates that the eddy and its Rossby wave wake signals propagate westward with phase speed of $-2.30 \pm 0.22$ km day$^{-1}$. Analysis of other experiments in which the prescribed eddy-like features also reached the western boundary at $\sim 26.5^\circ$N (SENS-E01 to SENS-E07) indicates that signals propagated westward with average phase speed of $-2.36 \pm 0.18$ km day$^{-1}$. These propagation speeds are roughly consistent with the expected theoretical speeds of first-mode Rossby waves with wavelengths matching the eddy scales (Figure 2.2b). Similar westward propagation is also observed for signals prescribed North of the Florida Straits in experiments SENS-E08 and SENS-E09 (not shown).

While westward propagation is a prominent feature during the initial phase of SENS-E0x experiments, the prescribed perturbations also display meridional motion as they propagate westward. In general, anticyclonic eddies prescribed in experiments SENS-E01 to SENS-E06, and SENS-E08 exhibited a tendency for moving southwestward, while cyclonic prescribed eddies in experiments SENS-E07, and SENS-E09 tended to move
northwestward. The opposing weak meridional drifts of cyclonic and anticyclonic eddies are the result of the $\beta$ effect combined with self advection [Chelton et al., 2007].

Following westward propagation, eddy-like features reached the western boundary region at the Bahamas archipelago after $\sim$100 days of simulation in experiments SENS-E01 to SENS-E07. Before investigating the different processes causing the variability in the Florida Straits, a description of total transport variability observed for the different experiments is provided. Time-series of volume transport anomalies in the Florida Straits (Figure 2.3a) show anomalies reaching values of: (i) $\sim$1 Sv for experiments SENS-E02 to SENS-E06 (anticyclonic eddies associated with 15 cm SSH and wavelengths of 300-700
Comparison of signal characteristics at the Florida Strait (a) Longitude-time Hovmoller diagram of SHA along 27°N for experiment SENS-E04. (b) Dispersion relationship for first mode baroclinic Rossby waves setting Rd equal to 47 km at 27°N, and using equal zonal and meridional wavenumbers (k=l, blue line), and also setting setting the meridional wavenumber to zero (l=0, red line). The black dots on panel (b) indicate the dispersion of SHA signals at 27°N for simulation experiments SENS-E0x.

km); (ii) 0.5 Sv for experiment SENS-E01 (anticyclonic eddy associated with 7.5 cm SSH and wavelength of 500 km); and (iii) -0.8 Sv in experiment SENS-E07 (cyclonic eddy associated with -15 cm SSH and wavelength of 500 km). The time-frame of the response in the Florida Straits is consistent between the different experiments that are associated with eddy-like features reaching the boundary at ~26.5°N (SENS-E01 to SENS-E07), and lasts for a total duration of about 200 days. This timescale is within the 73-525 days frequency band that is the focus of this study.

Compared to experiments that were associated with signals reaching the boundary at 26.5°N (SENS-E01 to SENS-E07), experiments that were initialized with eddy-like features north of the Florida Straits (SENS-E08 and SENS-E09) exhibited lower range transport anomalies (Figure 2.3b,c), with maximum amplitudes of ~0.4 Sv. The characteristics of these responses are more variable, but the corresponding transport fluctuations have similar time-scales of ~100-200 days.
In order to investigate the mechanisms driving volume transport anomalies in the Florida Straits, an analysis based on fields of surface velocity, SHA, and velocity sections in the Florida Straits are shown here for snapshots the model solutions denoted in the transport time-series (black circles, Figure 2.3). Results for experiment SENS-E04 indicate that transport anomalies in the Florida Straits at day 70 (Figure 2.3a) started building up before the eddy made initial contact with the topographic slope at the Bahamas (Figure 2.4a-c). At day 70, the transport anomaly of \( \sim 0.5 \) Sv is associated with SHA signals that reach the Bahamas side of the Florida Straits before the eddy (Figure 2.4b). SHA signals are rapidly transmitted along topographic contours, reaching the east...
side of the Florida Straits as broad SHA signals (not shown). These SHA signals are linked with northward velocity anomalies of \( \sim 1 \text{ cm s}^{-1} \) that occupy most of the water column in the eastern Florida Straits (Figure 2.4c), indicating a barotropic response. This result indicates that prescribed eddies generate some band-limited range of Rossby wavenumbers, some of which propagate faster than the eddy. The Rossby wave field is associated with broad SSH anomalies that rarely exceed 2 cm. Once the eddy reaches the continental slope east of the Bahamas, it remains in an approximate fixed location as its volume is slowly drained, forming a jet-like feature resembling a wall-jet [Nof, 1999] that propagated through the NWPC (Figure 2.4d). At the same time, fluid is drawn eastward along the northern side of the Bahamas Banks that joins the anticyclonic flow around the eddy. As this process takes place, the eddy diameter and magnitude becomes progressively smaller, taking approximately 200 days for all the eddy volume to be drained through the arc of islands. These wall-jets had a width of \( \sim 40 \text{ km} \) and surface velocities reaching values of \( \sim 20 \text{ cm s}^{-1} \), and were often associated with the development of other smaller eddy-like features within the NWPC (Figure 2.4d). At the time of maximum transport anomaly (day 119), the wall-jet had just reached the Straits. Velocity anomalies at this time showed some vertical shear, with surface velocity reaching values of \( \sim 5 \text{ cm s}^{-1} \). At day 210, transport anomalies in the Florida Straits started decaying (Figure 2.3). At this time, surface velocity field (Figure 2.4e) indicates that the jet originated a small anticyclonic eddy within the Florida Straits that is characterized by a velocity structure (Figure 2.4i) dominated by vertical shear, with surface values reaching values of \( \sim 20 \text{ cm s}^{-1} \). The evolution of experiment SENS-E04 is characteristic of each of the experiments SENS-E01 to SENS-E06, which were all associated with anticyclonic features reaching the western boundary at 26.5°N.

The evolution of the experiment based on a cyclonic eddy (SENS-E07) is overall similar to the evolution of experiments with anticyclonic eddies (SENS-E01 to SENS-
Figure 2.4: Fields of surface velocity (left column), SSH (center), and meridional velocity at the Florida Straits (right column) at different snapshots of model solution for experiment SENS-E04.

E06), except that a negative transport anomaly reaching -0.8 Sv is observed at day 189 in the Florida Straits (Figure 2.3a). On this experiment, negative volume transport anomalies in the Florida Straits develop before the eddy reaches the Bahamas archipelago on day 119 (Figure 2.5a). At this time, transport anomalies are linked with broad SHA signals that dominate the east side of the Straits (Figure 2.5b). Similar to the anticyclonic cases, SHA signals linked with the fast propagating Rossby wavenumbers may drive broad barotropic
velocity anomalies in the Florida Straits (Figure 2.5c). Once the cyclonic eddy reaches the continental slope east of the Bahamas, it also remains in an approximate fixed location as its volume is slowly drained. For this experiment, however, the eddy forms a wall-jet feature that propagates along the topographic slope north of the Bahamas (Figure 2.5d). As the eddy leaks along the topographic slope, the eddy diameter and magnitude becomes progressively smaller. At the time of minimum transport (day 189), the jet feature formed an enclosed cyclonic circulation cell including the Florida Straits and the northern islands from the Bahamas archipelago (Figure 2.5d). Velocity anomalies at this time (Figure 2.5f) exhibited the signature of the wall-jet feature in the eastern part of the Straits, and were characterized by the presence of vertical shear. At day 245, a small cyclonic eddy forms within the Florida Straits (Figure 2.5g), and the velocity structure (Figure 2.5i) is fully dominated by baroclinic velocities that are closely associated with the structure from the original eddy.

On experiments SENS-E08 and SENS-E09, where eddies were initialized north of the Straits, the prescribed anticyclonic and cyclonic features propagated westward until reaching the western boundary region at the east U.S. coast after approximately 196 days of simulation (Figures 2.6a, 2.7a). On experiment SENS-E09, the minimum transport anomaly is observed at day 63, when negative SHA reach the east side of the Florida Straits (Figures 2.7b) as the cyclonic eddy propagate westward. At this time, broad negative velocity anomalies occupy the entire water column in the Straits (Figures 2.7c). As the anticyclonic and cyclonic eddies reach the western boundary in experiment SENS-E08 and SENS-E09, animations show that they rebound once, moving slightly northward, and touch the boundary again. As they touch the sloping topography for the second time, they stay in an approximate fixed location as some of their volume was slowly leaked along the continental slope (Figures 2.6d, 2.7d). Similar to experiments SENS-E01 to SENS-E07, wall-jet features are also observed to develop in experiments SENS-E08 and SENS-E09.
Figure 2.5: Fields of surface velocity (left column), SSH (center), and meridional velocity at the Florida Straits (right column) at different snapshots of model solution for experiment SENS-E07.

On these last two experiments, however, the wall-jet propagates equatorward along the east U.S. continental slope. On both experiments, the formation of these wall-jet was closely linked with the southward flowing branch of prescribed eddies: linked with the right side on the anticyclonic eddy (SENS-E08, Figures 2.6d); and with the left side on the cyclonic eddy (SENS-E09, Figures 2.7d). The velocity structure associated with these
Figure 2.6: Fields of surface velocity (left column), SSH (center), and meridional velocity at the Florida Straits (right column) at different snapshots of model solution for experiment SENS-E08.

Wall-jets (Figures 2.6i, 2.7f) shows a southward flowing jet above 200 m with velocities reaching values of $\sim 10 \text{ cm s}^{-1}$.

To further investigate the mechanisms linked with the response in the Florida Straits, the rate at which these wall-jets propagate is assessed using Hovmoller diagrams of surface velocity anomalies along the propagation path (Figure 2.8). Longitude-time Hovmoller diagrams of surface velocity anomalies along the NWPC from experiments that
Figure 2.7: Fields of surface velocity (left column), SSH (center), and meridional velocity at the Florida Straits (right column) at different snapshots of model solution for experiment SENS-E09.

Simulated anticyclonic eddies (SEN S-E01 to SENS-E06, Figure 2.9a-f) show a slanted pattern indicative of westward propagation of the wall-jet feature through the NWPC. A similar pattern is also observed for the experiment initialized with a cyclonic eddy (SENS-E07, Figure 2.9g), except that in this experiment the wall-jet feature develops and propagates along the continental slope north of the Bahamas. Westward propagation rates range between -1.6 km day$^{-1}$ in experiment SENS-E01 — anticyclonic eddy associated with 500 km wavelength and 7.5 cm SSH — to -3.6 km day$^{-1}$ in experiment SENS-E03.
Figure 2.8: Location of propagation paths (black lines) used to analyze rate of wall-jet propagation for experiments: (a) SENS-E01 to SENS-E06 along Northwest Providence Channel; (b) SENS-E07 north of the Bahamas; and (c) SENS-E08 and SENS-E09 along the east U.S. coast.

— anticyclonic eddy associated with 400 km wavelength and 15 cm SSH. On experiment SENS-E07, westward propagation of the wall-jet feature (Figure 2.9g) occurs north of the Bahamas archipelago at a rate of -3.3 km day$^{-1}$.

The average westward propagation speed of all experiments SENS-E01 to SENS-E07 is -3.2 ± 0.9 km/day. On experiments SENS-E08 and SENS-E09, Latitude-time Hovmoller diagrams of surface velocity along the U.S. continental slope (Figure 2.10) depict the southward propagation of the wall-jet. For these experiments, the wall-jet features were observed to propagate southward at a rate of approximately -2.5 ± 0.2 km day$^{-1}$.

Even though results from the SENS-E0x experiments exhibited the development of wall-jets with similar characteristics (zonal and meridional length scales), these features were observed to propagate at different phase speeds. Further analysis shows that the rate of the wall-jet propagation is correlated (r=0.8, Figure 2.11d) with the averaged barotropic velocity at the location of original formation of the wall-jet (thick black lines, Figure 2.11a-c). In addition, in all SENS-E0x experiments these wall-jet features were observed to propagate approximately along contours of $f/H$. Consequences of these results will be further assessed in the discussion session 2.3.
Results from the SENS-E0x experiments show that the response in the Florida Straits is marked by the development of volume transport anomalies that are sensitive to the
intensity of the features reaching the Bahamas, but not to their wavelength characteristics. The response is initially linked with the development of barotropic northward velocity anomalies associated with the Rossby wave field excited by the eddy, and then influenced by the development of wall-jet features directly linked with the eddy.

In summary, surface velocity variability associated with four experiments that are representative from the SENS-E0x set (Figure 2.12) indicate that the response driven by single eddy-like features in the absence of background flow is strictly confined to the region that is directly under the influence of the eddy field. In the next section, similar single-eddy experiments are developed to evaluate the response in the Florida Straits in the presence of the background flow.
2.2 With Background Flow Case Study

Experiments from set ModFC-E0x are designed to assess the impact of westward propagating signals in the Florida Straits variability in the presence of a strong background flows associated with the mean FC. To accomplish this, ten experiments (Table 2.2) are developed using single eddy-like features prescribed at different meridional locations to assess the changes in the FC transport forced by these prescribed eddy-like features. Since the previous experiments from set SENS-E0x suggested that the transport response was
Figure 2.12: Surface velocity standard deviation for experiments (a) SENS-E04 (anticyclone at 27°N), (b) SENS-E07 (cyclone at 27°N), (c) SENS-E08 (anticyclone at 28.5°N), and (d) SENS-E09 (cyclone at 28.5°N). All experiments shown here were initialized with features associated with a 500 km wavelength and 15 cm absolute SSH signal.

roughly proportional to the eddy intensity and insensitive to its wavelength, experiments are developed using eddy-like features associated with only wavelengths of 500 km, and absolute SHA of 15 cm (intensity).

2.2.1 Experiment setup

Unlike the previous experiments from set SENS-E0x, experiments from set ModFC-E0x are started using modeling outputs including well-developed ocean circulation. Initial
Table 2.2: Properties of experiments from set ModFC-E0x in terms of simulated wavelength (twice the eddy diameter) and intensity (eddy maximum SSH), initial location of eddy features, and temporal standard deviation of volume transport variability in the Florida Straits for each experiment.

<table>
<thead>
<tr>
<th>Exp. Name</th>
<th>Wavelength</th>
<th>Intensity [SSH]</th>
<th>Initial location</th>
<th>FL Straits STD.</th>
</tr>
</thead>
<tbody>
<tr>
<td>Cntr-E02</td>
<td>N/A</td>
<td>N/A</td>
<td>N/A</td>
<td>0.3 Sv</td>
</tr>
<tr>
<td>ModFC-E01</td>
<td>500 km</td>
<td>15.0 cm</td>
<td>25.5°N, 74°W</td>
<td>0.3 Sv</td>
</tr>
<tr>
<td>ModFC-E02</td>
<td>500 km</td>
<td>15.0 cm</td>
<td>26°N, 74°W</td>
<td>0.2 Sv</td>
</tr>
<tr>
<td>ModFC-E03</td>
<td>500 km</td>
<td>15.0 cm</td>
<td>26.5°N, 74°W</td>
<td>0.3 Sv</td>
</tr>
<tr>
<td>ModFC-E04</td>
<td>500 km</td>
<td>15.0 cm</td>
<td>27°N, 74°W</td>
<td>0.3 Sv</td>
</tr>
<tr>
<td>ModFC-E05</td>
<td>500 km</td>
<td>15.0 cm</td>
<td>28.5°N, 75°W</td>
<td>0.2 Sv</td>
</tr>
<tr>
<td>ModFC-E06</td>
<td>500 km</td>
<td>-15.0 cm</td>
<td>25.5°N, 74°W</td>
<td>0.2 Sv</td>
</tr>
<tr>
<td>ModFC-E07</td>
<td>500 km</td>
<td>-15.0 cm</td>
<td>26°N, 74°W</td>
<td>0.2 Sv</td>
</tr>
<tr>
<td>ModFC-E08</td>
<td>500 km</td>
<td>-15.0 cm</td>
<td>26.5°N, 74°W</td>
<td>0.2 Sv</td>
</tr>
<tr>
<td>ModFC-E09</td>
<td>500 km</td>
<td>-15.0 cm</td>
<td>27°N, 74°W</td>
<td>0.3 Sv</td>
</tr>
<tr>
<td>ModFC-E10</td>
<td>500 km</td>
<td>-15.0 cm</td>
<td>28.5°N, 75°W</td>
<td>0.3 Sv</td>
</tr>
</tbody>
</table>

and boundary conditions fields used in ModFC-E0x are derived from a six year average from a data-assimilation run of the North Atlantic Ocean [Kourafalou et al., 2016] using the HYbrid Coordinate Ocean Model (HYCOM). Boundary conditions are also set to constant values based on parameters derived from HYCOM fields, and are configured according to boundary condition options described in section 1.4.1.

Initial conditions derived from HYCOM simulations are characterized by a realistic FC, and show the characteristic filament-like warm SST signature linked with northward FC flow (Figure 2.13a). Fields derived from HYCOM also include a realistic FC and AC velocity structure (Figure 2.14). For example, the simulated FC has a well-defined velocity maximum reaching values of ~2 m s⁻¹ over the western side of the straits along the topographic slope, which is consistent with observations [Leaman and Molinari, 1987; Leaman et al., 1987; Beal et al., 2008]. The simulated AC is also consistent with the
Figure 2.13: Initial SST conditions from experiment Cntr.E02, which is derived from fields based on a six years average from HYCOM simulations for the North Atlantic Ocean.

Figure 2.14: Average (a) FC, and (b) AC meridional velocity at the Florida Straits and east of the Bahamas, respectively, from control run experiment Cntr-E02. Please note that the bottom topography at the Florida Straits in the model is $\sim 20$ shallower than the real bottom topography due to topographic smoothing to comply with ROMS requirements to reduce pressure gradient errors (please refer to section 1.4.2).

observed AC structure which exhibits a subsurface velocity maximum east of the Bahamas [Johns et al., 2008].
Experiments from set ModFC-E0x are initialized using HYCOM-derived fields perturbed with different configurations of eddy-like westward propagating signals (Table 2.2). These experiments simulate the actual FC response in the Florida Straits forced by single cyclonic or anticyclonic eddies reaching the western boundary at the Bahamas archipelago, and downstream from the Florida Straits. Eddy-like perturbations on HYCOM-derived fields are prescribed and spun-up using the method described in section 1.4.3. Experiments are then carried out for 500 days. A total of ten ModFC-E0x experiments (Figure 2.15) is carried out by varying the type of feature prescribed and the location, to assess potential changes in the FC response. One additional control run experiment (Cntr-E02) is also developed using the same initial conditions and model configuration from ModFC-E0x experiments, except that no eddy-like mesoscale features are prescribed in this simulation.

2.2.2 Results

Control Run Experiment Cntr-E02

Before proceeding with the analysis of FC experiments including single-eddy like mesoscale features (ModFC-E0x), a detailed assessment of the control run experiment Cntr-E02 is performed. This assessment aims to investigate the background levels of variability in the absence of eddy-like features explicitly prescribed in the ocean interior. It is worth emphasizing, however, that, given that the control run experiment Cntr-E02 derives from a realistic simulation from the North Atlantic Ocean [Kourafalou et al., 2016], mesoscale features originating in the interior implicit to the control run may still be expected. In addition, considering that this study employs a realistic non-linear ocean model and that the FC / Gulf Stream is a regime associated with the development of instabilities [Lee and Csanady, 1994] and strong meandering activity [Johns and Watts, 1986; Rossby et al., 2010], some degree of internal variability is expected for our control run
Figure 2.15: Initial conditions of surface velocity for (a) the control run experiment Cntr-E02, and for (b-k) experiments from set ModFC-E0x. Note that the velocity scale is saturated at 30 cm s$^{-1}$ to highlight the surface velocity structure associated with the prescribed eddy-like features in the ocean interior.
simulation. We generically define these as the intrinsic variability of the western boundary region. This type of variability is expected despite the fact that constant boundary conditions are enforced in all Cntr-E02 and ModFC-E0x experiments.

The Cntr-E02 experiment includes a mean FC (Figure 2.14a) that is characterized by a well-defined jet flowing along the topographic slope in the Florida Straits at 27°N, and a mean AC that also captures the characteristic subsurface velocity structure associated with this current (Figure 2.14b). The AC from the control run experiment carries a volume transport of 4.4 Sv, with temporal standard deviation of 2.7 Sv, which compares reasonably well with previous estimates based on observations [Lee et al., 1990, 1996; Johns et al., 2008]. In the control simulation, the FC carries a volume transport of 29.7 ± 0.6 Sv, which is slightly smaller than previous estimates of the FC transport based on in situ observations [Leaman et al., 1987; Baringer and Larsen, 2001; Beal et al., 2008; Meinen et al., 2010]. The slightly smaller FC transport obtained by our control simulation is attributed to the relatively shallower Florida Straits in the ROMS experiments, with respect to fields from HYCOM.

Time-series of FC transport for experiment Cntr-E02 (Figure 2.16) indicate that about 25% of the variance is linked with variability within the frequency band that is the focus of this study (73-525 days period, STD = 0.3 Sv), while the remaining is dominated by high frequency events. Filtered time-series of FC transport for periods larger than 73 days (black line, Figure 2.16) indicate changes in transport have an absolute range of ∼1 Sv. This is important given that part of the intrinsic FC variability from the control experiment is within the frequency-band that is the focus of this study (73 - 525 days). Below, potential mechanisms driving changes in the FC transport in the control run are addressed. From now on, unless otherwise stated, time-series analysis reported in this result section are based on time-series low-passed filtered for 73 days.
Figure 2.16: FC transport time-series (thin gray line) from the control run experiment Cntr-E02. The black line shows values low pass filtered at 73 days.

Standard deviation of meridional velocity in the Florida Straits (Figure 2.17a) indicates that most of the intrinsic FC variability is linked with velocity changes in the western side of the Florida Straits, along the sloping topography. Estimates of the velocity standard deviation for both the high-frequency (<73 days periods, Figure 2.17b) and transient (73-525 days, Figure 2.17c) components are also linked with similar variability along the western side of the Florida Straits. In fact, during maximum and minimum transport in the Florida Straits (Figure 2.18), velocity anomalies in the Florida Straits exhibit maximum absolute values over the continental slope. These results suggest that similar mechanisms may control both the high frequency and transient two components of the intrinsic FC variability in the control run Cntr-E02 experiment.

Results from Cntr-E02 experiment suggest that most of the variability in the control run is linked with the intense Gulf Stream meandering that occurs downstream of the Straits approximately between 28°N-36°N. For example, surface velocity fields for selected snapshots of model solution (Figure 2.19) shows that meandering can propagate anomalies along the Gulf Stream and often shed eddies (Figure 2.19b), that may recirculate southwestward and later be reincorporated into the main jet stream (Figure 2.19). Typical eddies shed by the meander at 32°N have diameters of ~100 km diameters, and surface velocity reaching values of 50 cm s⁻¹, and are generally smaller and more
Figure 2.17: Standard deviation of meridional velocity in the Florida Straits for experiment Cntr-E02: (a) total standard deviation; (b) high-frequency component (<73 days period); and (c) transient seasonal component (73-525 days period).

Figure 2.18: Velocity anomalies in the Florida Straits referenced during (a) maximum, and (b) minimum transport anomalies for experiment Cntr-E02.

intense than the eddy-like mesoscale features that are explicitly prescribed in experiments ModFC-E0x (Figure 2.15).

From now on, snapshots of sea height anomaly referenced to the average dynamic topography from the control run (hereafter referred as SHA for simplicity) are used to qualitatively assess the evolution of Cntr-E02 and ModFC-E0x experiments. Analysis of
SHA fields for the control run (Figure 2.20) indicates that, as meandering develops, SHA propagate northward along the Gulf Stream and generally intensify. Approximately at 35°N, as these SHA signals approach Cape Hatteras, they often interact with the shelf topography there (Figure 2.20b,e), causing broad signals of the same sign to propagate southward along the U.S. coast, reaching the Florida Straits few days afterwards (Figure 2.20c,f). These results suggest that coastally trapped signals may provide the mechanism linking FC and Gulf Stream meandering with changes in the variability in the Florida Straits. In fact, the propagation of signals along the coastal waveguide may be one of the main processes accounting for the the meridional velocity variability in the Florida Straits described above for the control run experiments (Figure 2.17).
Figure 2.20: Maps of SHA for control run experiment Cntr-E02 at different snapshots of model solution. Emphasis is given to specific positive (magenta circle) or negative (green circles) SHA signals that are advected by the background circulation, that may generate coastally trapped signals as they reach shallow waters north of 35°N.

To further verify the generation of coastally trapped signals, latitude-time Hovmoller diagrams of SHA for experiment Cntr-E02 are analyzed for changes along the coastal waveguide (Figure 2.21a). Latitude-time Hovmoller diagram of SHA along the coastal waveguide (Figure 2.21b) shows the presence of slanted signals indicative of southward propagation. Analysis of the slope associated with the dominant signals indicates a southward propagating rate of $82.4 \pm 13.4 \text{ km day}^{-1}$, which is consistent with the propagation of first-mode baroclinic coastally trapped waves. Careful examination of the coastal latitude-time Hovmoller diagram of SHA shows that dominant signals occur at periods faster than 73 days, and also suggest that signals in lower frequency bands (e.g., in the transient seasonal component) may also be transmitted along the coastal waveguide. Dominant signals...
Figure 2.21: (a) Location of grid-points used to evaluate sea-level variability along the east US coast. (b) Latitude-time Hovmoller diagram for SHA along the east coast of US during experiment Cntr-E02. The black box in panel (b) emphasizes features emphasized in the text.

are mostly generated in the proximity of Cape Hatteras and take approximately 21 days to reach the Florida Straits, while secondary signals are also observed to develop midway. Secondary signals are likely associated with lateral meandering along the Georgia and Florida coasts. For example, between days 217 and 252 (green square, Figure 2.22), lateral meandering at \( \sim 30^\circ \text{N} \) generates a positive SHA signal over the shelf, triggering a southward propagating signal that reaches the Florida Straits approximately 7 days after. It is worth pointing out that all simulations are developed here in the absence of wind forcing. Hence, generation of coastally trapped signals is only possible through shelf interactions with offshore features.

So far, results suggest that the intrinsic FC variability is largely driven by meandering / instabilities in locations downstream from the Florida Straits, which may later reach the Florida Current through southward propagating coastally trapped signals. Here, an
additional analysis is carried out to assess potential other sources of FC variability in
the control run that originate in the open ocean. To accomplish this, Longitude-time
Hovmoller diagrams of SHA along 27°N (Figure 2.23) are analyzed. This diagram shows
westward propagating SHA signals with amplitude smaller than 10 cm, which indicates
that, despite the fact that eddy-like features are not explicitly prescribed in the control
run, the realistic model setup still allows the generation of open ocean variability with
characteristics similar to the signals studied here.

To quantify the relative contribution from each of these two mechanisms for driving
the intrinsic FC variability in the control simulation, the following approach is developed:
(step-1) coastal SHA time-series are retrieved at both sides of the Florida Straits at 27°N
as proxies for the FC variability induced by coastally trapped signals traveling along the
U.S. coast, and by the signals originating in the open North Atlantic ocean east of the
Bahamas; (step-2) time-series of coastal SHA and of FC transport are normalized by re-
moving their mean values and dividing by their standard deviation; (step-3) a multi-linear
regression analysis (formula below) is employed using a bootstrap approach [Johnson,
2001], in which the normalized SHA time-series are used as predictors (X matrix) for the
FC transport (Y vector). The multi-linear regression analysis consists of solving:

\[
\begin{bmatrix}
A_{FL} \\
A_{BHS}
\end{bmatrix} = (X^T \times X)^{-1} \times X^T \times Y
\]

(2.6)

where,

\[
X = \begin{bmatrix}
SHA_{FL_{t=1}} & SHA_{BHS_{t=1}} \\
\vdots & \vdots \\
SHA_{FL_{t=n}} & SHA_{BHS_{t=n}}
\end{bmatrix}, \text{ and } Y = \begin{bmatrix}
FC_{t=1} \\
\vdots \\
FC_{t=n}
\end{bmatrix}
\]

(2.7)

Step-3 is applied employing the normalized time-series, and also using the actual SHA
Figure 2.22: Snapshots of model solution for SHA at different times for control run experiment Cntr-E02. Green squares indicates the region where lateral meandering at $\sim 30^\circ$N generates positive southward propagating SHA signals that reaches the Florida Straits approximately 7 days after.

and transport time-series, resulting in two sets of slope coefficients ($A_{FL}$ and $A_{BHS}$): one set of normalized coefficients, and one set with physical units of Sv cm$^{-1}$. The overall correlation coefficient ($R_{FC}$) from the multi-linear regression analysis is also reported.

Application of this approach to the control run experiment Cntr-E02 (Table 2.3) shows that approximately 50% ($R_{FC}$=0.70, $p<0.05$) of the variance of the FC transport in the control run can be explained by coastal SHA changes from both sides of the Florida Straits. Slope coefficients shows, for example, that a 1 cm sea-level rise in the Bahamas side is generally associated with a 0.4 Sv increase in the FC transport, while a similar sea-level increase in the Florida side is only associated with a -0.14 Sv decrease in the transport. At first glance, analysis of the physical slope coefficients may suggest that
SHA changes in the Bahamas side dominate the FC variability in the control run. However, the physical slope coefficients are largely biased by the unequal SHA variability across the Florida Straits: SHA variability is almost three times larger at the Florida side of the Straits. The slope coefficients based on normalized time-series provide an unbiased estimate of the contribution of each side to the total FC variability. Analysis of the normalized coefficients indicates that similar values for both sides of the Florida Straits, with $A_{FL}$ equal to -0.49, and $A_{BHS}$ equal to 0.52. This indicates that both southward propagating coastally trapped waves and the westward propagating signals originating in the open ocean may contribute similar amounts with the intrinsic FC variability in the control run. Below, a similar analysis will be employed to track mechanisms driving FC variability on the ModFC-E0x experiments. From now on, focus is given to the analysis of the slope coefficients based on normalized time-series.

*Experiments from set ModFC-E0x*

Experiments from set ModFC-E0x were initialized by the inclusion of single eddy-like mesoscale features in the control run fields from experiment Cntr-E02, and spun-up using ROMS in the diagnostic mode until the stable development of momentum fields. The inclusion of eddy-like features was implemented by prescribing localized eddy-like
Table 2.3: Results from the multi-linear regression analysis performed for the control run and experiments ModFC-E0x. Slope coefficients \( A_{xx} \) labeled as “FL” indicate coefficients from the west or Florida side of the Straits, while coefficients labeled as “BHS” indicate coefficients from the east or Bahamas side of the Straits. Standard deviation of SHA in cm for each side of the Florida Straits is shown in the first two columns. Physical coefficients have units of Sv cm\(^{-1}\), while normalized coefficients are non-dimensional. The asterisk indicates coefficients that are not significant at 95% confidence level.

<table>
<thead>
<tr>
<th>Exp. Name</th>
<th>std FL</th>
<th>std BHS</th>
<th>Physical</th>
<th>Normalized</th>
<th>( R_{FC} )</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td>( A_{FL} )</td>
<td>( A_{BHS} )</td>
<td>( A_{FL} )</td>
</tr>
<tr>
<td>Cntr-E02</td>
<td>2.0</td>
<td>0.8</td>
<td>-0.14</td>
<td>0.40</td>
<td>-0.48</td>
</tr>
<tr>
<td>ModFC-E01</td>
<td>2.2</td>
<td>0.8</td>
<td>-0.16</td>
<td>0.44</td>
<td>-0.59</td>
</tr>
<tr>
<td>ModFC-E02</td>
<td>2.0</td>
<td>0.8</td>
<td>-0.14</td>
<td>0.42</td>
<td>-0.51</td>
</tr>
<tr>
<td>ModFC-E03</td>
<td>2.3</td>
<td>0.8</td>
<td>-0.22</td>
<td>0.28</td>
<td>-0.76</td>
</tr>
<tr>
<td>ModFC-E04</td>
<td>2.2</td>
<td>0.8</td>
<td>-0.20</td>
<td>0.42</td>
<td>-0.72</td>
</tr>
<tr>
<td>ModFC-E05</td>
<td>2.1</td>
<td>0.7</td>
<td>-0.17</td>
<td>0.25</td>
<td>-0.60</td>
</tr>
<tr>
<td>ModFC-E06</td>
<td>2.1</td>
<td>0.6</td>
<td>-0.17</td>
<td>0.30</td>
<td>-0.65</td>
</tr>
<tr>
<td>ModFC-E07</td>
<td>1.8</td>
<td>0.8</td>
<td>-0.16</td>
<td>0.43</td>
<td>-0.45</td>
</tr>
<tr>
<td>ModFC-E08</td>
<td>2.1</td>
<td>0.7</td>
<td>-0.12</td>
<td>0.46</td>
<td>-0.44</td>
</tr>
<tr>
<td>ModFC-E09</td>
<td>2.1</td>
<td>0.7</td>
<td>-0.20</td>
<td>0.34</td>
<td>-0.70</td>
</tr>
<tr>
<td>ModFC-E10</td>
<td>2.7</td>
<td>0.9</td>
<td>-0.19</td>
<td>0.37</td>
<td>-0.65</td>
</tr>
</tbody>
</table>
perturbations in the pycnocline depth, which implies that areas outside of the perturbed region remained essentially the same as the unperturbed control run (Figure 2.15). It is estimated here that the inclusion of the additional single eddy-like mesoscale features in experiments ModFC-E0x caused a modest increase in the kinetic energy of numerical simulations. The domain integrated kinetic energy per unit mass was $165.4 \times 10^3 \text{ m}^2 \text{ s}^{-2}$ during initialization for the control run experiment Cntr-E02, while the same quantity for ModFC-E0x experiments had average values of $[165.7 \pm 0.2] \times 10^3 \text{ m}^2 \text{ s}^{-2}$. This resulted in an approximate increase of only $\sim 0.2 \%$ in the domain integrated kinetic energy per unit of mass for ModFC-E0x experiments. Given the modest increase in energy levels, one can hardly expect an abrupt increase in the overall variability in ModFC-E0x simulations with respect to the control run experiment. In fact, time-series of volume transports in the Florida Straits (Figure 2.24a) show that the overall variability remained consistent between the control and ModFC-E0x experiments (Table 2.2). For example, the mean FC transport from ModFC-E0x experiments, at $29.7 \pm 0.3 \text{ Sv}$, coincides with the mean FC transport from the control run Cntr-E02 ($29.7 \pm 0.3 \text{ Sv}$). It is interesting to note that, even though the Cntr-E02 and ModFC-E0x FC time-series are consistent in terms of mean and overall variability, these time-series seem to diverge in terms of timing and phase. Phase discrepancies are more easily identified analyzing the differences between FC time-series derived from ModFC-E0x experiments and from the control run Cntr-E02 (Figure 2.24b). These differences show that, while phase discrepancies result in marginal differences ($\sim 0.3 \text{ Sv}$) in the first 100 days, at $\sim 230$ days a peak of 1 Sv difference indicates the first main change in phase. At this time, the FC from the control experiment Cntr-E02 shows a decline in transport (Figure 2.24a), reaching values close to 29.2 Sv, while for most ModFC-E0x experiments the FC continues flowing with transports of $\sim 30 \text{ Sv}$ or above. For the ModFC-E0x experiments, the decline in the FC transport is postponed with respect to the control run, and lower values happened between
Figure 2.24: (a) FC transport time-series at the Florida Straits from numerical experiments ModFC-E0x. (b) Transport anomalies from ModFC-E0x experiments with respect to the Control run Cntr-E02.

simulation time 260-420 days for the different experiments. FC transport differences with respect to the control run also suggest that changes in the variability occur within dominant time-scales between 100-200 days. It is worth pointing that the ensemble of FC time-series from ModFC-E0x time-series shows smaller levels of variability before day 200, whereas after this day, the FC time-series from each simulation behaves independently, and the levels of ensemble variability increase. These results indicate that the inclusion of eddy-like perturbations in the ModFC-E0x experiments caused perturbations in the phase of the FC variability in the Florida Straits. Note that shifts in phase described here are consistent with variability in the FC within the 73-525 transient seasonal band that is the focus of this study.

Next, the analysis focuses on investigating the mechanisms driving the changes in the seasonal phase of the FC variability in the different ModFC-E0x experiments. From now on, when a more detailed analysis is required, results from selected ModFC-E0x
experiments are presented to illustrate a general response observed. This approach is adopted in order to optimize the presentation of results and to avoid redundant information, given that the experiments have sometimes similar initial conditions. Results from four experiments will be used more often: from one experiment initialized with an anticyclonic eddy at 27°N (ModFC-E04), from one experiment initialized with a cyclonic eddy at 27°N (ModFC-E09), from one experiment initialized with an anticyclonic eddy at 28.5°N (ModFC-E05), and from one experiment initialized with a cyclonic eddy at 28.5°N (ModFC-E10).

First, the structure of velocity anomalies at the Florida Straits with respect to the control run Cntr-E02 is evaluated at times of maximum transport difference. Results from experiments ModFC-E04, E05, E09, and E10 (Figure 2.25) show that the initial transport differences develop between days 90-160 and reach values ranging between 0.2-0.3 Sv with respect to the control run experiment. Differences were generally negative, with the exception of results from ModFC-E10 (Figure 2.25j), which reached positive values of ∼0.2 Sv at day 119. At the time of this initial transport difference for this experiment, velocity differences were characterized by broad positive velocity differences less than 1 cm s\(^{-1}\) occupying the entire Florida Straits (Figure 2.25k). For the other three experiments, the initial transport differences maxima were accompanied by negative velocity anomalies occupying the western side of the Florida Straits (Figure 2.25b,e,h). The maximum transport differences between experiments ModFC-E04, E05, E09, and E10 and the control run Cntr-E02 occurred at day 238 (Figure 2.25a,d,g,j). At that day, transport differences reached values of ∼1 Sv for most experiments. Velocity anomalies were characterized by positive anomalies reaching values larger than 20 cm s\(^{-1}\) confined to the western side of the Florida Straits. The structure of velocity anomalies at the time of maximum transport differences is consistent with the overall variability.
Figure 2.25: Time-series of volume transport anomaly at the Florida Straits (left column), and meridional velocity anomalies (central and right column) with respect to the control run experiment Cntr-E02 for experiments ModFC-E04, ModFC-E05, ModFC-E09, and ModFC-E10.

of the velocity field in this location (Figure 2.17), suggesting the importance of coastally trapped mechanisms as discussed above for control run experiment Cntr-E02.
As an attempt to assess the spatial distribution of the response driven by the inclusion of eddy-like features in the ModFC-E0x experiments, the standard deviation of surface velocity differences between the control simulation Cntr-E02 and the ModFC-E0x experiments for each snapshot of model solution are evaluated. The standard deviation displayed in Figure 2.26 summarizes, in terms of root mean squared differences, how two modeling experiments with slightly different initial conditions (Cntr-E02 and ModFC-E0x) may evolve into parallel realities. The slightly different initial conditions were accomplished by prescribing the eddy-like features in the ModFC-E0x experiments. In practice, ocean processes forced by the eddy may drive the velocity field from ModFC-E0x experiments at different locations to diverge from the velocity field from the control run Cntr-E02. In other words, the model solution from ModFC-E0x experiments at grid points that are directly or indirectly affected by anomalies caused by the eddy will eventually diverge from the solution of Cntr-E02. Results from experiments ModFC-E04, E05, E09, and E10 indicate that velocity discrepancies with respect to the control experiment reach a wide geographical extent (Figure 2.26). In addition, standard deviation very commonly reach values larger than 10 cm s\(^{-1}\), which was the magnitude of surface velocity variability response to eddies in the absence of background flow (Figure 2.12). Considering that the eddy-like features were prescribed at 27°N and 28.5°N, the large geographical extent in the “response” and the large values of standard deviation (considering the anomaly prescribed), suggests the active interaction of the prescribed perturbations with background circulation. Signals are observed over most of the western boundary region, and even eastward of their original location. The latter suggests a possible interaction with background eastward propagating short Rossby waves, however investigating such a mechanism is beyond the scope of this work.

The evaluation of snapshots of surface velocity solution for experiment ModFC-E04 (Figure 2.27a,b,c), indicates that, indeed, the prescribed feature interacts actively with the
Figure 2.26: Standard deviation from surface velocity anomalies referenced to control run experiment Cntr-E02 for experiments (a) ModFC-E04 (anticyclone at 27°N), (b) ModFC-E09 (cyclone at 27°N), (c) ModFC-E05 (anticyclone at 28.5°N), and (d) ModFC-E10 (cyclone at 28.5°N). All experiments shown here were initialized with features associated with a 500 km wavelength and 15 cm absolute SSH signal.

background circulation. For this experiment, the anticyclonic eddy propagates slightly westward as its volume is slowly drained by the AC through its northward (western) branch. The entire process takes approximately $\sim$150 days until the entire volume of the eddy has been drained by the current (Figure 2.27c). A similar process is also observed for experiment ModFC-E09, which was initialized with a cyclonic eddy at 27°N (Figure 2.27d,e,f). However, in this experiment, the eddy is mostly drained through its southward flowing (western) branch. On the experiments initialized with features downstream of
the Florida Straits at 28.5°N, features were also advected and drained by the background flow, which at this time was mostly associated with the offshore Gulf Stream circulation (Figure 2.27g,h,i,j,k,l).

Overall, the main signals associated with the prescribed eddies appear to be advected northward along with the main currents that compose the subtropical western boundary circulation in the North Atlantic ocean. In this following analysis, an attempt is made to track SHA that were originally linked with the eddy. The goal of this analysis is to identify the fate of the main signals associated with the prescribed features, and where and how they may drive changes in the FC seasonal phase. The leading edge of SHA signals associated with the eddy were visually followed at each snapshot of model solution for each experiment. As the nonlinear simulation evolved, signals from the eddy interacted and were eventually incorporated into other features. In these cases, the method would continue following the amplified signal from the merged feature; the goal was to gain some insight on the potential fate of those signals. Figure 2.28 shows results from this method applied for experiments ModFC-E04, which provides a good overview of the pattern generally observed for the other ModFC-E0x experiments. For example, it shows how positive SHA signals associated with the original prescribed anticyclonic eddy at 27°N propagate northward while the eddy is drained by the AC. At a later time, these positive SHA signals merge with other positive signals that originated in the FC, and are then advected northward towards Cape Hatteras.

Analysis of individual tracks from ModFC-E0x (Figure 2.29) indicate that all experiments behaved similarly, and that signals are largely advected northward by the background circulation towards Cape Hatteras. This is not surprising given that experiments have similar initial conditions and that prescribed perturbations correspond to a small percentage (~0.2%) of the total kinetic energy, indicating that the experiments are largely dominated by the background circulation. In fact, analysis of satellite altimetry data indi-
Figure 2.27: Surface velocity fields of snapshots of model solution for experiments ModFC-E04 (anticyclone at 27°N), ModFC-E09 (cyclone at 27°N), ModFC-E05 (anticyclone at 28.5°N), and ModFC-E10 (cyclone at 28.5°N).

cates that SHA signals may also behave similarly in the real ocean. For example, application of a similar SHA tracking method for satellite altimetry data shows that positive and negative features located north of the Bahamas on September 2011 (Figure 2.30) were
Figure 2.28: Fields of SHA from experiment ModFC-E04 at different times of model solution. The green lines with “x” marker indicates the track followed by traceable SHA signals linked with the original eddy-like features prescribed in this experiments.

also advected northward along the known paths of the AC and Gulf Stream, until the negative signals reached Cape Hatteras and the positive signals detached from the coast. Indeed, signals observed on December 1997 also showed a similar tendency to propagate northward along with other surrounding signals. It is interesting to note that it took
Figure 2.29: Track of SHA signals that were associated with the original eddy-like features prescribed in ModFC-E0x experiments. The color code indicate individual ModFC experiments as listed in Figure 2.24

approximately 100 days for signals to reach the proximity of Cape Hatteras, similar to the time-frame for signals simulated observed in this study. In addition, it was observed that the positive SHA signals showed a tendency to recirculate towards the interior of the basin and then southwestward for these two occasions evaluated (not shown for the first one), which is similar to the behavior previously described for anticyclonic features that were spontaneously shed by the meander at 32°N in the control run experiment.

In order to have a more quantitative picture of processes linking the eddy induced phase changes in the FC variability, an analysis based on cumulative transport along four sections at different locations at the U.S. coast (Figure 2.31) is carried out. The analysis focuses on Longitude-time diagrams of cumulative transport differences between the ModFC-E0x experiments and the control run. In essence, these differences capture the terms linked with the prescribed features (e.g. $u' \frac{\partial u'}{\partial x}$), but also the non-linear advection
linked with the background circulation field (e.g. \( u' \frac{\partial u}{\partial x} \), and \( u \frac{\partial u'}{\partial x} \)). Therefore, the amplification of original eddy anomalies resulting from non-linear interactions is an expected outcome.

Analysis of cumulative transport differences between experiment ModFC-E04 and the control run Cntr-E02 at 28°N (Figure 2.32) exhibits positive transport differences with values as large as 10 Sv (saturated at 5 Sv in this figure) during the first 50 days centered at the initial time and 75°W, which is consistent with the original anticyclonic feature prescribed for ModFC-E04 at this location. The slanted pattern of positive transport anomalies with values of \( \sim 2 \) Sv at 28°N is consistent with positive SHA signals being drained by the AC (Figure 2.28b). By day 70, signals originating in the eddy reach
30°N (green “x” at 30°N, Figure 2.32). At this time, the positive SHA signals from the eddy interact with the background circulation, reinforcing the offshore branch of the circulation associated with the Gulf Stream by 5 Sv. Once the SHA are incorporated in the Gulf Stream circulation, the transport differences with respect to the control run continue amplifying as they propagate northward. This indicates that the signals associated with the prescribed eddy can work as perturbations that may cause the Gulf Stream circulation to diverge from the control run. For example, by day ∼ 150, SHA signals associated with the eddy reach 32°N, causing transport differences with respect to the control experiment as large as 10 Sv (saturated at 5 Sv in Figure 2.32). In addition, positive transport differences at this latitude are also observed to move towards the open ocean, suggesting a change in the meandering variability with respect to the control run. It is interesting that, as the SHA signals continue moving northward, at 34°N they are advected towards the coastal flank of the Gulf Stream circulation, causing negative transport differences with
Figure 2.32: Longitude-time Hovmoller diagram of cumulative transport from west to east for control experiment Cntr-E02 (left column), for experiment ModFC-E04 (central column), and the difference (right column), at four latitudes. The green “x” on the difference plot shows the projection of the traceable SHA signals linked with the prescribed eddy from ModFC-E04 experiment (see Figure 2.28).

respect to the control run experiment around day 180. At 34°N transport differences are observed to reach shallow isobaths, and broad negative signals reach the coast around day 220, which is approximately 20 days before the peak difference between the FC transport in experiment ModFC-E04 and the control run Cntr-E02.

The analysis based on cumulative transport differences described above suggests that advection of anomalies originating in the prescribed eddy-like feature from ModFC-E0x experiments are possibly linked with a modulation in the meandering characteristics of the Gulf Stream. To further verify this, the meandering variability from ModFC-E0x experiments is compared with the equivalent variability from the control experiment.
Meandering time-series are computed at 28°N, 30°N, 32°N, and 34°N by following the longitude of maximum jet intensity at each of these latitudes (Figure 2.33). Analysis indicates that the meandering variability in ModFC-E0x experiments at 34°N (Figure 2.33a) starts to diverge from the control run around day 70, which coincides with the timing that signals linked with the eddy reach the offshore circulation associated with the Gulf Stream at ~30°N (green “x” at 30°N, Figure 2.32). At other latitudes, the ensemble meandering variability displayed by ModFC-E0x experiments shows largest discrepancies with respect to the control run only after day 200 (Figure 2.33b-d), which coincides with the approximate timing that maximum transport differences were also observed (Figure 2.24b). These results also indicate that, once the initial discrepancies develop, the meandering variability changes remarkably from the control run variability and also among the ModFC-E0x experiments. Hence, it is reasonable to consider that the signal from prescribed eddy-like features in the ModFC-E0x experiments caused adjustments in the offshore circulation, as described above, that perturbed the Gulf Stream meandering variability. In fact, given the energetic and highly nonlinear nature of the Gulf Stream, these results indicate that SHA signals from the eddies may have first indirectly changed the stochastic characteristic of the system, which may have lead to the changes in the meandering pattern. In other words, the addition of SHA signals linked with prescribed eddies that were advected by AC into the Gulf Stream may be analogous to the “butterfly effect”, where a small perturbation in a nonlinear system may lead to a change of the state of that system.

So far, results from ModFC-E0x experiments showed that SHA signals linked with prescribed eddy-like features were generally advected along with the background circulation towards Cape Hatteras, and that they could potentially drive changes in the meandering variability at 34°N. The analysis now focuses on showing that changes in the variability downstream at Cape Hatteras can largely explain the changes in phase of the
Figure 2.33: Time-series of jet meandering at (a) 34°N, (b) 32°N, (c) 30°N, and (d) 28°N for the different ModFC-E0x, in comparison to the control run experiment Cntr-E02. Lines are color-coded according to the legend shown in Figure 2.24.

FC transport observed for the ModFC-E0x experiments compared to the control run. To accomplish this time-series of coastal SHA are evaluated for ModFC-E04, E05, E09, and E10 experiments (Figure 2.34). Latitude-time Hovmoller diagrams of coastal SHA for ModFC-E0x exhibits the signature of signals similar to the previously described coastally trapped waves described for experiment Cntr-E02. Detailed analysis of coastal SHAs during experiment ModFC-E04 (Figure 2.34c) indicates that the phase and sign of coastal signals changed remarkably when compared to results from the control run Cntr-E02.
(Figure 2.34b). For example, between 200-300 days of simulation, mostly positive SHA signals were observed in the Cntr-E02 between 26°N-28°N. (black square, Figure 2.34b). During the same period, mostly negative SHA signals were observed for ModFC-E04 experiment (black square, Figure 2.34c). In fact, mostly negative signals were also observed for other ModFC-E0x experiments (Figure 2.34d,e,f). These differences in the coastal sea-level with respect to the control run at 27°N largely explain the FC phase changes in ModFC-E0x that resulted in positive transport differences of \( \sim 1 \) Sv at day 238 (Figure 2.24b). Therefore, these results indicate that advection of SHA signals originating in the open ocean to locations downstream from the Florida Straits (e.g. Cape Hatteras) may feedback to the local variability in the Florida Straits through coastally trapped mechanisms.

In order to provide one first quantitative assessment of this mechanism, an analysis of the spatial distribution of correlation coefficients between FC transport differences and SHA differences is evaluated. Differences are calculated by subtracting results from ModFC-E0x from results the control run Cntr-E02. For example, FC transport differences are displayed in Figure 2.24b. Significant correlation coefficients from this analysis indicate the location of areas where the inclusion of eddy features on ModFC-E0x experiments caused SHA variability to diverge from the control run, leading to changes in the FC variability. Results from this analysis for experiments ModFC-E04 and ModFC-E09 (Figure 2.35) exhibits strong negative and positive correlation coefficients on the western and eastern sides of the Florida Straits, respectively, with absolute values above 0.8. The strong correlation coefficients indicate that over 60% of the variance of changes in the FC variability is explained by SHA differences between ModFC-E0x experiments and the control run Cntr-E02. The very high correlation coefficients observed at the Florida Straits are expected, as the geostrophic dynamics of the FC dictates that the SHA gradient across the Straits should be directly proportional to the transport. The spatial distribution
Figure 2.34: (a) Location of grid-points used to evaluate sea-level variability along the east US coast. Latitude-time Hovmoller diagram for SHA along the east coast of US during experiment (b) Cntr-E02, (c) ModFC-E04, (d) ModFC-E05, (e) ModFC-E09, and (f) ModFC-E10.

The analysis of correlation coefficients also indicates that changes in FC variability in ModFC-E0x were largely linked with changes in the coastal SHA variability, and along the continental topographic slope along U.S. coast. This result suggests that coastal signals may propagate through the coastal waveguide both through shallow waters and along the continental shelf break. In fact, analysis of similar correlation coefficients between the experiment ModFC-E04 and the control run for -21 days lag (Figure 2.35c) shows that the nega-
Figure 2.35: Spatial distribution of correlation coefficients between SHA differences and FC transport differences evaluated as: (a) results from ModFC-E04 minus results from Cntr-E02 at 0 lag; (b) results from ModFC-E09 minus results from Cntr-E02 at 0 lag; (c) results from ModFC-E04 minus results from Cntr-E02 at -21 days lag. Correlation coefficients that are not significant at the 95% confidence level are masked.

tive correlation pattern along the U.S. at zero lag can be traced to locations north of the Florida Straits: (i) along the topographic slope near the Charleston Bump at $\sim 32^\circ$N; and (ii) along the coast in the proximity of Cape Hatteras. In addition, strong positive correlation coefficients are observed on the offshore flank of the Gulf Stream, which is close to the path of propagation followed by SHA signals (Figure 2.29) as the were advected by the background circulation. Therefore, the correlation developed here analysis provides further evidence to support the hypothesis that changes in the FC variability in ModFC-E0x were indirectly forced by the eddy-like features included in the ocean interior, which perturbed the intrinsic dynamics of the system, driving the observed phase changes in the FC transport.
To further quantify the impact of modulation of coastally-trapped signals in the FC transport variability changes, a modified multi-linear regression approach (originally presented as equation 2.6) is employed here. The application of the original multi-linear regression approach to ModFC-E0x experiments (Table 2.3) confirms results from the control run that coastal SHA time-series at the Florida Straits explain about 55% of the variance (average $R_{FC} = 0.74 \pm 0.07$, Table 2.3). In the ModFC-E0x experiments, however, SHA time-series from the U.S. coast have a slightly stronger influence on the overall FC variability than the time-series from the Bahamas, as values from $A_{FL}$ (average $A_{FL} = -0.60 \pm 0.11$) are approximately 20% larger in magnitude than $A_{BHS}$ (average $A_{BHS} = 0.48 \pm 0.11$). The modified multi-linear regression approach consist of replacing the original FC transport and SHA time-series for each experiment, by time-series of SHA and transport differences with respect to the control run. More specifically, the multi-linear regression approach differs in the following: the time-series employed in step-1 are of SHA differences between ModFC-E0x and the control run Cntr-E02, for both sides of the Florida Straits; similarly, in step-3, time-series of FC transport differences between results from ModFC-E0x simulations and the the control run (see. Figure 2.24b) are employed; results from this slightly modified approach are also reported in terms of slope coefficients ($A_{FC,diff}$, $A_{BHS,diff}$) and overall correlation coefficient ($R_{diff}$) (Table 2.4). This approach allows us to more quantitatively assess the influence of the mechanisms described above for driving the reported changes in the FC variability in ModFC-E0x experiments with respect to the control run. Results from this modified approach show that time-series of coastal SHA differences at the Florida Straits with respect to the control run may similarly explain about 53% (average $R_{diff} = 0.73 \pm 0.07$) of FC transport differences. Interestingly, analysis slope coefficients reveal that the impact of coastal SHA differences at the FL side of the Florida Straits ($A_{FL,diff} = -0.64 \pm 0.05$) on FC transport differences is twofold when compared to SHA differences ($A_{BHS,diff} = 0.31 \pm 0.07$). In
Table 2.4: Same as Table 2.3, except that values reported here are computed using time-series of SHA and transport differences from ModFC-E0x with respect to the control run experiment Cntr-E02.

<table>
<thead>
<tr>
<th>Exp. Name</th>
<th>Physical</th>
<th>Normalized</th>
<th>Physical</th>
<th>Normalized</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>mean FL</td>
<td>mean BHS</td>
<td>$A_{FL:diff}$</td>
<td>$A_{BHS:diff}$</td>
</tr>
<tr>
<td>ModFC-E01</td>
<td>2.2</td>
<td>0.7</td>
<td>-0.22</td>
<td>0.26*</td>
</tr>
<tr>
<td>ModFC-E02</td>
<td>2.5</td>
<td>0.6</td>
<td>-0.18</td>
<td>0.36*</td>
</tr>
<tr>
<td>ModFC-E03</td>
<td>3.0</td>
<td>0.6</td>
<td>-0.19</td>
<td>0.48</td>
</tr>
<tr>
<td>ModFC-E04</td>
<td>2.7</td>
<td>0.8</td>
<td>-0.18</td>
<td>0.32*</td>
</tr>
<tr>
<td>ModFC-E05</td>
<td>2.6</td>
<td>0.8</td>
<td>-0.15</td>
<td>0.44</td>
</tr>
<tr>
<td>ModFC-E06</td>
<td>2.8</td>
<td>0.8</td>
<td>-0.18</td>
<td>0.29*</td>
</tr>
<tr>
<td>ModFC-E07</td>
<td>2.0</td>
<td>0.7</td>
<td>-0.23</td>
<td>0.36</td>
</tr>
<tr>
<td>ModFC-E08</td>
<td>2.5</td>
<td>0.7</td>
<td>-0.20</td>
<td>0.21*</td>
</tr>
<tr>
<td>ModFC-E09</td>
<td>2.5</td>
<td>0.7</td>
<td>-0.20</td>
<td>0.34</td>
</tr>
<tr>
<td>ModFC-E10</td>
<td>2.9</td>
<td>0.8</td>
<td>-0.23</td>
<td>0.46</td>
</tr>
</tbody>
</table>

In other words, time-series of SHA differences at the U.S. coast have a impact twice as large SHA differences at the Bahamas for capturing FC transport differences. These results indicate that FC transport differences between ModFC-E0x experiments and the control-run Cntr-E02 are dominated by SHA differences along the coastal waveguide.

Therefore, results from ModFC-E0x experiments suggest that the mechanism by which single eddy-like mesoscale features originating in the ocean interior may modulate the FC transport variability includes: (1) advection of anomalies originally associated with these features by the background circulation towards downstream locations along the Gulf Stream system; (2) interaction with the background flow; (3) changes in the the meandering variability; and (4) are finally transmitted to the Florida Straits by modulations in the SHA variability along the coastal waveguide.
2.3 Discussion

Experiments from sets SENS-E0x and ModFC-E0x provided useful information on the variability forced at the Florida Straits by single eddy-like westward propagating signals originating in the open North Atlantic ocean for cases “with no background flow”, and “with the background flow”, respectively. Experiments were initially developed in the absence of background flows (SENS-E0x experiments, Figure 2.1) to more easily identify the potential mechanisms linking the open ocean variability with the response at the Florida Straits. For the experiments with the background (ModFC-E0x experiments, Figure 2.15), a more realistic simulation was employed to assess the dynamic mechanisms that are more likely to act in the real ocean as a link between the open ocean variability and changes in the FC transport.

Analysis revealed that the main mechanism driving the response in the Florida Straits in experiments “with no background flow” (SENS-E0x) differed remarkably from the mechanism driving changes in the FC for the case “with background flow” (ModFC-E0x). Results showed that, in the absence of background flow, the response in the Florida Straits was driven by Rossby-wave anomalies and wall-jet features that were directly linked with the prescribed eddy. Because of this direct relationship between response and the prescribed eddy feature, we define this as “the direct response mechanism” for practical purposes. In the presence of the background circulation, the observed phase modulation of the FC transport (or response in the Florida Straits) was driven mostly by changes in the SHA variability along the coastal waveguide and on the offshore flank of the Gulf Stream that were indirectly linked with the prescribed eddy. We define this here as the “indirect response mechanism”. In general, using similar eddy-like signals prescribed in the same location, a faster response is generally observed at the Florida Straits for experiment “with no background flow” forced by the direct response mecha-
nism, and a delayed response forced by the indirect mechanism is later observed in the equivalent experiment “with the background flow” (left column, Figure 2.36). In this section, a more detailed discussion of the direct and indirect mechanisms is provided based on results from this study, and based on previous results based on observations and earlier modeling experiments reported in the literature.

*The direct response mechanism*

The experiments from set SENS-E0x showed that the variability induced in the Florida Straits can be characterized as a two-stage response. In a first stage, broad SHA signals reach the Bahamas archipelago before the prescribed eddy touches the topographic slope (Figure 2.4a, 2.5a). These SHA signals are then observed to rapidly extend over shallow waters in the northern Bahamas that are approximately enclosed by isobath contours (Figures 2.5b, 2.5b), generating a barotropic response in the velocity structure in the Florida Straits (Figures 2.5c, 2.5c). The rapid propagation of SHA signals in shallow waters near the Bahamas could not be isolated using outputs of averaged model solution every 7 days. These results suggest the propagation of fast barotropic signals, that can rapidly distribute pressure anomalies along contours of $f/H$. In fact, one study [Elipot et al., 2013] observed fast barotropic signals propagating at speeds of 128 m s$^{-1}$ ($\sim$10,600 km day$^{-1}$) along the Northeast U.S. coast. The second stage of the response was characterized by the development of features resembling the wall-jets described by Nof [1988, 1999]. These wall-jets were observed to propagate westward along the NWPC (Figure 2.4d) and along the northern coast of Bahamas (Figure 2.5d), or southward along the east coast of U.S. (Figure 2.6g, 2.7d). Westward propagation exhibited speeds ranging between -1.6 km day$^{-1}$ (SENS-E01) and -3.6 km day$^{-1}$ (SENS-E03). Along the U.S. east coast, wall-jets propagated at $\sim$2.5 km day$^{-1}$ (SENS-E08 and SENS-E09). These rates of propagation observed are much smaller than speeds expected for first-mode coastally-trapped waves.
Figure 2.36: Comparison of equivalent outputs from SENS-E0x experiments (no background flow) with ModFC-E0x (with background flow). (left column) Transport anomalies in the Florida Straits observed for the SENS-E0x experiments (blue line) and ModFC-E0x (red line). (center column) Fields of surface velocity from experiments SENS-E04, SENS-07, SENS-E08, and SENS-E09. (right column) Fields of surface velocity from experiment ModFC-E04, ModFC-E09, ModFC-E05, and ModFC-E10.

at the east coast of U.S., which are generally on the order of 1 m s\(^{-1}\) (86.4 km day\(^{-1}\)) [Elipot et al., 2013]. Hence, this indicates that jet propagation may be dominated by other
processes. The formation of wall-jets is largely associated with the actual volume of the eddy leaking along the wall [Nof, 1988, 1999], or along topographic contours in the case of this study, suggesting that advective processes play an important role. Results reported here largely support this hypothesis. Comparison between the average rate of jet propagation with the average barotropic velocity at the origin of the jet explains approximately 64% (r=0.8, Figure 2.11d) of the variance observed for different SENS-E0x experiments. In other words, results indicate that jet propagation is essentially an advective process that is controlled by the “peeling rate”, which sets the time-scale for draining the eddy volume through the gap at the entrance of Northwest Providence Channel, north of the Bahamas, and along the east U.S. coast. As the eddy leaks along the topographic slope, the wall jet forms and propagates approximately along contours of $f/H$ (Figure 2.11a-c), until it reaches the Florida Straits, where it may detach from the topography and give origin to smaller features.

It is interesting that, even though experiments SENS-E0x were developed in the absence of background flows, results reported here show, to some extent, agreement with results described in previous studies based on real ocean observations. For example, observations reported by one study [Frajka-Williams et al., 2013] suggest that similar wall-jet features linked with anticyclonic eddies originating east of the Bahamas may indeed develop and affect the FC in time-scales (50 days to 1 year) similar to the ones described here (~200 days). This indicates that, even though experiments from set SENS-E0x did not include the dominant features driving the background ocean circulation, they were still useful to highlight a potential role of the direct response mechanism for linking the FC changes with the open ocean variability. In one way, observations show that the AC is associated with elevated levels variability [Lee et al., 1990, 1996] and that its flow can even reverse occasionally [Johns et al., 2008]. Therefore, it is reasonable to consider that
at some times the eddies may interact with the boundary in a way that can be similar to the results presented here for the “no background flow” case studies.

The analysis also revealed that in the absence of the background flows associated with the FC and AC and the Gulf Stream, westward propagating eddy-like signals forced changes in the velocity field at the Florida Straits leading to transport anomalies that reached values as large as $\sim 1$ Sv (Figure 2.3a, Table 2.1). In fact, a 0.5 Sv transport anomaly was observed in the experiment (SENS-E01) developed with an anticyclonic eddy with SHA of 7.5 cm and 500 km wavelength, while a $\sim 1$ Sv anomaly was observed for experiments (SENS-E02-SENS-E06) developed with single eddies with SHA of 15 cm and wavelengths varying between 300-700 km. For the experiment developed using a cyclonic eddy with SHA of -15 cm and 500 km wavelength (SENS-E07), a transport anomaly of -0.8 Sv was observed. These results suggest that the magnitude of the response in the Florida Straits is sensitive to the intensity of the eddies reaching the Bahamas, but not to their wavelength. An extended analysis based on Uncertainty Quantification [Iskandarani et al., 2016] using eddy intensity and wavelength as input parameters could potentially provide a comprehensive assessment of the absolute responses in the Florida Straits. It is argued here, however, that additional sources of uncertainty may need to be quantified for a more accurate analysis based on Uncertainty Quantification. For example, because simulated features have different sizes and had slightly different tracks, the specific mechanisms by which interaction with the bottom topography takes place may vary from one simulation to another. Such discrepancies may ultimately affect the overall energy path of the eddy energy into the Florida Straits, and the magnitude of the response. While a more sophisticated analysis is left for future studies, results obtained here provide initial evidence to indicate that the response in the Florida Straits is sensitive to the intensity of westward propagating signals reaching the North Atlantic western boundary east of the Bahamas archipelago.
The intrinsic FC variability

While experiments with no background flow helped identify the direct response mechanism, experiments with the background flow provided a more realistic view of the processes that are more likely to take place in the real ocean. The indirect response mechanism identified in this study implies that westward propagating signals can modulate the seasonal phase of the FC transport mostly by causing changes in the SHA variability along the coastal waveguide. In other words, the indirect response mechanism is associated with changes in the phase of the FC variability, but not in the overall amplitude of the FC variability. This hypothesis is also supported by the fact that values of the STD variability from ModFC-E0x experiments (∼0.3 ± 0.1) are equivalent to the STD variability from the control experiment Cntr-E02 (STD = 0.3 Sv). Therefore, in the presence of background flows, the overall magnitude of the variability in the Florida Straits is essentially the same for both cases, for the control run experiment Cntr-E02, and for ModFC-E0x experiments, indicating that the levels of variability are largely associated with the intrinsic dynamics of the FC and Gulf Stream.

The intrinsic FC and Gulf Stream variability may be objectively evaluated by assessing results from the control run experiment Cntr-E02, which does not include any explicit eddy-like feature prescribed in the ocean interior. Analysis showed that meridional velocity variability in the Florida Straits was greater at the west side of the Straits along the topographic slope (Figure 2.17), for all frequencies evaluated in this study. These results are consistent with findings based on real ocean observations for the region [Leaman and Molinari, 1987; Leaman et al., 1987; Beal et al., 2008], which generally attribute this type of velocity variability to lateral meandering of the FC jet at this location and to coastally trapped waves. Indeed, results from the control run showed that SHA variability along the coastal waveguide was dominated by signals that propagated southward at speeds of ∼82 km day⁻¹, which is compatible with the rate observed for first-mode coastally trapped
waves at the northeast coast of U.S. [Elipot et al., 2013]. Results showed that the generation of coastally trapped signals was largely associated with the Gulf Stream meandering near Cape Hatteras. Snapshots of model solution indicated that SHA signals linked with meandering were advected downstream by the Gulf Stream towards Cape Hatteras, and agree with the downstream propagation of Gulf Stream meanders reported by Watts and Johns [1982], which showed that meanders were overall associated with spatial scales ranging between 180-600 km, and temporal scales of 4-30 days. Analysis of modeling outputs from this study shows that when these meanders (SHA signals) reach shallow areas at \( \sim 35^\circ N \), coastal SHA signals are excited, and observed to propagate southward, taking approximately 21 days to reach the Florida Straits.

It should be acknowledged that the analysis developed here does not address in detail the meander-shelf interactions triggering the southward propagation of coastally trapped signals. Previous studies, however, based on satellite altimetry observations [Hughes and Meredith, 2006] indicate that the interior oceanic variability can indeed be transmitted to the shelf through the generation of coastal-trapped waves, which generally translate in coherent coastal sea-level signals along large stretches of coastlines. Huthnance [1992] acknowledged that waves can be generated a full variety of mechanisms over the continental shelf and slope, including scattering by irregularities in the shelf, and generation by oceanic eddies offshore. The main mechanism for triggering these signals consists of imposing fluctuating offshore forcing in the form of pressure anomalies, from which waves will respond to the off-shelf forcing according to the similarities between the length and time scales in the forcing and the dispersion curve of a free wave [Chapman and Brink, 1987]. When the alongshore velocity of the forcing approximates that of a free mode (i.e. near-resonance), the response is dominated by the free mode, and the signal can be amplified towards the coast [Power et al., 1990]. The transmission of the off-shore signal to the shelf will be further affected by several parameters, such as the slope and shelf topogra-
phy, stratification, and friction [Ezer, 2016]. Interestingly, while waveforms are strongly influenced by the depth and by the density profiles [Huthnance, 1978], large friction improves the transmission of oceanic signals across the slope and shelf, as it reduces frictional decay distances to be comparable with the oceanic forcing [Huthnance, 2004]. At the shelf, the resulting response will be mostly barotropic [Chapman and Brink, 1987], or given by a hybrid baroclinic Kelvin wave/barotropic continental shelf wave, from which the structure depends on the steepness of the topography and strength of stratification [Huthnance, 1978]. At the North Atlantic western boundary, Gulf Stream meandering onto the shelf can provide a significant source of pressure anomalies for the triggering trapped waves along the coastal waveguide. In fact, previous studies [Ezer, 2016] found that variations in Gulf Stream transport can produce variations in sea level gradient across the entire length of the jet, and that this large-scale signal can be transmitted into the shelf by the generation of coastal-trapped waves.

The unstable regime and meandering pattern observed in the proximity of Cape Hatteras, and the generation of coastally trapped signals are the main feature associated with the intrinsic variability from our simulations. Lee and Csanady [1994] showed that baroclinic instability may be the main process driving the meandering variability of the Gulf Stream, which can extract energy from the mean flow into the eddy field. Their results also showed that instabilities were mainly baroclinic, and that 75-85% of the energy supply of the growing perturbations came from the basic flow potential energy. The more recent study by Kang and Curchitser [2015] showed that both barotropic and baroclinic instability mechanisms may play a role in driving the Gulf Stream instability upon separation from the coast at Cape Hatteras. Previous analysis based on in situ observations indicated that the Charleston Bump plays an important role in driving the amplification of both barotropic and baroclinic instabilities, which is largely associated with the Gulf Stream meandering downstream from this location [Dewar and Bane Jr, 1985]. While a
more detailed analysis on the specific mechanism driving the meandering variability of the Gulf Stream is beyond the scope of this work, the intrinsic variability from our experiments captures a realistic characteristic of the western boundary circulation in the North Atlantic ocean that is linked with the unstable regime of the Gulf Stream.

A secondary component of the intrinsic FC variability observed in the simulations developed in this study was linked with features at 27°N that could be traced to the interior of the ocean. Despite employing constant boundary conditions for all boundaries, lower amplitude signals behaving like first-mode baroclinic Rossby waves were observed to develop in the ocean interior. These signals, however, were much weaker than the explicit eddy-like features prescribed in the interior for ModFC-E0x experiments. Still, it is reasonable to consider that variability coming through the Northwest Providence Channel linked with these baroclinic Rossby waves played an noticeable role on the FC variability, given that observations indicate that the Northwest Providence Channel provides an active source for the Florida Current flow [Beal et al., 2008]. Previous analysis of in situ observations have also showed the development of intermittent flow in a band 10–15 km wide off Little Bahama Bank, on the east side of the Florida Straits [Leaman and Molinar, 1987], which were generally associated with meandering of the FC flow within the Florida Straits.

Results obtained here for the control run experiment Cntr-E02 indicate that the intrinsic FC variability simulated here is largely associated with the mechanisms described above. In terms of variance partitioning, a multi-linear regression analysis has been carried out using SHA time-series at both sides of the Florida Straits as proxies for the contribution of coastally-trapped waves, and of open ocean signals. This analysis revealed that, for the control run experiment, both processes contributed similarly, and that about 50% of the FC transport variance could be accounted using SHA time-series retrieved from opposite sides of the Florida Straits.
The indirect response mechanism

Once eddy-like mesoscale features were explicitly included in the ocean interior in the ModFC-E0x experiments, changes in the FC transport variability with respect to the control run experiment were observed linked with the indirect forcing mechanism. The observed changes in the FC variability were restricted to shifts in the phase of seasonal oscillations, whereas the overall variability in ModFC-E0x (STD = 0.3 ± 0.1) experiments remained equivalent to the variability in the control run Cntr-E02 (STD = 0.3 Sv). These results indicate that, in the presence of background flows, westward propagating signals may modulate the phase of the FC transport through the indirect response mechanism, while the magnitude of the variability is largely linked with the intrinsic dynamics of the systems and not directly dependent on the magnitude of the eddy-like signals reaching the western boundary. It is worth pointing out that this mechanism is remarkably different from the direct response mechanism described above, in which the magnitude of the response at the Florida Straits was sensitive to the intensity of the eddy-like features reaching the boundary. Below, the main characteristics of the indirect response mechanism are addressed.

Phase modulation of seasonal oscillations in the FC transport is likely the main outcome of the indirect response mechanism forced by westward propagating eddy-like features. Results suggest that this mechanism is characterized by complex processes that involve the nonlinear dynamics of the system through interaction of eddy anomalies with the unstable background circulation. For example, the initial stage of ModFC-E0x experiments was marked by an active interaction of prescribed eddies with the background flow of the AC: anomalies linked with the original eddy were predominantly advected northwestward as they were drained by the AC. Analysis based on cumulative transport differences with respect to the control run (Figure 2.32) showed that the propagating anticyclone caused a 10 Sv signal at 75°W and 28°N. This analysis also revealed that
westward drift and advection associated with the background AC flow resulted in variability with a time-scale of \( \sim 50 \) days. Assuming the prescribed anticyclone eddy to be associated with the positive crest of a baroclinic Rossby wave, under eddy-full configurations, the resulting periodicity would likely be approximately 100 days. Results reported here are consistent with previous analysis from in situ mooring observations collected at 26.5\(^\circ\)N east of the Bahamas [Lee et al., 1990, 1996], which have shown that transport variability in this location is generally dominated by westward propagating eddies moving at 4 cm s\(^{-1}\) (\(\sim 3.4 \) km day\(^{-1}\)) and mean periods of 100 days. Results from [Johns et al., 2008] also showed that the baroclinic transport variability in this location in the upper 1000 m ranged from \(\sim -5 \) Sv to 20 Sv. Böning and Budich [1991] also applied one of the earliest eddy-resolving models to the North Atlantic western boundary, reporting that strong fluctuations with time-scales on the order of 100 days masked the annual cycle east of the Bahamas, potentially providing one source of transient variability for this location. Results based on our controlled approach confirm the general understanding that eddy-like westward propagating eddies can drive strong volume transport variability east of the Bahamas with amplitude on the order of 10 Sv and time-scales of \(\sim 100 \) days.

Further analysis from ModFC-E0x experiments showed that, following the initial advection by the AC, anomalies associated with the eddy were incorporated into the offshore circulation the Gulf Stream. Cumulative transport differences with respect to the control run simulation showed that the interaction of small anomalies with the background circulation could lead to transport differences as large as 5 Sv at 30\(^\circ\)N (Figure 2.32). These results suggest that signals coming from interior may play a role in sustaining the offshore circulation associated with the Gulf Stream. In fact, one study Lee [2001] addressed the structure and energy pathway for supercritical western boundary currents, and found that energy was actively transported from the interior of the ocean to the western boundary layer via ageostrophic pressure work divergence. His study also showed that about 53%
of the mean energy was then converted back to eddy energy via turbulence production in the outer offshore layer. Results obtained here indicate that similar processes may have also occurred in ModFC-E0x experiments. For instance, when anomalies that could be originally traced to prescribed eddies reached latitudes in the proximity of Cape Hatteras, they were generally amplified and associated with intense Gulf Stream meandering. Analysis showed that the meandering pattern at 34°N varied remarkably among the ModFC-E0x experiments with respect to the control run experiment, suggesting that the small perturbations included in the ModFC-E0x could lead to large changes in the variability. Therefore, while a detailed energy budget is not performed in this study, there is evidence to suggest that the indirect forcing mechanism is linked with eddy-to-mean and mean-to-eddy energy exchanges. Further studies are needed to confirm this.

Observed changes in the SHA variability along the coastal waveguide in ModFC-E0x experiments with respect to the control run Cntr-E02 provided the ultimate link for understanding the modulation of the FC variability. Results showed that coastally trapped signals were largely triggered by the meandering activity at Cape Hatteras. Hence, changes in the meandering variability forced by the nonlinear perturbations linked with the prescribed eddy in the interior had likely an important role in changing the coastal SHA variability. It is important to emphasize that the modulation of the FC variability observed in ModFC-E0x experiments resulted from net signals reaching the Florida Straits. In other words, the FC variability will “feel” SHA signals that reach the Florida Straits, since constructive / destructive interactions between coastally trapped signals generated at Cape Hatteras with signals generated by lateral meandering midway (e.g. at 30°N, Figure 2.22) may sometimes be observed. In addition, coastally trapped signals generated due to lateral meandering midway between Florida Straits and Cape Hatteras forced by eddy anomalies may have also played a role. Nevertheless, the modulation of the FC variability was linked with integrated changes in the coastal SHA variability reaching the Straits.
Our results are largely complementary to findings from Domíngues et al. [2016], which reported that approximately 50% of the variance from the transient component of the FC transport was accounted for by integrated coastal SHA changes along the east coast of U.S. linked with westward propagating signals based on analysis of satellite altimetry observations and in situ tide gauges. Therefore, both the modeling component developed in this study, and the observational component reported in Domíngues et al. [2016] indicate that westward propagating signals can play an important role in modulating the seasonal variability of the FC transport through the indirect response mechanism, which is linked with changes in the coastal SHA variability.

Rousset and Beal [2011] observed that during 2001-2006 the FC and Yucatan current exhibited similar seasonal variability, which, for the FC, was characterized by maximum of 33 Sv in summer and minima of 28-30 Sv in spring and fall, and a secondary maximum in winter. It is possible that changes induced by westward propagating signals in the Florida Straits variability identified here can be transmitted along the U.S. coast towards the Gulf of Mexico, and may also modulate the Yucatan Current seasonality. The regional domain employed in this study, however, is restricted to areas outside the Gulf of Mexico, and prevents the evaluation of such mechanism. Further studies based on similar approaches may be able to provide insightful information on the role of westward propagating signals on modulating the Gulf of Mexico variability.

In summary, the response observed in the Florida Straits for experiments with a background flow (ModFC-E0x) differed remarkably from the response for experiments developed without the background flow (SENS-E0x). In addition, some gaps in knowledge have been potentially addressed by the simulations developed here regarding the potential blocking impact of the Bahamas archipelago to signals coming from the interior, as previous studies relying on low resolution [Czeschel et al., 2012] or simplified ocean models [Pedlosky and Spall, 1999; Simmons and Nof, 2002] could not realistically repro-
duce the dynamics the of the region. For example, results from experiments developed in the absence of background flows indicate that indeed most of the open ocean variability may reach the Florida Straits through Northwest Providence Channel as broad barotropic signals, followed by baroclinic wall-jet features. Experiments developed with the background flow, on the other hand, suggest that the AC flow may function as a more effective barrier for signals coming from the ocean interior, which often get advected northwestward, and can later modulate the FC variability only through the indirect response mechanism. However, since the AC flow is associated with strong variability [Lee et al., 1990, 1996], and it is oftentimes observed to halt and reverse [Johns et al., 2008], it is likely that westward propagating signals may drive variability in the Florida Straits through the direct response mechanism under these specific circumstances. In fact, one study Frajka-Williams et al. [2013] reported changes in the FC and AC transport with seasonal time-scales associated with the development of elongated anomalies in the Florida Straits resembling wall-jets that could be traced to anticyclonic eddies originating east of the Bahamas. Therefore, it is likely that both the direct response and the indirect response mechanisms play a role in linking the open ocean variability to changes in the FC seasonal variability.
Chapter Three  
On the Impact of Realistic Westward Propagating Signals Based on Satellite-Altimetry Data

Experiments developed in the previous chapter helped assess the response and mechanisms linked with seasonal variability in the Florida Straits forced by single eddy-like westward propagating signals (questions Q1 to Q6, section 1.3). In this chapter, an additional step is taken in an effort to simulate conditions that are more likely to be observed in the real ocean, using realistic boundary conditions and an initial state derived from satellite-altimetry observations filtered for the transient seasonal band.

To accomplish this, seven additional experiments are developed (Table 3.1): experiments EdFul-E01 and EdFul-E02 are developed in the absence of FC flow, while five experiments from set EdFulFC-E0x are developed with the background flows. It is hoped that analysis of these experiments will support findings reported in Chapter Two, and also provide additional insight on question Q7 (section 1.3), which aims to understand whether
Table 3.1: Properties of experiments from set EdFul-E0x and EdFulFC-E0x.

<table>
<thead>
<tr>
<th>Exp. Name</th>
<th>FC flow</th>
<th>Description</th>
<th>FL Straits STD.</th>
</tr>
</thead>
<tbody>
<tr>
<td>EdFul-E01</td>
<td>NO</td>
<td>initialized with meridionally elongated Rossby waves</td>
<td>0.2 Sv</td>
</tr>
<tr>
<td>EdFul-E02</td>
<td>NO</td>
<td>initialized with zero-flow; eastern boundary conditions based on filtered altimetry data</td>
<td>0.6 Sv</td>
</tr>
<tr>
<td>EdFulFC-E01</td>
<td>YES</td>
<td>HYCOM-derived initial conditions; ocean interior initialized with filtered altimetry data for 01/06/1995</td>
<td>0.5 Sv</td>
</tr>
<tr>
<td>EdFulFC-E02</td>
<td>YES</td>
<td>HYCOM-derived initial conditions; ocean interior initialized with filtered altimetry data for 01/03/1997</td>
<td>0.4 Sv</td>
</tr>
<tr>
<td>EdFulFC-E03</td>
<td>YES</td>
<td>HYCOM-derived initial conditions; ocean interior initialized with filtered altimetry data for 01/05/2001</td>
<td>0.3 Sv</td>
</tr>
<tr>
<td>EdFulFC-E04</td>
<td>YES</td>
<td>HYCOM-derived initial conditions; ocean interior initialized with filtered altimetry data for 01/03/2003</td>
<td>0.3 Sv</td>
</tr>
<tr>
<td>EdFulFC-E05</td>
<td>YES</td>
<td>HYCOM-derived initial conditions; ocean interior initialized with filtered altimetry data for 01/07/2005</td>
<td>0.5 Sv</td>
</tr>
</tbody>
</table>

variability forced by these signals can drive year-to-year changes in the seasonality of the FC transport.
Figure 3.1: Initial condition fields of SSH for experiments: (a) Cntr-E01 - zero flow; (b) EdFul-E01 - elongated Rossby waves; (c) EdFul-E02 - eastern boundary conditions based on satellite altimetry data; (d) Cntr-E02. EdFulFC-E0x are initialized using satellite altimetry data for (e) EdFulFC-E01 - 01-06-1995; (f) EdFulFC-E01 - 01-03-1997; (g) EdFulFC-E01 - 01-05-2001; (h) EdFulFC-E01 - 01-03-2003; (i) EdFulFC-E01 - 01-07-2005.

3.1 Experiment setup

Before the application of more realistic initial and boundary conditions based on altimetry data, an initial experiment designated EdFul-E01 (Table 3.1) is developed using meridionally elongated Rossby waves as initial conditions (Figure 3.1b).

Because the satellite altimetry record contains the signals of both Rossby waves and mesoscale eddies [Oliveira and Polito, 2013; Polito and Sato, 2015], experiment EdFul-E01 is designed to gain initial insight on the response forced at the Florida Straits purely
due to incoming Rossby waves that have a meridional scale scale that could simultaneously influence much of the southeast U.S. coast. To accomplish this, perturbations on isopycnal surfaces are performed applying equation 1.5d using the original Gaussian function $G_{xy}(x)$ on a meridionally elongated elliptical reference. The wavelength is set to 500 km and the maximum SSH to 7.5 cm, as the objective is to simulate a broad Rossby wave field.

Experiment EdFul-E02 (Table 3.1) is also developed in the absence of background flows. This experiment is started from rest (Figure 3.1c), and altimetry-derived perturbations are applied as boundary conditions on the eastern side of the domain (at 61°W, Figure 1.9). In order to apply altimetry-derived as boundary conditions, we first assume that the filtered satellite-altimetry SHA signals are linked with first mode baroclinic perturbations. This is a reasonable assumption, given that these signals propagate westward at speeds similar to those expected for first mode baroclinic Rossby waves (Figure 1.7a, see Section 1.2.2). The filtered SHAr data are then extracted at 61°W, and used to prescribe perturbations in the pycnocline depth. To accomplish this, an approach similar to the method described in Section 1.4.3, is applied, except that the Gaussian function $G_{xy}(x)$ used in equation 1.5d is replaced by the actual SHA data along the boundary. Perturbations in isopycnal surfaces are applied so that changes in dynamic height integrated down to the bottom of the ocean match the actual SHA signal. For the velocity field, the actual altimetry-derived geostrophic velocity is used at the surface, and extended to subsurface depths using the first baroclinic mode for horizontal velocity (Figure 1.10f). Experiment EdFul-E02 is then carried out for 8 years using the filtered SHA data starting in January, 1995. The objective of this longer simulation is to evaluate year-to-year changes in the seasonality of the response forced in the Florida Straits.

The five experiments from set EdFulFC-E0x (Table 3.1) are developed in the presence of background flows. For these experiments, initial conditions are defined using snapshots
of the satellite altimetry between 1995-2005 in a eddy-full configuration for the ocean interior (Figure 3.1d-h). Perturbations in the initial temperature and salinity fields are prescribed using the same approach as the one described above for experiment EdFul-E02; except that perturbations on isopycnal surfaces are applied everywhere in the domain where water column is deeper than 4000 m, and are not restricted to the eastern boundary. These simulations are carried out for two years, which is the approximate time that it takes for signals to cross the domain. The experiment that would otherwise start in January 1999 is skipped, given that there is a gap in the FC transport time-series derived from the telephone cable in the Florida Straits. A comparison with the observed FC variability is performed in this study with the goal of verifying how changes in the FC seasonality observed by the cable can be reproduced by simplified simulations forced by westward propagating signals.

3.2 Results

In this section, results are presented starting from the simplest case for EdFul-E01 — experiment developed with no background flow and elongated Rossby wave field in the interior (Figure 3.1b), and concluding with the more complex experiments from set EdFulFC-E0x — experiments developed with background FC flow and initialized with satellite altimetry signals in the interior (Figure 3.1e-i). Unlike the analysis developed in Chapter 2, which focused more on the dynamic mechanisms driving the response in the Florida Straits, the analysis developed here focuses on the overall characteristics of the response driven by the more realistic conditions based on eddy-full configurations. Some of the characteristics assessed include the overall magnitude of the response, the periodicity, and the timing.
Before proceeding with the analysis of experiments based on eddy-full ocean interior configurations, experiment EdFul-E01 (Figure 3.1b) was designed to gain additional insight on the response in the Florida Straits driven by a broad Rossby wave field. As expected, westward propagation is controlled by Rossby wave dispersion (not shown), and signals reach the eastern side of the Bahamas after \(\sim 84\) days of simulation (Figure 3.2b). As the simulation evolves, wave crests tend to get tilted with latitude (Figure 3.2c,d). This results from Beta-dispersion, which causes Rossby waves crests to propagate faster (slower) at lower (higher) latitudes. Another interesting feature is that as signals reach shallower areas, smaller scale features are observed to develop. This indicates that interaction with the bottom topography may play an important role in breaking these waves into smaller features. Investigating this specific mechanism, however, is beyond the scope of this work.
Figure 3.3: (a) Time-series of meridional volume transport anomalies at the Florida Straits for experiment EdFul-E01. (b) Longitude-time Hovmoller diagram of SHA for experiment EdFul-E01. (c) Standard deviation of meridional velocity anomalies developed at the Florida Straits for experiment EdFul-E01.

The time-series of volume transport at the Florida Straits for experiment EdFul-E01 (Figure 3.3a) exhibits the development of a maximum anomaly with magnitude of approximately 0.4 Sv. The absolute range of the transport variability is $\sim 0.7$ Sv, with a STD of 0.2 Sv. The time-series further indicates a response with periodicity of approximately 200 days, which is similar to the time-scales observed for the response to the SENS-E0x experiments (Figure 2.3). Longitude-time Hovmoller diagram of SHA for this experiment (Figure 3.3c) shows that the periodicity of 200 days observed in the Florida Straits is in good agreement with the open ocean time-scales set by westward propagation, indicating the important role of Rossby wave dispersion in setting up the response time-scales in the Florida Straits.

Analysis of the standard deviation of the meridional velocity at the Florida Straits (Figure 3.3c) also shows that the variability is linked with broad changes in the velocity field associated with small vertical shear, suggesting the dominant effect of barotropic
mechanisms. These results are consistent with the first stage of the response discussed for the SENS-E0x experiments, in which Rossby waves that propagated faster than the eddy forced barotropic velocity anomalies in the Florida Straits before the development of the wall-jet features. Therefore, results from experiment EdFul-E01 can be understood as the isolated first stage of the response from experiments SENS-E0x, and it indicates that a field forced by purely Rossby wave-like signals may drive barotropic anomalies causing transport values as large as 0.4 Sv, with periodicity of \(\sim 200\) days. However, It should be noted that under more realistic conditions, real ocean observations in the Florida Straits indicates the dominance of baroclinic variability over barotropic mode variability [Meinen and Luther, 2016].

The predominantly barotropic response in the Florida Straits forced by impinging baroclinic waves is likely largely associated with the effect of the bottom topography in regulating signals that will be admitted over shallow bathymetry, and the signals that will be reflected/dispersed along the sloping topography. For example, the study by [Smith IV, 1986] showed that baroclinic eddies with a weak lower-layer expression evolve quickly to upper-layer features in the presence of topography because of dispersion of the lower-layer feature by topographic Rossby waves. The lower-layer may “feel” the slope as vertical walls, leading to rapid topographic dispersion that can erode the lower layer structure, leaving an upper layer signals to propagate independently of topography. The resulting response over the shelf or over regions with shallower topography will then be characterized by mostly barotropic signals [Chapman and Brink, 1987], or given by a hybrid baroclinic Kelvin wave/barotropic continental shelf wave, from which the structure depends on the steepness of the topography and strength of stratification [Huthnance, 1978]. The baroclinic structure prescribed on the various experiments developed in this study was associated with an upper-layer above \(\sim 1500\) m, and a lower layer below (Figure 3.4). Therefore, it is likely that only signals from the upper-layer (of the same sign) are admit-
Figure 3.4: Cross-section at 25.6°N showing the meridional velocity component for experiment EdFul-E01 at time: (a) initial condition, and (b) 84 days. The green circle indicates the entrance to Northwest Providence Channel.

109

...ted through the shallow bathymetry at the Bahamas slope, which is associated with depths shallower than 1500 m at Northwest Providence Channel and in the Florida Straits (Figure 1.9), causing a barotropic-like response over shallow topography. In fact, baroclinic signals can indeed radiate significant barotropic energy upon interaction with the bottom topography [Louis and Smith, 1982].

Proceeding towards an increased level of realism, experiment EdFul-E02 was started from rest (Figure 3.1c) and developed with boundary conditions that were perturbed with satellite altimetry data on the eastern boundary. Analysis of snapshots of model solutions for SHA (Figure 3.5) indicate that broad and small amplitude signals reach the western boundary after approximately one year of simulation, while for the dominant signals it takes approximately 2 years to cross the domain. This indicates that the perturbed boundary condition excited long Rossby waves that propagated faster than the eddy-like signals that are the focus of this study. It is interesting to note, however, that, while some snapshots of model solutions show a pattern dominated by eddy-like anomalies in the
Figure 3.5: Snapshots of model solution of SHA for experiment EdFul-E02.

...ocean interior (Figure 3.5f), other snapshots show a meridionally-elongated pattern (Figure 3.5e) resembling the solution from experiment EdFul-E01, with wave crests tilted as a function of latitude (Figure 3.2c,d). Because experiment EdFul-E02 was solely forced with satellite altimetry data at the eastern boundary, these results provide further evidence to previous observations [Oliveira and Polito, 2013; Polito and Sato, 2015] that the altimetry record contains the signals of both Rossby waves and mesoscale eddies.

Time-series of volume transport at the Florida Straits for experiment EdFul-E02 (Figure 3.6) indicates the development of anomalies reaching values with magnitudes as large...
The absolute range of transport variability is 2.8 Sv, with a STD of 0.6 Sv. The first three-years of the time-series are generally characterized by transport anomalies smaller than 0.5 Sv in magnitude, and the STD during these first three years is 0.2 Sv. After this initial period, anomalies often reach values larger than 1 Sv, in magnitude, and the STD is 0.7 Sv. This result indicates that three years is the approximate time-frame that it takes for the western boundary to become saturated with anomalies that originated in the eastern boundary of the domain. For the purposes of this study, the saturated state provides a closer representation of the real ocean.

Snapshots of the model solution for experiment EdFul-E02 further indicate that similar mechanisms reported for experiments SENS-E0x (developed single eddy-like features) may also be observed once the saturated eddy-full state is reached. For example, peak volume transport anomalies observed during the EdFul-E02 simulation are largely linked with the development of wall-jet features (Figure 3.7a). The velocity structure at maximum response time (Figure 3.7b) exhibits a vertical shear characterized by maximum velocities above 200 m associated with the jet, and by barotropic-like velocities below. These results suggest that variability in the Florida Straits forced by eddy full configurations is also consistent with the two-stage response discussed for experiments SENS-E0x, which is initially dominated by the development of barotropic anomalies.
linked with the Rossby wave field, and then by the development of baroclinic anomalies linked with the wall-jet.

The time-series of volume transport for experiment EdFul-E02 exhibits well-defined transport oscillations with periods ranging between 250-440 days (Figure 3.6). This indicates that westward propagating signals originating in the open ocean can drive transient seasonal variability (73-525 days) in the Florida Straits. To illustrate how these anomalies would impact the annual phase of the FC, the transport anomalies from individual years are plotted together in Figure 3.8. This analysis shows that the variability forced by the westward propagating signals is not linked with a fixed annual phase. For example, while low transports were observed from between May-October on simulation year-5 (yellow line, Figure 3.8), low transports were observed from November-February in year-7 (red-line, Figure 3.8). In fact, a statistical evaluation based on a t-test at 95% confidence level indicates that the average annual cycle computed using the seven years of simulation (black line, Figure 3.8) is statistically equal to zero. In other words, the variability at the Florida Straits forced by westward propagating signals in the absence of background
flows is characterized as stochastic variability that is not linked with any consistent annual phase.

The analysis now turns to the evaluation of experiments based on eddy-full configurations in the presence of background flows. For simulations with the background flow, setting up a perturbed boundary condition on the eastern edge of the domain using the altimetry data, as in experiment EdFul-E02, was not technically possible. This is because the active and constant inflow associated with the background circulation caused boundary condition inconsistencies when added to the variable perturbations on the eastern boundary, resulting in unstable and unrealistic circulation patterns inside the domain. The five experiments from set EdFulFC-E0x using snapshots of satellite altimetry were then developed in order to achieve the same objective, but using five independent simulations.

Time-series of FC transport at the Florida Straits for experiments EdFulFC-E0x exhibit a mean volume transport of $29.7 \pm 0.4$ Sv (Figure 3.9). Considering the FC time-
series from all EdFulFC-E0x experiments, the FC transport shows values ranging between 28.8 Sv and 30.8 Sv, an absolute range of ≈2 Sv. The STD variability for individual experiments ranges between 0.3 Sv for experiments EdFulFC-E03 and E04, and 0.5 Sv for experiment EdFulFC-E01 and E05 (Table 3.1). The STD from experiments EdFulFC-E0x has a value of 0.4 ± 0.1 Sv and is statistically equivalent at 95% confidence level to the STD from the control run experiment Cntr-E02 (STD of 0.3 Sv). It should be acknowledged here that the simulated FC variability is artificially low compared to real ocean observations of the FC variability for time-scales of interest in this study (STD = 1.6 Sv for the FCt), which is further discussed below. In addition, these results suggest that adding westward propagating signals to the interior circulation did not result in a significant increase in the overall western boundary variability. In fact, as previously mentioned for experiments ModFC-E0x, the inclusion of eddy-full configurations on EdFulFC-E0x implied an increase of only about 4% in the domain integrated kinetic energy per unit mass, from $165.4 \times 10^3 \text{ m}^2 \text{s}^{-2}$ in the control run Cntr-E02 to $[172.5 \pm 0.3] \times 10^3 \text{ m}^2 \text{s}^{-2}$ in EdFulFC-E0x. While the kinetic energy showed a modest increase with respect to the control run, the total energy remained the same for all simulations. This result is in agreement with findings from ModFC-E0x, and suggest that the overall magnitude of variability in the Florida Straits is dominated by the intrinsic variability of the background circulation features in the western boundary. It is worth emphasizing, however, that the phase of FC time-series from individual EdFulFC-E0x experiments varies considerably from one experiment to another (Figure 3.9), indicating that different configurations of eddy-full interior may lead to different responses.

Results from experiments EdFulFC-E0x indicate that the main link between westward propagating signals with changes in the FC variability is through the indirect response mechanism discussed for experiment ModFC-E0x (see Discussion 2.3). For example, snapshots of the model solutions of SHA for experiment EdFulFC-E01 (Figure
3.10) show westward propagating signals prescribed in the interior reaching the western boundary at different latitudes along the Gulf Stream. As SHA signals interact with the Gulf Stream, broad signals resembling coastally trapped waves are seen to propagate along the coastal waveguide towards the Florida Straits (Figure 3.10b,c). It is interesting to note, however, that even though the indirect response mechanism appears to be the dominant mechanism indicated by these results, experiments EdFulFC-E0x also provide evidence that the direct response mechanism may occur under certain circumstances. For example, snapshots of model surface velocity for experiment EdFulFC-E01 (Figure 3.11) exhibit the development of features in the Northwest Providence Channel resembling the wall-jets identified for experiments SENS-E0x (Figure 2.4). In those cases, these features provide a direct link between the FC in the Florida Straits and anticyclonic features located east of the Bahamas. Therefore, results from EdFulFC-E0x experiments indicate that in more realistic eddy-full configurations, westward propagating signals may drive changes in the FC variability through both the direct and indirect response mechanisms.

To assess the changes in the seasonality of the FC arising from these westward propagating signals, Figure 3.12 compares the seasonality of the first year of simulation with the seasonality of the second year for each individual EdFulFC-E0x experiment. This analy-
Figure 3.10: Snapshots of model solution of SHA for experiment EdFulFC-E01 referenced to the averaged dynamic topography from the control run experiment Cntr-E02.

Figure 3.11: Snapshots of surface velocity fields for experiment EdFulFC-E01 in the proximity of the Florida Straits.

sis indicates that, even within the same experiment, the seasonal FC variability may vary from one year to the following. For example, experiment EdFulFC-E02 (Figure 3.12b) exhibits a weak semi-annual oscillation with amplitude of $\sim0.5$ Sv during the first year of the simulation, and a annual oscillation with amplitude of $\sim1$ Sv peaking in January-
March during the second year. Another interesting example, experiment EdFulFC-E03, also exhibits a slightly stronger semi-annual oscillation with amplitude of 0.8 Sv in the first year of simulation, and an annual oscillation with 0.7 Sv peaking in November-December. Overall, results from EdFulFC-E0x experiments using different eddy-full initializations for the ocean interior resulted in remarkably different seasonal variability in the Florida Straits (Figure 3.13), with dominant signals showing periods ranging between ~180-400 days. In fact, analysis shows that the ensemble average seasonal cycle from the different EdFulFC-E0x (black line, Figure 3.13) is statistically equal to the mean FC transport 29.7 Sv at the 95% confidence level. This indicates that the resulting seasonal variability observed in the Florida Straits forced by westward propagating signals can be characterized as stochastic seasonal oscillations that are not linked to any consistent annual phase, both for the case with no background flow and for the case with background flow.

Finally, the seasonality simulated by the EdFulFC-E0x experiments can be compared to the actual transient component of the FC transport (the FCt) observed at the Florida Straits during 1995-2007. This period was selected to avoid the edges of the time-series that were filtered for the 73-525 days band, and considering the limited computational resources used to develop the simulations from this study. This comparison aims to evaluate if the modeled variability is consistent with the observed FC variability derived from the cable. Analysis of FCt transport anomalies with respect to the mean FC transport (Figure 3.14) for the period indicate that the FCt seasonality also varies remarkably during 1995-2007. In fact, analysis show that during this period the FCt is not linked with a statistically significant annual cycle, which is consistent with results from experiment EdFulFC-E0x. It is evident, however, that the actual range of FCt (-3.5 Sv to 3.5 Sv) is much larger than the simulated range of transport anomalies in the Florida Straits (-1 Sv to 1 Sv). The STD FCt is 1.6 Sv, and is approximately three times larger than the STD
Figure 3.12: Plots of volume transport oscillations for individual ModFC-E0x experiments as function of months. The red line shows the transport for the first year of simulation, and the blue line shows the transport for the second year of simulation.

Simulated for the EdFulFC-E0x and for the EdFul-E02 experiment. Further consequences of these results will be discussed below.
Figure 3.13: Similar to Figure 3.8, but for results from experiments EdFulFC-E0x.

Figure 3.14: (a) Volume transport anomalies from EdFulFC-E0x experiments as a function of month. (b) Same as panel (a), but using the cable-derived observed transient component of the FC transport.
3.3 Discussion

In this chapter, two sets of numerical experiments were developed with the objective of further investigating the role of westward propagating signals in driving variability within the transient seasonal band (73-525 days) in the Florida Straits for the “no background flow” case (EdFul-E0x), and for the case “with background flow” (EdFulFC-E0x). More specifically, eddy-full configurations of boundary or initial conditions based on satellite altimetry data were employed to enable a more realistic approach to investigate how these signals may affect the seasonality of the FC transport in the Florida Straits.

In situ observations show that seasonal (73-525 days) changes in the FC transport correspond to approximately 35% of the FC total variance. From these 35%, 8% correspond to the average annual cycle, which is generally associated with local / non-local wind forcing [Schott et al., 1988], while the remaining 27% correspond to the transient FC component. One study [Domingues et al., 2016] has shown that the transient FC component is largely associated with year-to-year changes in the FC annual cycle that were first reported by Baringer and Larsen [2001]. Given that the levels of variability associated with the transient FC component are three times larger than that of the average FC annual cycle, the wind-forced annual cycle only emerges from the FC cable record after averaging the data over decades of continuous observations [Atkinson et al., 2010].

In fact, one study [Meinen et al., 2010] showed that at least 25 years of observations are required to obtain an average annual cycle with an accuracy of 0.2 Sv, and that at shorter observational periods, the annual cycle calculations can be obscured by the stochastic signals. In this study, results indicate that westward propagating signals originating in the open ocean can indeed provide a source of stochastic variability causing changes in the FC seasonal variability, as hypothesized by Meinen et al. [2010].
Westward propagating signals were simulated in this chapter using real satellite altimetry observations (Figure 3.1) filtered for the transient seasonal band for both the experiments with no background flow (EdFul-E02), and with the background flow (EdFulFC-E0x), which include the dominant (semi-annual and annual) first-mode baroclinic Rossby waves for the North Atlantic Ocean [Polito and Liu, 2003]. These experiments showed the development of volume transport variability in the Florida Straits with time-scales of $\sim 250-440$ days for experiments with no background flow (EdFul-E0x), and of $\sim 180-400$ days for experiments with background flow (EdFulFC-E0x). These results are consistent with findings from Domingues et al. [2016], which suggested, based on the analysis of satellite and in situ observations, that the transient component of the FC transport was largely forced by westward propagating signals originating in the open North Atlantic Ocean. Therefore, this study provides the modeling framework to confirm that these signals may indeed drive variability in the Florida Straits within the 73-525 days frequency band.

The assessments performed in this study showed that the resulting seasonality in the Florida Straits is stochastic in terms of not displaying a consistent annual phase. In other words, numerical simulations showed that westward propagating signals do not produce a statistically significant annual cycle, and that the resulting variability is strictly associated with transient seasonal component. These results diverge slightly from findings of Czeschel et al. [2012], which was based on an adjoint model approach designed for investigating the FC annual cycle. While their study also acknowledge the role of baroclinic signals originating in the ocean interior for driving seasonal variability in the FC, they reported that baroclinic Rossby waves produced a fixed phase in the seasonal variability. Results obtained here, however, suggest that seasonal transport oscillations produced by westward propagating signals is mostly stochastic. It should be noted here that our seasonality assessments are based on an 8 year record for the “no background flow” case,
and on a 10 year record for the case “with background” flow, and that at least 25 years of observations are required to obtain an average FC annual cycle with an accuracy of 0.2 Sv [Meinen et al., 2010]. It is argued here, however, that some of the sources of variability and error included in the real ocean observations considered by [Meinen et al., 2010] are explicitly not included in the simulations or are insignificant. For example, sources such as interannual wind-driven variability or high-frequency signals coming from the Gulf of Mexico are intentionally not included in the numerical experiments to enable controlled simulations focused on the impact of westward propagating features on the FC.

The detailed analysis presented in Meinen et al. [2010] showed that year-to-year changes in the FC annual cycle, such as the ones reported by Baringer and Larsen [2001] were largely influenced by the elevated levels of background stochastic variability linked with other processes and time-scales. Results presented in this study suggest that westward propagating signals may potentially provide an important source for the stochastic variability found by Meinen et al. [2010]. It should be noted here, however, that the overall magnitude of the simulated stochastic variability forced by westward propagating signals in the Florida Straits is much smaller than the values observed in the FC cable record. For example, simulations showed an absolute range of transport anomalies of ∼2 Sv (2.8 Sv) and STD variability of 0.4 Sv (0.6 Sv) for the experiments with the background flow (with no background flow), while cable observations show that the transient component of the FC transport exhibit an absolute range of ∼8 Sv and STD of 1.6 Sv. These results indicate that the variability linked with westward propagating signals simulated here can only account for 25% of the observed variability. While this seems to indicate that westward propagating signals may correspond to a minor contributor of the transient component of the FC transport, it is argued here that this discrepancy is probably associated with limitations in our modeling experiments.
The main sources of discrepancies between the simulated and the observed magnitude of the transient seasonal FC variance can be related to two factors: (1) the model configurations employed in this study, and (2) the actual nature of the response mechanisms proposed. First, the fixed boundary and no-wind forcing conditions configurations used in experiment EdFulFC-E0x were employed to enable a controlled environment to investigate the influence of westward propagating signals on the FC seasonal variability. An outcome of these configurations is that other processes that can drive changes in the FC transport are explicitly not included in the simulations. For example, changes in the FC transport forced by the local \cite{Wunsch1969, Schott1988} or remote \cite{DiNezio2009} wind forcing, or by the upstream Loop Current dynamics in the Gulf of Mexico \cite{Lin2010, Mildner2013} are not reproduced by our numerical experiments. The nudging-radiation boundary conditions adopted for the simulations developed in this study allow for setting up constant inflow conditions to a known value at the boundary (nudging timescale set to 1 day), while still permitting anomalies generated inside the domain to be radiated out. Hence, the total FC variance estimated for our experiments is solely due to the internal and intrinsic dynamics of the domain, and is therefore, much smaller than the values from the FC cable in the Florida Straits. In addition, another limitation of experiments developed in this study is the latitudinal extent of the domain, which only covers the band between 21°N and 37°N. \cite{Domingues2016} showed that westward propagating signals reaching the western boundary between 25°N and 42°N can potentially affect the FC seasonality. Results reported in that study showed an abrupt increase in the explained FCt variance when signals between 37°N and 42°N were also considered. If only the latitudinal range between 25°N and 37°N is considered, SHA signals only explained 35% of the FCt variance (closer to the 25% simulated here), as opposed to 50% when signals between 25°N and 42°N were considered. Therefore, it is likely that the low FC variance simulated in this study may be partly because the
simulations developed here do not include the full extent of the western boundary linked with the FC variability.

Second, the nature of the mechanisms proposed here imply that westward propagating signals can: (i) drive changes in the FC transport through the direct response mechanism — in which case the magnitude of the response (or variance) is sensitive to the intensity of the signals; or (ii) modulate the phase of the FC transport through the indirect response mechanism — in which case the magnitude of the response (or variance) is largely insensitive to the magnitude of the signals. Results from experiments EdFulFC-E0x suggest the the indirect response mechanism probably plays a dominant role in the real ocean, while the direct response mechanism may still be able to drive FC variability. Therefore, it is likely that under more realistic ocean conditions, when other sources of variability are available to the FC (larger FC transport STD), westward propagating signals may drive transient FC variability by modulating the phase of its intrinsic variability (or variability related to other processes), and not by directly driving transport fluctuations. In other words, results obtained here suggest that westward propagating signals may behave as perturbations that can drive the stochastic and nonlinear dynamics of the system to evolve into a different state, in a way similar to the “butterfly effect”. These results are consistent with findings from Domingues et al. [2016] based on the analysis of tide-gauges and of satellite altimetry data along the east coast of U.S. According to that study, changes in the phase of the transient seasonal component of the FC transport were largely correlated with the phase of incoming westward propagating SHA signals traveling along different latitudes. Domingues et al. [2016] also showed, that, once these signals reached the coast, approximately 50% of the variance of the transient FC component was accounted for by integrated coastal SHA signals traveling along the east coast of U.S. Observations analyzed in that study are consistent with the indirect response mechanism proposed here, which is largely linked with the modulation of the SHA variability along the coastal
waveguide. Therefore, both the modeling components presented in this study and the observational components reported by Domingues et al. [2016] indicate that westward propagating signals play an important role modulating the FC seasonal variability, and most likely through the indirect response mechanism.

Finally, results reported in this thesis indicate that westward propagating signals may play an important role on the transient seasonal component of the AMOC and MHT. This is because seasonal changes in the FC transport are largely associated with the MHT at 26°N [Böning and Budich, 1991]. In fact, analysis developed in this study showed that the transient component of the FC transport can account for at least 30% of the transient seasonal variability of the AMOC and MHT (Figure 1.3), with seasonal anomalies reaching values as large as 5 Sv for the AMOC, and 0.4 PW for the MHT. Values obtained here are overall consistent with values reported by Atkinson et al. [2010]. It is reasonable to consider then, a relationship between westward propagating signals and year-to-year changes in the seasonal variability of the AMOC and MHT. For instance, results presented here are largely complementary to findings from Clément et al. [2014], which reported that 42% of the variance of the AMOC trans-basin thermocline transport east of the Bahamas inferred from geostrophic calculations at 26.5°N can be attributed to first-mode variability associated with eddies and Rossby waves at periods of 80–250 days.
Chapter Four
Conclusion

This study focused on improving the understanding of the role of westward propagating signals for driving year-to-year changes in the seasonal variability of the FC transport. Year-to-year changes in the FC seasonal variability account for 27% of the total FC variance and are linked with the transient seasonal component of the FC transport, which is defined as variability with periods at 73–525 days that are associated with variable annual phase. The transient FC component is also closely associated with year-to-year changes in the seasonality of the MHT and MOC. Therefore, understanding the main sources and mechanisms driving this type of variability in the FC is important.

Controlled realistic numerical simulations were carried out using the Regional Ocean Modeling System (ROMS) for different sets of idealized experiments with and without background flows associated with the FC to access the different mechanisms in play. Numerical simulations performed here included sets initialized: (1) with single eddy with different sizes and intensity in the ocean interior without the background flow (SENS-E0x); (2) with single eddy experiments at different latitudinal locations with the background flow; and with eddy-full configurations in the ocean interior (3) with and (4) without background flows.

The main finding from this study is that westward propagating signals can cause stochastic seasonal variability in the FC transport by means of two main mechanisms: the direct forcing mechanism, and the indirect forcing mechanism. In the direct forcing
mechanism, westward propagating signals may drive a two stage response in the Florida Straits through the Northwest Providence Channel, in which the first stage is characterized by barotropic velocity anomalies linked with the Rossby wave field, and the second stage is linked with the development of baroclinic wall-jets. In the indirect forcing mechanism, as the name suggests, westward propagating signals originating in the open ocean play an indirect role in modulating the FC seasonal variability. Results showed that, as the westward propagating signals approach the western boundary, they can perturb the nonlinear and stochastic variability associated with the Gulf Stream downstream from the Florida Straits, which once perturbed, evolves into a different state. In this mechanism, westward propagating signals behave as perturbations analogous to the butterfly effect, which may drive the system to evolve into a completely different state. The perturbed Gulf Stream variability is then linked to the Florida Straits through coastally trapped signals that travel along the east U.S. coast. Velocity anomalies in the Florida Straits linked with the indirect forcing mechanism were characterized by baroclinic changes in the velocity field confined to the left side of the Straits. Results indicate that the in the real ocean, the indirect forcing mechanism may play a dominant role in linking the open ocean variability from westward propagating signals to the changes in the FC transport, since the flow from the AC may partially block and advect signals northward. However, given the large variability of the AC, which often is observed to halt or reverse, the direct response mechanism may also occasionally play an important role in linking the open ocean variability with changes in the FC transport.

Results reported here based on numerical simulations confirm findings from previous studies based on observations, which reported that year-to-year changes in the FC seasonality are largely linked with elevated levels of background stochastic variability [Meinen et al., 2010], and that the main source for such stochastic variability is provided by westward propagating signals originated in the open ocean [Domingues et al., 2016].
Numerical experiments developed here showed that westward propagating signals with wavelengths ranging between 300–700 km generate transport anomalies at the Florida Straits with dominant time-scales of \(\sim\)200–400 days, and are essentially associated with stochastic seasonal variability. In other words, the variability generated by these signals is not characterized by a fixed annual phase or well-defined annual cycle.

It should be acknowledged here that, even though results from numerical simulations performed here showed that westward propagating signals can indeed modulate the FC transport causing stochastic seasonal variability, the simulated amplitude of the response (\(\sim\)2 Sv) is about 25% from the amplitude of the FC transient component observed in the real ocean (\(\sim\)8 Sv). This discrepancy, however, is probably due to the nature of the mechanisms revealed here, and with some inherent limitations in our numerical experiments. Results showed that westward propagating signals modulate the FC transport mostly through the indirect forcing mechanism, which is likely the dominant mechanism in the real ocean. In this mechanism, the magnitude of the variability is not likely dependent on the intensity of westward propagating signals, it depends solely on the intrinsic FC variability. The intrinsic FC was largely underestimated by the controlled environment adopted for the experiments developed in this study. For example, the use of constant boundary conditions at all boundaries and no wind forcing implied that important processes causing variability in the FC transport were not included in our simulations. Nevertheless, the mechanism reported here is consistent with observations reported by previous studies [Domingues et al., 2016], which showed that the phase of the transient component of the FC transport was indeed strongly correlated with the phase of westward propagating signals reaching the boundary. Therefore, it is likely that under more realistic ocean conditions, westward propagating signals can also modulate the amplified FC variability forced by other processes that were not included in the simulations developed here.
In conclusion, results reported in this study provide new insight on the understanding of mechanisms driving the Florida Current and Gulf Stream variability, and may be also applicable to other western boundary current systems. In addition, the mechanisms studied here can be potentially linked to year-to-year changes in the seasonality of the MOC and MHT, given that the FC corresponds to an important component of the MOC and MHT. This study also shows how the application of numerical simulations based on controlled realistic simulations can provide valuable information about the underlying dynamics of western boundary currents. More specifically, the approach and techniques employed here can also be applied to other areas and adjusted to investigate the influence of other processes and time-scales in the variability of western boundary currents.
References


Huthnance, J. (1978), On coastal trapped wave response to wind over the deep ocean, Ocean Modelling, 17, 1–3.


