Decadal Variability of the Atlantic Meridional Overturning Circulation

by

Martha Weaver Buckley

Submitted to the Department of Earth, Atmospheric and Planetary Sciences
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Author . . .: ........................................

Department of Earth, Atmospheric and Planetary Sciences

May 19, 2011

Certified by . . .: ........................................

John Marshall

Cecil and Ida Green Professor of Oceanography

Thesis Supervisor

Accepted by . . .: ........................................

Maria Zuber

E.A. Griswold Professor of Geophysics, Head of the Department
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Abstract

In the mean, the Atlantic Ocean transports 1 to 1.5 PW of heat northward, and estimates suggest that 60% of this heat transport is associated with a circulation that reaches the cold waters of the abyss. Due to the role of the Atlantic Meridional Overturning Circulation (AMOC) in ocean heat transport, numerous studies have suggested that AMOC variability plays a role in climate variability on a wide range of timescales. My focus is AMOC and ocean buoyancy variability on decadal timescales.

Decadal variability of sea surface temperature (SST) has been observed in the instrumental record and climate proxy data and is thought to be linked to variability in the AMOC. On the other hand, according to the thermal wind relation, buoyancy anomalies on the boundaries lead to anomalies in the AMOC, a fact that has been utilized in order to reconstruct the MOC at 26.5°N using data collected by the RAPID array. Here, I study decadal AMOC and buoyancy variability in a coupled and ocean-only GCMs run in idealized geometries. I focus on understanding the mechanisms of decadal variability of the AMOC, both the role of the AMOC in creating decadal buoyancy anomalies and the response of the AMOC to buoyancy anomalies.

I find that decadal AMOC variability is driven by buoyancy anomalies near the western boundary of the subpolar gyre. When a buoyancy anomaly hits the western boundary, it is advected southward by the deep western boundary current. Via the thermal wind relation, buoyancy anomalies on the boundaries result in anomalies in the shear of the zonally integrated meridional velocity. Buoyancy anomalies on the eastern boundary are observed to be negligible, except in the subpolar gyre, indicating that negative (positive) buoyancy anomalies on the western boundary lead to a spin up (down) of the AMOC. The AMOC is observed to respond passively to buoyancy anomalies on the western boundary: although variability of the AMOC does lead to variability in the meridional transport of heat and salt, these transports are not responsible for creating the buoyancy anomalies on the western boundary that drive the AMOC variability.
While the structure of the buoyancy anomalies is found to change with model bathymetry, in all the models studied the buoyancy variability is due to an ocean-only mode. In some cases, the mode is weakly damped (large Q-factor), resulting in regular, predictable oscillations. In other cases, the ocean-only mode is highly damped (small Q-factor) and must be excited by stochastic atmospheric variability, resulting in irregular, less predictable variability.

In nature, buoyancy anomalies along the western boundary might be created in a number ways, including local baroclinic instability, baroclinic Rossby waves impinging on the western boundary, advection of anomalies from tropics, and advection/propagation of convectively created anomalies from polar regions. In our models the dominant sources of buoyancy anomalies on the western boundary are local baroclinic instability and the propagation of baroclinic Rossby waves originating near the eastern boundary. However, we expect the response of the AMOC to decadal buoyancy anomalies on the western boundary to be similar regardless of the origin of these buoyancy anomalies.

Thesis Supervisor: John Marshall
Title: Cecil and Ida Green Professor of Oceanography
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Everything should be made as simple as possible, but not simpler.

-Albert Einstein
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Chapter 1

Introduction

1.1 Motivation for studying decadal variability

Numerical weather prediction (NWP) can successfully predict weather on daily to weekly timescales. Global coupled class models, such as those used in the IPCC projections, provide estimates of “forced” climate change on long timescales. However, variability in the climate system on decadal timescales is poorly understood. Due to the shortness of the observational record, particularly for the interior of the ocean, it is difficult to distinguish between decadal variability and secular trends. Additionally, mechanisms of decadal variability are poorly understood. In a recent paper, Zhang et al. (2007) used a climate model to demonstrate that changes in radiative forcing or variability of the Atlantic ocean circulation are both capable of producing the observed (detrended) variability in global sea surface temperatures.

Although decadal variability of the climate is poorly understood, understanding and perhaps predicting decadal climate variability is of great importance. Decadal variations in sea surface temperatures may lead to changes in atmospheric temperature, rainfall, and other societally relevant climate variables. For example, current models suggest that decadal changes in Atlantic sea-surface temperatures play a role in decadal hurricane variability (Zhang and Delworth, 2006) and global precipitation changes, including droughts (Danabasoglu, 2007). If predicting decadal climate variability is possible, it will have significant implications for long-range forecasting and
societal decision making.

Understanding decadal climate variability is essential in order to distinguish between anthropogenically induced warming and natural variability in the climate system. Additionally, human-induced warming of the climate system in the future will be modulated by natural climate variability. For example, Keenlyside et al. (2008) projected that global surface temperatures may not increase in the next decade, as natural climate variations in the North Atlantic and tropical Pacific temporarily offset the projected anthropogenic warming. However, a warm period due to natural variability has the potential to exacerbate the effects of human-induced climate change, and these effects must be understood when accessing the potential risks of climate change.

Global warming may also alter both the mean state and the variability of the climate system. For example, models run under global warming scenarios (Bryan et al., 2006) and limited data (Bryden et al., 2005; Wunsch and Heimbach, 2006) suggest that the Atlantic meridional overturning circulation (AMOC) has decreased in strength and may further decrease in strength as the climate warms. However, it is difficult (and perhaps impossible using currently available data) to distinguish between a secular trend and natural variability (Cunningham et al., 2007; Balmaseda et al., 2007). Only by understanding the mechanisms of climate variability, will we be able to predict how variability may change in a warming world.

In this work, I will focus on decadal variability of the Atlantic Ocean. While low-frequency variability in the Pacific Ocean is dominated by the interannual ENSO signal, evidence suggests that the Atlantic Ocean may have significant variability on decadal timescales. Additionally, the Atlantic Ocean is the best observed region of the ocean, allowing some estimates of variability on decadal timescales from observations.

### 1.2 Overview of Atlantic decadal variability

On short timescales, to first order the ocean responds passively to atmospheric forcing, reddening the essentially white spectrum of atmospheric variability (Hasselman,
1976). In contrast, both observations (Deser and Blackmon, 1993; Kushnir, 1994) and models (Dong et al., 2007) show that upper ocean heat content anomalies on decadal timescales are forced by ocean heat transport convergence and damped by air-sea heat fluxes, creating potentially important atmospheric temperature and circulation anomalies. Furthermore, both observations (Cunningham, 2010) and models (Jayne and Marotzke, 2001) suggest that variability in the overturning circulation plays a dominant role in meridional ocean heat transport variability. As a result, the deep inter-hemispheric meridional overturning circulation in the Atlantic is believed to play a role in setting sea surface temperatures on decadal timescales (Kushnir, 1994; Delworth and Mann, 2000; Dong and Sutton, 2003; Shaffrey and Sutton, 2004, 2006; Zhang, 2007; Danabasoglu, 2008).

The potential role of variability of the Atlantic Meridional Overturning Circulation (AMOC) in decadal sea surface temperature (SST) variability has lead to numerous studies of AMOC variability, including attempts to characterize variability of the AMOC from observations and explorations of AMOC variability in models ranging in complexity from box models to IPCC class GCMs. However, a theoretical framework for understanding the relationship between decadal buoyancy anomalies and AMOC variability is lacking. The crux of the problem is that the AMOC responds to buoyancy anomalies on the boundaries of the ocean basin according to the thermal wind relation, while meridional heat and freshwater transports due to AMOC variability may be responsible for creating decadal buoyancy anomalies. Teasing apart the response of the AMOC to buoyancy anomalies and its role in creating buoyancy anomalies is difficult, and as a result many studies focus on either the role the AMOC plays in creating buoyancy anomalies (see section 1.2.1) or the response of the AMOC to buoyancy anomalies (see section 1.2.2). Studies that focus on both the response of the AMOC to buoyancy anomalies and the role of the AMOC in creating buoyancy anomalies (see section 1.2.4) tend to represent the AMOC in very heuristic ways, which may not be justified.
1.2.1 Decadal buoyancy anomalies due to AMOC variability

In the mean the ocean transports approximately 2 PW of heat polewards (Trenberth and Caron, 2001; Wunsch, 2005), as can be seen from Figure 1-1. Meridional overturning circulations rather than horizontal circulations dominate the mean ocean heat transport (OHT) since vertical circulations act over large surface to deep temperature differences rather than smaller zonal temperature differences (Hall and Bryden, 1982; Roemmich and Wunsch, 1985; Cunningham, 2010). The use of heat transport streamfunctions has allowed the vertical dependence of OHT to be studied (Boccaletti et al., 2005; Ferrari and Ferreira, submitted). In the Atlantic, OHT is dominated by an overturning cell that includes both surface circulations and a deep circulation associated with North Atlantic Deep Water. As a result of this deep, inter-hemispheric circulation (see Figure 1-2), the Atlantic Ocean transports heat northward in both hemispheres, with a peak OHT of 1-1.5 PW at 20°N. Thus, the deep meridional overturning circulation in the Atlantic plays a role in maintaining the current mean climate. As a result, it has been suggested that variability of the AMOC may play a role in climate variability on a variety of timescales. Here I will focus on the potential role of the AMOC in decadal climate variability.

Kushnir (1994) first suggested that AMOC variability might play a role in creating decadal SST anomalies in the Atlantic. Kushnir analyzed SST and sea level pressure (SLP) variability in the Atlantic. He concluded that while interannual SST variability is primarily driven by air-sea heat fluxes forced by atmospheric variability, ocean dynamics plays a role in setting SST on decadal timescales. The decadal SST anomalies found by Kushnir were generally one polarity over the entire basin and maximal in the area around Iceland, in the Labrador Sea, and northeast of Bermuda. Maximal anomalies in these dynamically active regions suggested that these anomalies were due to changes in the large-scale ocean circulation, and Kushnir hypothesized that the decadal SST anomalies were due to variations in the strength of the AMOC. A more intense AMOC implies increased advection of warm waters northward at the ocean’s surface and warm conditions in the areas associated with the sinking branch
Figure 1-1: Mean meridional ocean heat transport (PW) as a function of latitude for the global ocean, the Pacific, the Atlantic, and the Indian oceans. The ocean heat transports are calculated from surface fluxes using (top panel) the NCEP and (bottom panel) the ECMWF atmospheric fields. Figure reproduced from Trenberth and Caron (2001).
Figure 1-2: Mean Atlantic MOC (Sv) from the GECCO state estimate over the 50 year period 1952-2001. Figure reproduced from Kohl and Stammer (2008).

of the overturning circulation. Similar cold-warm patterns in the North Atlantic were found by Mann and Park (1994, 1996) and Kushnir et al. (1997) using joint SST and SLP multivariate frequency domain methods. Kushnir et al. (1997) isolated a spectral peak at a timescale of 50-60 years, although the peak is not clearly resolvable from secular variations in their 136 year timeseries. The spatial patterns of SST and SLP variability associated with the multidecadal peak is shown in Figure 1-3.

The inability to isolate a multidecadal SST signal in the relatively short instrumental record inspired Mann and his colleagues to look at multidecadal SST variability in climate proxy data. Mann et al. (1995, 1998) demonstrated that a significant multidecadal signal is present in global SST proxy data and the variability has gross spatial features similar to those observed in the instrumental record.

The Atlantic Multidecadal Oscillation

The multidecadal, basin-wide fluctuation of SST observed in the instrumental record and climate proxy data was termed the “Atlantic Multidecadal Oscillation (AMO)” by Kerr (2000). Numerous studies have used observations to compute AMO indices,
generally defined as a detrended area-weighted mean of SST over the North Atlantic, smoothed to removed high frequency variability. Figure 1-4 shows an AMO index computed using the HadISST data set, as well as the surface temperature patterns associated with the AMO index (Knight et al., 2005). A high AMO index is associated with warm surface temperatures over the North Atlantic and much of the Northern Hemisphere.

However, attributing the detrended SST variability to internal variability of the ocean may be problematic as the radiatively forced signal is not well-described by a linear trend. An alternative approach for extracting the signal of natural variability is to use the global mean surface temperature as a proxy for the externally forced signal (Trenberth and Shea, 2006; Mann and Emanuel, 2006). Figure 1-5 compares an estimate of the AMO index computed by removing a linear trend from the average North Atlantic SST (top panel) to one computed by removing the signal associated with the global average SST anomaly (SSTg, middle panel) or global average surface temperature anomaly (Tg, bottom panel). The amplitude of the AMO index is significantly smaller, particularly for recent years, when the SST anomaly associated with SSTg or Tg is used rather than a linear trend. It is unclear exactly which method of removing the forced signal is most accurate, demonstrating the difficulty of separating forced and internal variability in observations.

**Estimates of AMOC and OHT Variability on Decadal Timescales**

Despite numerous studies that implicate AMOC variability and the resulting OHT variability in creating decadal buoyancy anomalies, observations of AMOC and OHT variability are extremely limited. Repeated hydrographic sections at several latitudes in the North Atlantic (Koltermann et al., 1999; Bryden et al., 2005) have been used to estimate changes in the AMOC, but the sparseness of these observations makes it difficult to determine whether observed changes are due to secular trends or variability on shorter timescales.

One method for estimating changes in the AMOC despite limited data is using state estimates, which combine the available data with an ocean circulation model
Figure 1-3: Multidecadal SST (colors) and SLP (contours) anomalies in the Atlantic from Kushnir et al. (1997). The spatial patterns were computed using a multivariate frequency domain singular value decomposition analysis of SST and SLP. The multidecadal peak (period of approximately 50 years) is significant at the 90% level.
Figure 1-4: (A) AMO index derived from detrended area-weighted mean of North Atlantic SST from HadISST data set (Rayner et al., 2003). Index is smoothed using a Chebyshev filter with half-power at a period of 13.3 years. (B) Surface temperature anomalies associated with one positive standard deviation of the AMO index, calculated by regression of surface temperatures with normalized AMO index. Combined land and SST data are from an optimally interpolated version of HadCRUTv data set (Jones et al., 2001). The solid contour bounds regions significant at the 90% level using a two-sided t-test. Figure reproduced from Knight et al. (2005).
Figure 1-5: Multiple methods for attempting to remove the forced signal from North Atlantic basin average SST anomaly (NASSTI). (a) Linear trend (black line) and detrended NASSTI (shaded). (b) An alternative approach is to use the global mean SST (SSTg) as a proxy for the externally forced signal, and regress the SST field onto SSTg. The basin average of the regression (black line) is an estimate of the forced signal and the local difference between the total field and the regression pattern (shaded) is an estimate of the internal variability. (c) Same as (b) but the global average surface temperature (Tg) is used instead of SSTg to estimate the anthropogenically forced variability. Figure reproduced from Ting et al. (2009).
using its adjoint. Using the 14 year ECCO state estimate, Wunsch and Heinbach (2006) concluded that the upper ocean northward flow decreased in strength while both the mid-depth return flow and northward bottom water transport increased in strength over the period 1992-2004.

An analysis of the GECCO state estimate (mean AMOC in GECCO is shown in Figure 1-2) shows a general increase in the strength of the AMOC at 25°N over the period 1952-2001 (see Figure 1-7). Figure 1-6 shows the spatial pattern of AMOC variability. AMOC variability with a standard deviation of over 10 Sv is observed in the 50 year record (top panel). The AMOC variability is dominated by variability on intra-annual timescales (middle panel), but an analysis yearly mean data demonstrates that the AMOC has interannual variability with a standard deviation of 1 Sv (bottom panel).

Observations of OHT variability are even more sparse (in both space and time) than observations of MOC variability. However, limited measurements (Johns et al., 2010; Fillenbaum et al., 1997), ocean state estimates (Wunsch and Heimbach, 2006), and models (Jayne and Marotzke, 2001) indicate that velocity variations advecting mean temperature gradients \( (v'T) \) account for the vast majority of OHT variability in the North Atlantic. Variability in vertical circulations is expected to make a larger contribution to OHT variability since mean vertical temperature gradients in the ocean are larger than horizontal temperature gradients. Therefore, MOC variability is expected to be a good indicator of meridional OHT variability.

**Modeling studies**

The hypothesized link between AMOC variability and SST variability on decadal timescales has prompted numerous modeling studies. A large number of climate models exhibit decadal variability of the AMOC (Delworth et al., 1993; Delworth and Greatbatch, 2000; Dong and Sutton, 2005; Knight et al., 2005; Zhang, 2008, 2010; Danabasoglu, 2008), although the timescale of the variability varies substantially between models. Figures 1-8a and 1-10a show timeseries of AMOC indices for the HadCM3 (Knight et al., 2005) and GFDL CM2.1 models, respectively. Powerspectra
Figure 1-6: Standard deviation of the AMOC from the GECCO state estimate. The top panel shows the standard deviation and the middle and bottom panels show the standard deviation on intra-annual and inter-annual timescales, respectively. Figure reproduced from Kohl and Stammer (2008).
Figure 1-7: Maximum MOC (1 year running mean) at 25°N from the GECCO state estimate for the reference (MITgcm with no assimilation of observations, green), the 50-year optimization (red), and an earlier 11 year optimization (black). Yellow shading indicates one standard deviation of the intra-annual variability, which is 2.4 Sv in the model. The difference between the reference state and the 50 year optimization is shown in blue (scale on the RHS axis). The asterisks show the values of the AMOC at 1000 m depth estimated by Bryden et al. (2005). Figure reproduced from Kohl and Stammer (2008).
of the AMOC timeseries indicate a broad peak at a timescale of 100 years in the HADCM3 model and a red spectra with a significant peak at a timescale of 20 years in CM2.1. Models demonstrate that decadal ocean heat transport (OHT) variability is associated with MOC variability. In Figure 1-8a, the OHT at 30°N is compared to the MOC index, and it is apparent that the two are highly correlated.

Models show decadal SST anomalies are associated with AMOC variability, although the spatial patterns of SST variability differ substantially between models. For example, in HadCM3 a relatively uniform warming of SST and surface air temperature over much of the Northern hemisphere (See Figure 1-9) is associated with a strong AMOC. On the other hand, in CM2.1 warming of the subpolar gyre and cooling along the Gulf Stream path are associated with a strong AMOC (see Figure 1-11). Yet, in both models the authors suggest that the observed decadal SST anomalies are due to changes in the strength of the AMOC.

**Implications for Predictability**

The idea that variability of the AMOC is responsible for decadal SST variability has provoked much interest in the AMOC. If AMOC variability has predictability on decadal timescales, SST and hence important climate variables may have a predictable component on decadal timescales. Msadek et al. (2010) conducted a series of perfect model predictability experiments using the GFDL CM2.1 model. The experiments were conducted by randomly selecting initial conditions from the control run, and for each initial condition conducting an ensemble of experiments that have the same ocean, land, and sea-ice conditions but perturbed atmospheric conditions. Msadek et al. (2010) concluded that the AMOC is potentially predictable for up to 20 years, but the degree of predictability of the AMOC depends on its initial state, with higher predictability when the experiments are initialized at either a maximum or minimum of the AMOC. However, due to the drastically different spectra of AMOC variability seen in models, the degree of predictability of the AMOC is almost certainly model dependent.
Figure 1-8: (a) Decadal mean AMOC index (black) and meridional heat transport at 30°N (red) in HadCM3. The AMOC index is defined as the maximum of the meridional streamfunction at 30°N. (b) Wavelet analysis of the annual mean AMOC index. (c) Power spectrum of the annual mean AMOC index. Dashed lines show the 95% confidence intervals. Figure from Knight et al. (2005).

Figure 1-9: Pattern of decadal surface temperature (left) and AMOC anomalies (right) in HadCM3. The patterns were computed using a multivariate frequency domain analysis of decadal mean surface temperature and AMOC anomalies for years 400-900 of a 1400 year run of the HadCM3 model. The black contours on the left figure show the climatological mean AMOC. Figure from Knight et al. (2005).
Figure 1-10: (a) Time series of the AMOC maximum (0 – 60°N, 500-5000m depth) in the control run of GFDL CM2.1. (b) AMOC power spectrum calculated with the multitaper method using four windows. A fitted red noise spectrum (thin line) and the associated 95% confidence interval are shown. (c) AMOC autocorrelation in CM2.1 (plain line) and autocorrelation function of an AR(1) model (dashed line). Shading indicates the 95% significance level. From Msadek et al. (2010)

Figure 1-11: Regression of subsurface (400m depth) temperature anomalies (left panel) and AMOC anomalies (right panel) on the AMOC index from the 1000-year control integration of GFDL CM2.1. The regressions correspond to a 1 standard deviation anomaly of the AMOC index, which has a standard deviation of 1.8 Sv. Left panel is reproduced from Zhang (2008) and right panel is courtesy of Rong Zhang.
1.2.2 Response of the MOC to buoyancy anomalies

While the focus of the previous section was on decadal buoyancy (or temperature) anomalies created by AMOC variability, buoyancy anomalies on the boundaries result in AMOC anomalies in accord with the (zonally integrated) thermal wind relation:

$$\frac{\partial V'}{\partial z} = \frac{1}{f} (b'_e - b'_w). \quad (1.1)$$

$V'$ is the zonally integrated meridional velocity anomaly, $f$ is the Coriolis parameter, and $b'_e$ and $b'_w$ are the buoyancy anomalies on the eastern and western boundaries, respectively. Thus, anomalies in the buoyancy difference between the eastern and western boundaries result in anomalies in the shear of the zonally integrated meridional velocity.

However, the thermal wind component is just one contribution to anomalies in the meridional overturning streamfunction $\Psi'$. Following Lee and Marotzke (1998) $\Psi$ (dropping primes) can be split into contributions related to the barotropic (depth averaged) velocity $\bar{v}$, the geostrophic shear (or thermal wind) $v_{sh}$, and the Ekman transport $v_{ek}$:

$$\Psi(y, z) = \int_{-H}^{z} \int_{x_w}^{x_e} \bar{v} dz dx + \int_{-H}^{z} \int_{x_w}^{x_e} (v_{ek} - \bar{v}) dz dx + \int_{-H}^{z} \int_{x_w}^{x_e} v_{sh} dz dx, \quad (1.2)$$

where $x_w$ and $x_e$ are the western and eastern limits of the basin, respectively and $H(x, y)$ is the depth of the ocean. The first term on the right hand side is the contribution that the barotropic flow makes to the MOC in the presence of topography, and it is often referred to as the external mode. It is the most difficult term to measure in the ocean. The second term on the right hand side is the contribution due to Ekman transport:

$$v_{ek} = -\frac{1}{\rho_0 f \delta z} \tau^x, \quad (1.3)$$

where $\rho_0$ is a reference density, $f$ is the Coriolis parameter, and $\tau^x$ is the zonal surface windstress. The Ekman layer transport is assumed to occur in a layer of thickness $\delta z$. 


and is compensated by a barotropic return flow:

$$ u_{ck} = -\frac{1}{\rho_o f H} \tau^x. $$

The third term on the right hand side is the contribution of the geostrophic shear to the MOC, which can be related to the buoyancy field on the boundaries according to the thermal wind relation (Equation (1.1)).

Although the focus here is on MOC anomalies, Equation (1.2) holds for both the mean and anomalies. However, the assumption that the Ekman transport is returned by a barotropic return flow is only true on short timescale and does apply for the mean Ekman transport (Jayne and Marotzke, 2001). This framework has been used to reconstruct the AMOC at 26°N, as described in the following section.

### 1.2.3 The RAPID-MOCHA Array

The RAPID-MOCHA\(^1\) program has monitored the AMOC at 26.5°N continuously since March 2004. Since only 6 years of data are currently available, the RAPID data obviously cannot be used to discuss decadal variability of the AMOC. However, no discussion of the AMOC would be complete without a mention of the RAPID-MOCHA array. Additionally, the RAPID-MOCHA array demonstrates that the AMOC can be reconstructed using boundary current transports, the windstress, and the thermal wind relation, according to Equation (1.2). These techniques can certainly be used to diagnose changes of the AMOC on longer timescales.

The basic principle of the array is to estimate the zonally integrated geostrophic northward velocity profile in the interior using measurements of temperature and salinity throughout the water column on the eastern and western boundaries. Moorings are also included on the flanks of the Mid-Atlantic Ridge to resolve flows in either sub-basin. Variability of the Gulf Stream flow is derived from cable voltage measurements across the Straits of Florida. Ekman transports are derived from satel-

\(^1\)Meridional Overturning Circulation and Heatflux Array (MOCHA) is a collaborate project, partnered with the UK RAPID program, to measure the MOC and ocean heat transport in the North Atlantic.
lite winds. Precision bottom pressure gauges are also employed to monitor absolute transports, including the barotropic circulation. The Ekman, geostrophic, and direct current observations are combined with an overall mass conservation constraint to continuously estimate the basin-wide AMOC strength and vertical structure.

Figure 1-12 shows a reconstruction of the maximum meridional overturning at 26.5°N (red line) for March 2004 to March 2008. The maximum overturning is the sum of the Gulf Stream transport (blue), Ekman transport (black), and upper mid-ocean transport (magenta). The Gulf Stream transport is based on electromagnetic cable measurements across the Florida Straits. The Ekman transport is based on Quick-Scat winds. The upper mid-ocean transport is calculated via the thermal wind relation using the RAPID mooring array data and is the vertical integral of the transport over the northward flowing layer (approximately the top 1100 m).

Figure 1-13 shows an estimate of the OHT (top panel) and MOC (bottom panel) at 26°N using data from the RAPID-MOCHA array and satellite SST and wind fields (Johns et al., 2010). The gray lines show the total transports and the black lines show the transport due to the geostrophic circulation, after the direct influence of the Ekman transport has been removed. The mean OHT across 26.5°N is 1.35 PW, but the OHT varies greatly, ranging from 0.2 to 2.5 PW over the 3 year timeseries. Much of the high-frequency variability of the OHT is due to Ekman transport variability, but significant geostrophic variability is present, even on short timescales. The variability of the OHT and the AMOC are highly correlated, suggesting that OHT variability is primarily due to velocity variations due to AMOC variability.

1.2.4 Heuristic models of decadal AMOC and buoyancy variability

A number of studies have attempted to create simple models of AMOC variability using, for example, box models. Most of these models rely on some sort of delayed negative feedback to create oscillations in the strength of the AMOC. These models are useful because they provide a simplified setting in which the processes of AMOC
Figure 1-12: Reconstruction of the maximum meridional overturning at 26.5°N (red line) for March 2004 to March 2008. The maximum overturning is the sum of the Gulf Stream transport (blue), Ekman transport (black), and upper mid-ocean transport (magenta). The Gulf Stream transport is based on electromagnetic cable measurements across the Florida Straits. The Ekman transport is based on Quick-Scat winds. The upper mid-ocean transport is calculated via the thermal wind relation using the RAPID mooring array data and is the vertical integral of the transport over the northward flowing layer (approximately the top 1100 m). Figure courtesy of the RAPID Program (Baringer et al., 2011).
Figure 1-13: Timeseries of meridional heat transport (top) and maximum value of the MOC (bottom) at 26.6°N. Light lines show the total variability and heavy lines show the variability attributed to the geostrophic circulation after the direct influence of Ekman transport is removed. Figure reproduced from Johns et al. (2010).
and buoyancy variability may be explored, allowing mechanisms of variability to be isolated. However, due to their simplicity, these models tend to represent the AMOC in a very heuristic way.

Griffies and Tzipperman (1995) showed that a simple four-box ocean model run under mixed boundary conditions exhibits damped interdecadal oscillations in the strength of the MOC. In their model the transport between the northern box and the southern box depends on the buoyancy difference between the boxes. A strengthened MOC leads to increased salt and heat transport into the sinking region. The salt anomalies decrease the buoyancy in the sinking region, providing a positive feedback on the MOC (salinity-advection feedback), while the temperature anomalies increase buoyancy and provide a negative feedback (temperature-advection feedback). However, the temperature anomalies are initially damped, and thus the negative feedback is delayed, leading to an oscillation in the strength of the MOC.

However, ocean circulation models run with mixed boundary conditions may exhibit unrealistically large variability of the MOC because atmospheric temperatures are not allowed to adjust to SST, leading to unrealistically large damping of large-scale SST anomalies and an underestimate of the negative temperature-advection feedback (Rahmstorf and Willebrand, 1995). Additionally, the assumption that the strength of the MOC depends on the buoyancy difference between the northern and southern boxes is not consistent with the thermal wind relation and may not be justified.

Marshall et al. (2001a) proposed a simple four box model to consider the effects of air-sea coupling and advection by ocean currents on SST anomalies. The model has two boxes in each the atmosphere and the ocean, and the boxes are separated by the mean zero windstress curl line, which separates the subtropical and subpolar gyres. The advantage of this choice of boxes is that (in the model) there are no mean currents between the boxes in either the atmosphere or the ocean so only anomalous currents (the so-called “intergyre gyre”) can transport heat between the boxes and create SST anomalies.

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2 Although in the absence of bathymetry, the barotropic streamfunction at the zero curl line will be zero, there may be baroclinic transport between the gyres, which may transport heat. This is not included in the model.
Assuming that heat storage in the atmosphere is negligible, $T$ the temperature (or buoyancy) anomaly in the northern box, can be written in non-dimensional form as:

$$\frac{\partial T}{\partial t} = m\Psi_m + g\Psi_g + \chi T + F_T$$  \hspace{1cm} (1.4)

$$-\frac{\partial \Psi_g}{\partial t} + \frac{\partial \Psi_g}{\partial x} = -\tau$$  \hspace{1cm} (1.5)

$$\tau = F_\tau - fT.$$  \hspace{1cm} (1.6)

$$F_T = -\alpha F_\tau$$  \hspace{1cm} (1.7)

$$\frac{\partial \Psi_m}{\partial t} = -sT$$  \hspace{1cm} (1.8)

Equation (1.4) describes the evolution of temperature anomalies, forced by stochastic thermal forcing $F_T$ and advection across the boundary between the boxes by the anomalous baroclinic gyre circulation just inside the western boundary $\Psi_g$ and the MOC $\Psi_m$ and damped on a timescale $\lambda^{-1}$. $g$ and $m$ are measures of the efficiency of the heat transport by the anomalous gyres and MOC, respectively. $\lambda$ includes both air-sea damping and Ekman effects, which provide a small positive feedback on SST anomalies forced by stochastic air-sea heat fluxes and thus effectively reduce the magnitude of the damping by air-sea heat fluxes. Equation (1.5) describes changes in the baroclinic streamfunction $\Psi_g$ forced by windstress anomalies $\tau$ according to the first baroclinic Rossby wave equation, described in detail by (Frankignoul et al., 1997). Equation (1.6) expresses the windstress forcing in terms of the stochastic wind forcing $F_\tau$ and the windstress forcing determined by atmosphere-ocean coupling. Equation (1.7) states that the stochastic turbulent air-sea heat flux is linearly related to the stochastic surface windstress, with $\alpha > 0$ implying that increased westerlies ($F_\tau > 0$) result in anomalous air-sea heat fluxes out of the ocean in the northern box. Equation (1.8) describes the MOC variability in the model, assuming that the strength of the MOC is instantaneously determined by the temperature difference between the two boxes; $s$ represents the efficiency of thermal dipoles in driving the MOC.

While the equations in this model are quite heuristic, particularly the equation for
the MOC, the advantage of this model is that it provides a framework for studying both thermal variability and variability of the ocean circulation. The role of thermal forcing, wind forcing, gyres, the MOC, and feedback in creating SST anomalies can be examined in a systematic way.

1.2.5 Role of convection

The strong vertical density gradients found in the ocean thermocline generally inhibit the vertical exchange of fluid between the surface and the abyss. However, in a few regions in the ocean, characterized by weak stratification and large air-sea buoyancy losses in winter, convection penetrating to a depth of several thousand meters occurs in wintertime. These regions are said to be “preconditioned” for convection. Not only must the stratification in these regions be weak (perhaps due to previous convection), the prevailing circulation must allow weakly stratified underlying waters to be brought to the surface so that they may be exposed to intense surface buoyancy forcing. This condition is favored by cyclonic circulations with isopycnals that dome up toward the surface (Marshall and Schott, 1999).

In the present climate, deep convection occurs only in the Atlantic (primarily in the Labrador and Greenland Seas) and in the Weddell Sea in the Southern Ocean. Deep convection and water mass formation in the North Atlantic is responsible for the deep MOC in the Atlantic. As a result, variability in deep convection and water mass formation is often cited as a cause of MOC variability. In this section I will review observational evidence of variability in convection and water mass formation and describe how this variability may be linked to AMOC variability.

Observations of changes in water mass volumes and properties indicate that the intensity of deep convection is highly variable from one year to the next (Dickson, 1997). For example, convection in the Labrador Sea was extremely weak and shallow in the 1960's, only penetrating to a depth of 1 km. From 1966 until 1995 convection in the Labrador Sea increased in strength, and by 1995 convection reached a depth of 2.3 km (Sy et al., 1998). Variability in deep convection may be due to changes in air-sea buoyancy fluxes as a result of variability of the atmospheric circulation,
such as the NAO. On the other hand, changes in deep convection may be the result of preconditioning. Changes in stratification due to previous convection, changes in the atmospheric circulation, or buoyancy anomalies moving into the region may precondition the ocean for enhanced convection. Additionally, the role of horizontal advection in re-stratifying the water column may affect the intensity of convection and the depth to which convection penetrates.

The hypothesis that changes in high latitude water mass formation lead to changes in the strength of the AMOC relies on the following assumptions: (1) The deep western boundary current (DWBC) is the dominant pathway for the export of water masses formed due to high-latitude convection. (2) Changes in water mass formation lead to changes in the strength of the DWBC. (3) Changes in the strength of the DWBC lead to changes in the strength and position of the Gulf Stream. The combined changes in the warm, northward flowing branch of the AMOC (Gulf Stream and North Atlantic Current) and the deep, southward flowing branch of the AMOC (DWBC) lead to changes in the strength of the AMOC. However, each one of these causal links is tentative, and I will briefly examine the evidence supporting (or refuting) each below.

(1) Observations along the continental slope in the Western North Atlantic, show the existence of a DWBC carrying dense, weakly stratified waters southward (Swallow and Worthington, 1961; Pickart and Jr., 1998; Schott et al., 2004; Joyce et al., 2005; Toole et al., 2011). However, recent float observations by Fischer and Schott (2002) and Bower et al. (2009) suggest that a large fraction of Labrador Sea Water (LSW) that reaches the subtropics does so through an interior pathway.

(2) The hypothesis that changes in high-latitude water mass formation lead to changes in the strength of the DWBC is complicated by the fact that weakly stratified water masses are generated by convection in several different regions in the North Atlantic. Changes in the formation rate or properties of each water mass may effect the AMOC differently. Additionally, the formation rates of each water mass do not necessarily vary in concert with each other, making it difficult to isolate the effects of changes in one water mass on the strength of the DWBC. For example, when there
is deep convection in the Labrador Sea, convection in the Nordic Seas tends to be shallow (Dickson, 1997).

A study of the relationship between water mass formation and the transport of the DWBC at Line W along the continental slope south of New England (near 40°N, 70°W) demonstrates that larger DWBC transports are associated with colder, fresher and more weakly stratified Overflow Water (OW) and more stratified deep Labrador Sea Water (LSW) (Pena-Molino, 2010). A theoretical analysis of potential vorticity (PV) anomalies shows both enhanced LSW production and OW production lead to strengthening of DWBC, but the overall response at line W integrates different contributions and is mostly driven by changes in deep OW rather than changes at intermediate depths. PV anomalies formed in the Labrador sea take approximately 6-9 years to reach Line W.

(3) Numerous modeling studies have shown that the presence of the DWBC affects the separation of the Gulf Stream from the coast and its subsequent path into the interior (Thompson and Schmitz, 1989; Spall, 1996; Zhang and Vallis, 2007). Zhang and Vallis (2007) showed that bottom vortex stretching induced by a down-slope DWBC south of the Grand Banks leads to the formation of a cyclonic northern recirculation gyre (NRG), the separation of the Gulf Stream from the coast downstream of Cape Hatteras, and thus a more southerly Gulf Stream path.

The relationship between the DWBC and the Gulf Stream path has lead to the hypothesis that meridional excursions in the Gulf Stream path might be due to variations in the strength of the DWBC. Observational (Joyce and Zhang, 2010; Pena-Molino, 2010; Toole et al., 2011) and modeling studies (Zhang, 2008) have found that southward displacements of the Gulf Stream are associated with a stronger DWBC, with changes in the DWBC leading changes in the Gulf Stream by several months (Joyce and Zhang, 2010).

The relationship between the Gulf Stream path and the strength of the AMOC is more tentative as long-term measurements of the AMOC are lacking. Zhang (2008) showed that in the GFDL CM2.1 model, a more southerly Gulf Stream path is associated with an increase in the strength of the AMOC. Additionally, Joyce and Zhang
used observations to show that the SST anomalies associated with a southerly Gulf Stream path have a similar spatial pattern to the SST anomalies associated with a strong AMOC in CM2.1, leading credence to Zhang’s modeling results.

Despite the tentative string of arguments theoretically linking variability in polar water mass formation to changes in the strength of the AMOC, variability of deep convection is often implicated as being the cause of AMOC variability. Numerous models show that variability of the AMOC is correlated with variability in deep convection (Dong and Sutton, 2005; Danabasoglu, 2008; Msadek et al., 2010) and lead/lag relations are often used to hypothesize that convective variability is driving AMOC variability. Whether this is indeed the case or both convection and the AMOC are responding to some other process, such as the North Atlantic Oscillation, is unclear.

1.3 Approach: GCMs run in idealized geometries

As outlined in the previous section, a theoretical framework for understanding decadal AMOC variability and associated buoyancy anomalies is lacking. In this thesis I will study decadal AMOC and buoyancy variability in coupled and ocean-only GCMs run in idealized geometries. By simplifying the geometry, I endeavor to create models that are simple enough that the mechanisms of decadal AMOC and buoyancy variability can be isolated, but are complex enough to exhibit interesting and perhaps realistic behavior. I will focus on understanding the mechanisms of decadal variability of the AMOC, both the role of the AMOC in creating decadal buoyancy anomalies and the response of the MOC to buoyancy anomalies.

In chapter 2 I will introduce the idealized GCMs which will be used to study decadal buoyancy and MOC variability. Additionally, I will briefly describe the mean state of the models, focusing on the aspects of the models that make them well-suited for studying decadal variability. In Chapter 3 I will show that the MOC and buoyancy in the subpolar gyre exhibit variability on decadal timescales and demonstrate that MOC variability is due to buoyancy anomalies on the boundaries in accord with the thermal wind relation. In Chapter 4 I will address the origin of the decadal MOC
and buoyancy anomalies. Specifically, I will answer the following questions: (1) Does large-scale MOC variability play an active role in creating buoyancy anomalies in the subpolar gyre? (2) Is the mode of buoyancy and MOC variability a coupled mode, a damped ocean-only mode, or a self-sustained ocean-only mode? Finally, in chapter 5 I will summarize the main conclusions of this work, focusing on the aspects of MOC and buoyancy variability that I believe are robust. I will make connections between the decadal MOC and buoyancy variability seen in my idealized models to more complex (and hopefully realistic) GCMs as well as data.
Chapter 2

Idealized Coupled Models

In this thesis I will study decadal MOC and buoyancy variability in coupled and ocean-only GCMs run in idealized geometries. By simplifying the geometry, I endeavor to create models that are simple enough that the mechanisms of decadal MOC and buoyancy variability can be isolated, but are complex enough to exhibit interesting and perhaps realistic behavior. In this chapter I will briefly describe the idealized models and their mean states, focusing on the aspects of the models that make them well suited for studying decadal climate variability.

2.1 The model

The model used in this study is the Massachusetts Institute of Technology GCM (MITgcm) run in a coupled atmosphere-ocean-sea ice setup as described in Marshall et al. (1997). The model has a realistic three dimensional atmosphere and ocean, but it is run in idealized geometry. The planet is covered entirely by water except for two ridges that extend from the north pole to 34°S, dividing the ocean into a small basin (almost 90° wide), a large basin (almost 270° wide), and a zonally unblocked southern ocean. As described in Ferreira et al. (2010) this idealized setup captures many of the gross features of the earth’s climate: a meridional asymmetry (circumpolar flow near the pole in the southern hemisphere and blocked flow in the northern hemisphere) and a zonal asymmetry (a small basin and a large basin). We will see that the small
basin resembles the Atlantic, and my focus will be on the small basin. The model will be run in two idealized bathymetries: one in which the bottom is flat and the ocean has depth of 3 km and the other in which bowl bathymetry is added to the small basin and the ocean depth varies from 3 km in the center of the basin to 2.5 km next to the meridional boundaries (See Figure 2-1).

![Flat and Bowl Bathymetry](image)

**Figure 2-1**: Ocean geometry and depth (km) for Flat (left) and Bowl (right). Two strips of land (white) extend from the north pole to 34°S, dividing the world ocean into a small basin, a large basin, and a zonally unblocked southern ocean. In Bowl bathymetry is added to the small basin and the ocean depth varies from 3 km in the center of the basin to 2.5 km next to the meridional boundaries.

Much of the following model description has been borrowed from Ferreira et al. (2010). The model exploits an isomorphism between the atmosphere and ocean dynamics in order to render atmospheric and oceanic models from one hydrodynamical core, as discussed in Marshall et al. (2004). The atmosphere and ocean are integrated forward on the same “cubed sphere” horizontal grid at C24 (each face of the cube is split in to a 24 x 24 matrix of cells) yielding a resolution of 3.7° at the equator. The use of the cubed sphere coordinates avoids problems associated with the converging meridians of a latitude-longitude grid and ensures that the model dynamics at the poles are treated with as much fidelity as elsewhere. The model uses the following (isomorphic) vertical coordinates: the rescaled pressure coordinate $p^*$ for the atmosphere and the rescaled height coordinate $z^*$ for the Boussinesq ocean (Adcroft and Campin, 2004).
The atmosphere is of “intermediate” complexity, employing the Simplified Parameterization, Primitive Equation Dynamics (SPEEDY) physics package described in Molteni (2003). The atmospheric model includes a four-band radiation scheme, a parameterization of moist convection, diagnostic clouds, and a boundary layer scheme. The atmospheric model has low vertical resolution, comprised of 5 vertical levels.

The ocean is has a maximum depth of 3 km and has 15 vertical levels, increasing from a thickness of 30 m at the surface to 400 m at depth. As eddies are not resolved by the low-resolution model, the effects of mesoscale eddies are parameterized as an advective process (Gent and McWilliams, 1990) and an isopycnal diffusion (Redi, 1982) with a transfer coefficient of 1200 m$^2$s$^{-1}$ for both processes. Convective adjustment, implemented as an enhanced vertical mixing of temperature and salinity, is used to represent ocean convection (Klinger et al., 1996). The background vertical diffusivity is uniform and set to $3 \times 10^{-5} m^2 s^{-1}$.

A two and a half layer thermodynamic ice model following Winton (2000) is incorporated into the model. There is no sea ice dynamics. The land model is a simple two-layer model with prognostic temperature, groundwater, and snow height. The land albedo is set to 0.25 except where snow is present in which case the albedo depends on the snow height, snow age, and surface temperature.

Orbital forcing and CO$_2$ levels are prescribed at present day values. The seasonal cycle is represented, but there is no diurnal cycle. The whole system is integrated forward on a parallel computer. Fluxes of momentum, heat, and freshwater are exchanged every hour (the ocean model time step). The model achieves perfect (machine accuracy) conservation of freshwater, heat, and salt during extending integrations, as discussed in Campin et al. (2008).

The flat bottomed model, launched from a state of rest with temperature and salinity distributions taken from a zonal-average ocean climatology, was integrated for 20000 years to a quasi-equilibrium (integration performed by David Ferreira). I picked up the model at the end of the 20000 year run, and initiated two additional 1000 year runs, one with flat bathymetry (henceforth Flat) and one with bowl bathymetry (henceforth Bowl). As the model takes some time to adjust to the changes
in bathymetry, I analyzed the last 800 years of these 1000 year runs.

2.2 Mean state

The mean state of Flat (the “Double Drake” model) is analyzed in detail in Ferreira et al. (2010). As the focus of this work is decadal variability, I will only briefly describe the mean state, focusing on the features of the model that make it suited for studying MOC variability and the changes to the mean state when bowl bathymetry is added to the model. All diagnostics discussed below are averages over the last 800 years of the 1000 year runs.

2.2.1 Sea surface climatologies

The mean SST and sea ice thickness in Flat are shown in the top two panels of Figure 2-2. (The mean SST and sea ice thickness in Bowl are almost identical.) In both models, the south pole is significantly colder than the north pole. The lack of meridional boundaries in the southern ocean reduces the ability of the ocean to transport heat poleward, and an ice cap forms around the south pole. Additionally, in both models SSTs in the small basin are warmer than in the large basin.

The mean sea surface salinity (SSS) and surface density in Flat are shown in the bottom panels of Figure 2-2. (Bowl is almost identical.) Like in the present climate, the equatorial regions and high latitudes are relatively fresh due to excess precipitation and the subtropics are salty due to excess evaporation. The small basin is saltier than the large basin, similar to the higher salinity of the Atlantic relative to the Pacific. Although SSTs are warmer in the small basin than the large basin, the surface density is higher in the small basin due to the higher SSS.

Due to the higher surface density in the small basin, deep convection is restricted to the small basin. The top left panel in Figure 2-3 shows the convective index averaged over the top 620 m for Flat. (Bowl is almost indistinguishable.) The convective index indicates the percent of the time over which convective adjustment is operative. Below 250 m, convection is confined to the small basin, where it reaches as deep as 1500 m.
about 15% of the time, typically one winter month per year.

![Figure 2-2: (Top) Mean SST and sea ice thickness in Flat. Mean SSS (bottom left) and surface density (bottom right) in Flat.](image)

### 2.2.2 Mean Ocean Circulation

#### Horizontal Circulation

The zonal mean zonal surface windstress, shown in left panel of Figure 2-4a, is easterly in the tropics, westerly in midlatitudes and easterly near the poles. This large scale pattern of windstress forces the ocean’s gyre circulation and subtropical overturning cells. The mean barotropic streamfunction for Flat and Bowl are shown in Figure 2-4. The patterns of the barotropic streamfunction $\Psi_{BT}$ can be understood using
Figure 2-3: (Top left) Mean convective index for the Flat model. The convective index indicates the fraction of the time that convective adjustment is active. (Top right) The mean meridional overturning streamfunction in the large basin for Flat. (Bottom panels) The mean meridional overturning streamfunction in the small basin for Flat (right) and Bowl (left). The black box shows the latitude and depth range used to define the MOC timeseries.
Sverdrup balance:
\[
\beta \frac{\partial \Psi_{PT}}{\partial x} = \frac{1}{\rho_o} \mathbf{z} \cdot \nabla \times \tau_{\text{wind}} - f w_h, \tag{2.1}
\]
where \( f \) is the Coriolis parameter, \( \beta \) is the meridional gradient of \( f \), \( \rho_o \) is a reference density, \( \tau_{\text{wind}} \) is the windstress, and \( w_h \) is the vertical velocity at the bottom of the ocean. In Flat, \( w_h = 0 \) and the streamfunction has zeros at the lines of zero windstress curl. In the northern hemisphere, the streamfunction is anticyclonic in the subtropical gyres and cyclonic in the subpolar gyres. There are also anticyclonic polar gyres in the northern hemisphere due to the presence of polar easterlies. In the southern hemisphere, there are anticyclonic subtropical gyres, but no subpolar gyres due to the lack of meridional boundaries below 34°S. When bathymetry is added to the small basin, the weak anticyclonic polar gyre vanishes and the cyclonic subpolar gyre extends to the poles (compare left and right panels of Figure 2-4b). Even though the windstress curl is negative in the polar region, \( w_h \) is also negative and has a larger magnitude than the windstress curl term, resulting in a cyclonic barotropic streamfunction.

Figure 2-5 shows the mean currents in the small basin at the surface and at a depth of 1735 m for Flat and Bowl. The surface circulation is anticyclonic in the subtropical gyres and cyclonic in the subpolar gyre. At the surface the cyclonic subpolar gyre extends to the north pole in both Flat and Bowl. The strongest surface currents are found along the western boundaries, except in the subpolar gyre. In the subpolar gyre, the strongest currents are found on the eastern boundary, as the warm, northward flowing branch of the MOC continues from the western boundary of the subtropical gyre, into the interior, and along the eastern boundary of the subpolar gyre. This circulation is analogous to the North Atlantic current.

Deep water formation in the small basin feeds a deep western boundary current, which flows southward from 64°N to the exit of the small basin. North of 64°N, the maximum of the polar easterlies, the circulation at depth is in the opposite direction to the surface circulation. In Flat the anticyclonic deep circulation is stronger than the cyclonic surface circulation, creating a weak anticyclonic barotropic streamfunc-
tion, as is required by the negative windstress curl in this region. Highly baroclinic circulations such as these are a signature of buoyancy induced circulations.

**Meridional Overturning Circulation**

A result of deep water formation, a deep meridional overturning circulation develops in the small basin, similar to the overturning circulation in the present-day Atlantic Ocean. The residual-mean meridional overturning circulation (MOC) in the small basin for Flat and Bowl are plotted in the bottom panels of Figure 2-3. The majority of the water that sinks in the small basin is still at depth when it exits the small basin, supporting the idea that the water upwells primarily in the southern ocean. In contrast, the MOC in the large basin (see top right panel of Figure 2-3) is dominated by shallow wind-driven cells.

The mean meridional volume transport in the small basin as a function of depth is shown in Figure 2-6. The zonally integrated volume transport is shown in the top panels and in the lower panels the meridional volume transport is decomposed into the western boundary, interior, and eastern boundary contributions. Consistent with the mean MOC, the zonally integrated meridional volume transport is northward at the surface and southward at depth. Northward flow occurs in the interior of the southern hemisphere (SH) subtropical gyre and on the western boundary in the tropics and in the northern hemisphere (NH) subtropical gyre. In the NH subpolar gyre the northward flow occurs primarily along the eastern boundary. Southward flow at depth occurs along the western boundary, except north of 60°N where the southward transport in the deep layer is primarily along the eastern boundary.

The ocean heat and freshwater transports in the small and large basins of the model bear a striking similarity to those observed in the Atlantic and Indo-Pacific sectors of the modern climate (see top panels of Figure 2-7). Like in the Indo-Pacific, the heat transport in the large basin is due to the gyre circulations and Ekman transport and is poleward in both hemispheres. Similar to the Atlantic, the heat transport in the small basin is northward in both hemispheres.

Decomposing the temperature and meridional velocity into their zonal means,
Figure 2-4: (a) (Left) Average zonal mean zonal winds in Flat (black) and Bowl (blue). (Right) Average barotropic streamfunction in Flat. The black contours show the barotropic streamfunction over the southern ocean where the color-scale is saturated. (b) Average barotropic streamfunction in small basin in Flat (left) and Bowl (right). The thick black line marks the equator and the thin black lines mark the latitudes of the maxima/minima of the zonal windstress which predict the boundaries of the barotropic gyre circulations in the absence of bathymetry.
Figure 2-5: Mean horizontal currents in the small basin at the surface (top panels) and at a depth of 1735 m for Flat (left) and Bowl (right). Pink shading in the top panels indicates the regions where the mean meridional velocity is northward.
denoted by $[V]$ and $[T]$, and the deviations from their zonal means, denoted by $V^*$ and $T^*$, the mean meridional OHT can be separated into the contribution by the zonally averaged circulation $\rho_s C_P [V][T]$ and the deviations from the zonal average circulation $\rho_s C_P [V^*T^*]$ (see Figure 2-8). The majority of the OHT in the small basin is accomplished by the zonal mean overturning circulation, except in the subpolar gyre where the gyre circulation also makes a substantial contribution.

The bottom panels of Figure 2-7 show the meridional freshwater transport (FWT) in the model (left) and in observations (right). While the exact structure of the FWT is different in the model and the (quite uncertain) observations, they share many gross features. In the tropics and high-latitudes, where precipitation exceeds evaporation, the ocean freshwater transport is divergent. In the subtropics, where evaporation exceeds precipitation, the ocean freshwater transport is convergent. The ocean freshwater transport of the small basin/Atlantic is almost everywhere more northward than that of the large basin/Indo-Pacific.
Figure 2-6: The mean meridional volume transport in the small basin as a function of depth (Sv/m) for Flat (left panels) and Bowl (right panels). The zonally integrated volume transport is shown in the top panels and in the lower panels the meridional volume transport is decomposed into the western boundary, interior, and eastern boundary contributions.
Figure 2-7: (Top panels) Ocean heat transport in (left) the small and large basins of Flat (solid lines) and Bowl (dashed black lines) and (right) the Atlantic and Indo-Pacific sectors as calculated by Trenberth and Caron (2001). (Bottom Panels) Ocean freshwater transport in (left) the small and large basins of Flat (solid lines) and Bowl (dashed black line) and (right) the Atlantic and Indo-Pacific sectors, as calculated by Wijffels et al. (1992).
Figure 2-8: Meridional ocean heat transport (OHT) in the small basin for Flat (solid lines) and Bowl (dashed black lines). Total advective OHT (blue line) is decomposed into contributions by the zonal mean circulation (green line), and deviations for the zonal mean circulation (cyan line). Diffusive OHT is small (magenta line).
Chapter 3

Decadal Variability in the small basin

In the previous section I showed that the mean state of the Flat and Bowl models have many similarities to the present climate. Specifically, the small basin is saltier than the large basin and a deep-interhemispheric MOC develops in the small basin, which can be thought of as an idealized “Atlantic Ocean”. In this section I will analyze decadal variability of the small basin. I will show that the MOC and buoyancy in the subpolar gyre exhibit variability on decadal timescales and demonstrate that MOC variability is due to buoyancy anomalies on the boundaries in accord with the thermal wind relation.

3.1 Decadal MOC variability

The MOC in the box 8°S to 40°N, 460 to 1890 m depth (box shown in black in Figure 2-3) is used as a measure of the large-scale MOC variability. At each latitude, a timeseries of the MOC is computed by taking the value of the meridional streamfunction at the depth of the maximum of the mean meridional streamfunction within the box. These timeseries are then averaged over all the latitudes in the box to create a timeseries of the MOC in the box, henceforth called the MOC timeseries. The right panels of Figure 3-1 show a 100 year segment of the MOC timeseries for Flat (top
right) and the Bowl (bottom right).

The left panels of Figure 3-1 show the spatial pattern of the MOC variability obtained by projecting\textsuperscript{1} the meridional streamfunction anomalies onto the normalized\textsuperscript{2} MOC timeseries (henceforth called the MOC index). As expected, the spatial patterns of the MOC anomalies are inter-hemispheric, with a single sign MOC anomaly between 20°S and 60°N. The spatial patterns strongly resembles the first empirical orthogonal function (EOF) of the MOC, which explains 26% of the variance for Flat and 22% of the variance for Bowl. The correlation between the MOC index and the first principle component (PC) timeseries of the MOC is 0.94 for Flat and .87 for Bowl.

Figure 3-6 shows the powerspectra of the MOC indices. The powerspectrum of the MOC index in Flat (top) is red at high frequencies, has a large peak at a timescale of 34 years, and flattens out at low frequencies. The powerspectrum of the MOC index for Bowl (bottom) is red at high frequencies and flattens out at low frequencies.

Figure 3-2 shows the meridional volume transport anomalies in the small basin that are associated with the MOC variability for Flat (left panels) and Bowl (right panels). Spatial patterns are created by projecting the volume transport anomalies as a function of latitude and depth onto the MOC index at lag=0. The zonally integrated volume transport anomalies (top panels) are northward near the surface and southward at depth, in accord with the positive MOC anomaly (see Figure 3-1). Additionally, the meridional volume transport anomalies are decomposed into the western boundary (middle panels) and interior/eastern boundary (bottom panels) contributions. The northward meridional volume transport anomaly near the surface and southward transport anomaly at depth that lead to the positive MOC anomaly are primarily in the interior in the SH subtropical gyre, along the western boundary in the tropics and NH subtropical gyre, and along the eastern boundary in the NH subpolar gyre. The decomposition of the meridional volume transport anomalies matches that of the mean meridional volume transport (see Figure 2-6).

\textsuperscript{1}Projecting a data field onto a timeseries means computing the covariance between the timeseries and the data field at each spatial location.

\textsuperscript{2}A normalized timeseries has mean zero and standard deviation of one.
Figure 3-1: (Right panels) A 100 year segment of the timeseries of the MOC in the box 8° S to 40° N, 460 to 1890 m depth (box shown in black in Figure 2-3) for the Flat (top) and Bowl (bottom). (Left panels) The spatial patterns of the MOC variability obtained by projecting the MOC anomalies onto the MOC index for Flat (top) and Bowl (bottom).
Figure 3-2: Meridional volume transport anomalies in the small basin as a function of latitude and depth (Sv/km) for Flat (left panels) and Bowl (right panels). Spatial patterns are created by projecting the meridional volume transport anomalies onto the MOC index at lag=0. The zonally integrated volume transport anomalies are shown in the top panels. Additionally, the meridional volume transport anomalies are decomposed into the western boundary (middle panels) and interior/eastern boundary (bottom panels) contributions.
3.2 Decadal buoyancy anomalies

In this section I will describe the buoyancy variability in the small basin. I will demonstrate that decadal buoyancy anomalies are damped by air-sea buoyancy fluxes, and thus the ocean circulation must play an active role in creating decadal buoyancy anomalies.

The top panel of Figure 3-3 shows the first two empirical orthogonal functions (EOFs) of buoyancy at a depth of 265 m in the northern hemisphere (NH) of the small basin in Flat. The first two EOFs are east west dipoles centered at 60°N, and together they explain 57% of the variance. The magnitude of the buoyancy anomalies is on the order of $10^{-3} m/s^2$. Buoyancy anomalies are associated with temperature anomalies with a maximum value of 0.8°C and compensating salinity anomalies with a maximum value of 0.065 psu, but the buoyancy anomalies are dominated by temperature.

The first two principal component (PC) timeseries (shown in the bottom panel of Figure 3-3) are one quarter cycle out of phase, indicating a propagating mode. The buoyancy anomalies propagate westwards with a velocity of 0.37 to 0.47 cm/s, taking approximately 34 years to cross the basin, as demonstrated by the a Hovmoller plot of subsurface buoyancy anomalies at 60°N (see left panel of Figure 3-5).

Figure 3-4 shows the first two EOFs of buoyancy at a depth of 265 m in the NH of the small basin in Bowl. The buoyancy anomalies are maximum on the western boundary of the subpolar gyre, and the first two EOFs explains 41% of the variance. The magnitude of the buoyancy anomalies is on the order of $10^{-3} m/s^2$. Buoyancy anomalies are associated with temperature anomalies with a maximum value of 0.5°C and salinity anomalies with a maximum value of 0.036 psu, but the buoyancy anomalies are dominated by temperature. A Hovmoller plot of subsurface buoyancy anomalies at 60°N (see right panel of Figure 3-5) demonstrates that although there is westward propagation of buoyancy anomalies, the largest buoyancy anomalies are confined to the region near the western boundary. It is somewhat difficult to estimate the propagation speed of buoyancy anomalies in Bowl from the Hovmoeller plot, and more sophisticated techniques, such as Radon transforms, would likely be useful.
However, I estimate that westward velocities range from 0.60 to 0.68 cm/s.

While both the spatial structure and timescale of the buoyancy anomalies are quite different in Flat and Bowl, both models exhibit significant buoyancy variability on the western boundary of the subpolar gyre. The mean buoyancy in a box along the western boundary between 40°N and 65°N latitude (box shown in black in the top panels of Figure 3-3 and 3-4) from 130 to 320 m depth is computed, henceforth called the western boundary buoyancy (WBB) timeseries. The normalized WBB timeseries (henceforth WBB index) for Flat and Bowl are plotted in black in the bottom panels of Figures 3-3 and 3-4, respectively. The WBB index is highly correlated with the first PC of subsurface buoyancy (correlation of 0.9 at lag=3 years for Flat and correlation...
Figure 3-4: (Top) The first 2 EOFs of buoyancy at a depth of 265 m in the NH small basin in Bowl. The buoyancy anomalies are maximum on the western boundary of the subpolar gyre, and the first two EOFs explains 41% of the variance. (Bottom) The first 2 PC timeseries (blue and green) and the western boundary buoyancy (WBB) index (black). The WBB index is a timeseries of the buoyancy anomalies in a box along the western boundary (box shown in black in top panels) normalized to have a mean of zero and a standard deviation of one.
Figure 3-5: Hovmoller plot of subsurface (depth of 265 m) buoyancy anomalies at 60°N for Flat (left panel) and Bowl (right panel).
of 0.9 at lag=0 for Bowl).

The powerspectra of the WBB index are shown in Figure 3-6 (black lines). For Flat the powerspectrum of the WBB index is red at high frequency, flattens out at low frequency, and has a peak at a timescale of 34 years, like the powerspectrum of the MOC index. The peak in the powerspectra of the MOC and WBB indicies matches the time it takes for the buoyancy anomalies to propagate across the basin (34 years). For Bowl the powerspectrum of the WBB index is red at high frequencies and flattens out at low frequencies, like that of the MOC index. In both Flat and Bowl the transition from a red spectrum to a flat spectrum (the “kink” in the spectrum) occurs at a timescale of approximately 24 years, although it is somewhat hard to estimate this timescale in Flat due to the presence of the large peak. The relationship between MOC variability and buoyancy anomalies on the western boundary of the subpolar gyre will be elucidated in section 3.3.

Despite the similarity of the powerspectra of the MOC index and the WBB index, it is worth noting that in both models the MOC index has more power at high frequencies than the WBB index. MOC variability on short timescales is primarily due to overturning cells forced by Ekman transport variability, which responds rapidly to changes in windstress (see theory developed in Jayne and Marotzke (2001)). In fact, if EOFs of the MOC are computed using monthly data (with the seasonal cycle removed), the first EOF of the MOC is a dipole of overturning anomalies forced by the wobbling of the northern hemisphere jet (not shown).

3.2.1 Westward Propagation of Anomalies

In this section I will show that the propagating buoyancy anomalies in Flat can be understood as baroclinic Rossby waves. Additionally, I will show that the timescale of the oscillation in Flat is set by the time it takes for first baroclinic Rossby waves to cross the basin. Despite the weaker signal of propagation in Bowl, I will show that the timescale at which the spectrum of the WBB index flattens out in Bowl is given by the time it takes a first baroclinic Rossby wave to cross the basin, as was suggested by Frankignoul et al. (1997). First, I will describe the direction of propagation of the
Figure 3-6: The power spectra for the MOC in the box 8°S to 50°N, 620 to 1890 m depth (blue) and the powerspectra of the average buoyancy in a box along the western boundary (black) for Flat (top) and Bowl (bottom).
waves heuristically. Next, I will solve the quasi-geostropic potential vorticity equation linearized about the mean state to estimate the magnitude of the (real part) of the phase velocity of the waves.

**Heuristic Description of Waves**

In order to understand the direction of propagation of the buoyancy anomalies, I consider the mean buoyancy gradient and velocity fields (mean quantities will be denoted by capital letters). The mean buoyancy gradient is primarily in the meridional direction with $\partial B/\partial y < 0$. By the thermal wind relation this implies that $\partial U/\partial z > 0$, a mean zonal current $U$ that increases with height, as is observed (mean zonally averaged zonal current is shown in the left panel of Figure 3-7). The mean meridional buoyancy gradient results in a mean meridional potential vorticity (PV) gradient in addition to the mean planetary vorticity gradient $\beta$. The mean meridional potential vorticity gradient is

$$Q_y = \beta - \left( \frac{f^2}{N^2} U_z \right)_z,$$

where $f$ is the Coriolis parameter and $\beta$ is its meridional gradient, $U$ is the mean zonal velocity, and $N^2$ is the mean buoyancy frequency. Subscripts denote partial derivatives. The mean meridional PV gradient zonally averaged over the small basin at 60°N is shown in the right panel of Figure 3-7. The meridional PV gradient due to the meridional buoyancy gradient is an order of magnitude larger than $\beta$.

Let us consider a buoyant perturbation: the buoyant perturbation is associated with high sea surface height and an anticyclonic circulation that is maximal near the surface. West of the anomaly’s center the northward velocity brings buoyant (low PV) water from the south, and east of its center, the southward velocity brings dense (high PV) water from the north, leading to westward motion of the anomaly. However, there is a mean eastward zonal velocity, and a competition arises between the mean advection and propagation. In both Flat and Bowl the westward propagation wins out, and buoyancy anomalies are observed to propagate westwards, as I will show theoretically in the next section.
Figure 3-7: (Left Panel) Time mean zonal mean zonal current at 60°N in the small basin (zonally averaged away from boundaries) for Flat (black line) and Bowl (blue line). (Right Panel) Time mean zonal mean meridional potential vorticity gradient at 60°N in the small basin for Flat (black line) and Bowl (blue line).
Linear Rossby Waves

In this section I will solve the quasi-geostrophic potential vorticity (QGPV) equation linearized about the mean state to estimate the magnitude of the (real part) of the phase velocity of the waves. I will utilize the approach described in Tulloch et al. (2009), and code for solving the vertically discretized equations that are shown here was provided to me by Ross Tulloch.

The local mean velocity \( \mathbf{U} = U(z) \mathbf{x} + V(z) \mathbf{y} \) and squared buoyancy frequency \( N^2(z) = -(g/\rho_0)\frac{d\rho}{dz} \) are assumed to be slowly varying functions of horizontal location. Denoting the slowly varying mean state with capital letters and perturbations with lower case letters, the QGPV equation can be written:

\[
\partial_t q + \mathbf{U} \cdot \nabla q + \mathbf{u} \cdot \nabla Q = 0, \quad -H < z < 0, \tag{3.2}
\]

where \( q = \nabla^2 \psi + (f^2/N^2 \psi_z)_z \) is the QGPV anomaly, \( \psi \) is the streamfunction anomaly, \( f \) is the local Coriolis parameter, and \( H(x, y) \) is the local depth of the ocean. The mean QGPV gradient \( \nabla Q \) is given by

\[
\nabla Q = \left[ V_{xx} - U_{yx} + \left( \frac{f^2}{N^2} V_z \right)_z \right] \mathbf{x} + \left[ \beta + V_{xy} - U_{yy} - \left( \frac{f^2}{N^2} U_z \right)_z \right] \mathbf{y}. \tag{3.3}
\]

At the upper and lower boundaries we assume \( w = 0^3 \), allowing the linearized buoyancy equation to be written as:

\[
\partial_t b + \mathbf{U} \cdot \nabla b + \mathbf{u} \cdot \nabla B = 0, \quad z = 0, -H. \tag{3.4}
\]

The buoyancy anomaly \( b = -g \rho/\rho_o \) is related to the streamfunction: \( b = f \psi_z \).

We seek wave solutions of the form

\[
\psi(x, y, z, t) = \Re \{ \hat{\psi}(z) \exp[i(kx + \ell y - \omega t)] \}, \tag{3.5}
\]

\(^3\)Slowly varying, small amplitude bathymetry can be incorporated into the QG model by including the vertical velocity at the bottom, \( w(-H) = -\mathbf{U} \cdot \nabla H \).
and likewise for \( q \) and \( b \). In general, both \( \omega \) and \( \hat{\psi} \) are complex. \( \hat{\psi} \) contains information about the amplitude of the wave and its vertical structure. \( \omega \) is the angular frequency, whose real part gives the phase velocity of the wave and whose imaginary part (if it exists) implies growth or decay of the wave. Substituting Equation (3.5) into Equations (3.2) and (3.4), one obtains the linear eigenvalue problem for frequency \( \omega \):

\[
(K \cdot U - \omega_n) \hat{q}_n = (\ell Q_x - k Q_y) \hat{\psi}_n, \quad -H < z < 0,
\]

\[
(K \cdot U - \omega_n) \hat{b}_n = (\ell B_x - k B_y) \hat{\psi}_n, \quad z = 0, -H,
\]

where \( K = (k, \ell) \) is the wavenumber, and \( \hat{\psi}_n \) is the \( n^{th} \) eigenvector. \( \hat{q}_n \) and \( \hat{b}_n \) can be written in terms of \( \hat{\psi}_n \):

\[
\hat{q}_n = \frac{\partial}{\partial z} \left( \frac{f^2}{N^2} \frac{\partial \hat{\psi}_n}{\partial z} \right) - K^2 \hat{\psi}_n, \quad (3.7a)
\]

\[
\hat{b}_n = f \frac{\partial \hat{\psi}_n}{\partial z}, \quad (3.7b)
\]

Rossby waves in a resting ocean

In a resting ocean, \( U = 0 \), implying \( B_x = B_y = Q_x = 0 \) and \( Q_y = \beta \). Equation (3.6) becomes:

\[
\omega \left[ \frac{\partial}{\partial z} \left( \frac{f^2}{N^2} \frac{\partial \hat{\psi}}{\partial z} \right) - K^2 \right] - \beta k \hat{\psi} = 0, \quad -H < z < 0 \quad (3.8a)
\]

\[
\frac{\partial \hat{\psi}}{\partial z} = 0, \quad z = 0, -H, \quad (3.8b)
\]

where \( K = |K| \). The above reduces to the standard Rossby wave dispersion relation

\[
\omega_m = \frac{-k \beta}{K^2 + K_m^2}, \quad (3.9)
\]
where $K_m$ is the $m^{th}$ deformation wavenumber, which is given by the following Sturm Liouville problem:

$$
\frac{d}{dz} \left( \frac{f^2}{N^2} \frac{d\Phi_m}{dz} \right) = -K_m^2 \Phi_m, \quad \left. \frac{d\Phi_m}{dz} \right|_{z=0} = \left. \frac{d\Phi_m}{dz} \right|_{z=-H} = 0. \quad (3.10)
$$

The eigenfunctions $\Phi_m$ are often called the “neutral modes”; they form an orthonormal basis of the vertical structure in a resting ocean. The eigenvalues $K_m$ are the deformation wavenumbers and their reciprocals are the radii of deformation $R_m \equiv 1/K_m$. Theoretically, there are an infinite number of eigenvalues, ordered from largest to smallest $K_1, K_2, ... K_\infty$.

In order to calculate the radii of deformation and neutral modes in the model, $N^2$ is calculated at each level, and (3.10) is discretized in the vertical. In the discrete case there are as many eigenvalues as there are layers. The top panels of Figure 3-8 show the first baroclinic deformation radius ($R_1$) zonally averaged over the small basin for Flat (left) and Bowl (right). Comparing the observed scale of the waves at 60°N in Flat, which is on the order of 3000 km, to the Rossby radius, which is only 11 km, it is apparent that we can make the longwave approximation ($K < < K_1$) when calculating the phasespeed of the waves. While smaller than that in Flat, the scale of the buoyancy anomalies in Bowl is still on the order of 1000 km, while the deformation radius is 12 km. Despite the weaker signal of westward propagation present in Bowl, we will see that the propagation of long Rossby waves is also relevant for understanding the buoyancy anomalies seen in Bowl.

In the longwave limit ($K < < K_1$), the first baroclinic mode has a westward phase speed:

$$
c_H = \omega_1/k = -\beta R_1^2 
$$

(3.11)

The bottom panels of Figure 3-8 show the predicted westward phase speed of long first baroclinic Rossby waves in a resting ocean zonally averaged over the small basin (blue lines). The predicted phase speed at 60°N is 0.15 cm/s for Flat and 0.16 cm/s for Bowl. Also included in Figure 3-8 is the range of phasespeeds seen in the models.
at 60°N (red bars). The observed phase speeds at 60°N range from 0.37 to 0.47 cm/s for Flat and 0.60 to 0.68 cm/s for Bowl, which is a factor of 2-4 times larger than the phases speeds predicted by linear Rossby waves in a resting ocean.

The reason for this discrepancy is that mean current shears change the structure of the waves by altering the mean PV gradient (in the previous section I showed that the mean meridional PV gradient at 60°N is primarily determined by the mean meridional buoyancy gradient rather than $\beta$) and by Doppler shifting the phase velocity. In the next section I will show that inclusion of the mean flow leads to a considerable improvement in the match between the observed and predicted phase velocities.

**Rossby Waves in the presence of observed currents and PV gradients**

In this section I will calculate the phase velocities and vertical structure of Rossby waves in the presence of mean velocities and PV gradients. I will show that the predicted phase speed of the waves better matches the phase speed observed in the model when mean velocities and PV gradients are included. At each spatial location the mean velocity and buoyancy fields are calculated from the model, and the mean PV gradients are computed. Equation (3.6) is used to compute the $\omega_n$ and $\psi_n$. We then chose the vertical shear mode $\psi_n$ whose real part has the largest projection on the first neutral mode $\phi_1$.

The bottom panels of Figure 3-8 show the westward phase speed of long first baroclinic Rossby waves zonally averaged over the small basin when the mean flow and PV gradient are included for $l = 0$ (black line) and $l = k$ (green line). In Flat the predicted phase speed at 60°N is 0.72 cm/s for $l = 0$ and 1.0 cm/s for $l = k$. Comparing to the observed phase speed of 0.37 to 0.47 cm/s, we see that the predicted phase speeds are closer to the observed phase speed when the mean flow and PV gradient are included, particularly when I make the assumption $l = 0$. In Bowl, the predicted westward phase speed at 60°N is 0.38 cm/s for $l = 0$ and 0.70 cm/s for $l = k$. Comparing to the observed phase speed of 0.60 to 0.68 cm/s, we see

---

4 These phase speeds were calculated for buoyancy anomalies averaged over the latitude range 55° - 65°N, hence the latitude range shown.
that the predicted phasespeeds are closer to the observed phase speed when the mean flow and PV gradient are included, although for Bowl the assumption that $k = l$ leads to phase speeds that match the observations better than does the assumption that $l = 0$.

### 3.2.2 Timescale of Variability

I have demonstrated that the propagating buoyancy anomalies in Flat can be understood as baroclinic Rossby waves. The time that it takes for these waves to cross the basin (34 years) sets the timescale of the oscillation. Interestingly, the Rossby waves are observed to speed up in the western part of the basin (see Figure Hovmoller) as a consequence of the deepening of the thermocline in the west, as has been seen in observations (Chelton and Schlax, 1996).

It should be emphasized, however, that the Rossby waves seen in Flat are quite unrealistic. Observed Rossby waves at 50°N have length scales on the order of 500 km (Chelton and Schlax, 1996), rather than several thousand kilometers as observed in Flat. Additionally, at high latitudes Rossby waves are not observed to propagate all the way across the basin. In fact LaCasce and Pedlosky (2004) and Isachsen et al. (2007) suggests that baroclinic Rossby waves are unstable and break up into deformation scale eddies, and thus only Rossby waves in the tropics are able to cross the basin before succumbing to instability. The origin of these (unrealistic) large-scale Rossby waves will be elucidated in section 4.2. However, the Flat bottomed model provides an excellent test case for understanding the response of the MOC to regular buoyancy anomalies which strike the western boundary, as we will see in Section 3.3.

The weaker, less coherent propagation of buoyancy anomalies seen in Bowl is more in accord with observations of baroclinic Rossby waves. The largest buoyancy anomalies are constrained to the western part of the basin. Interestingly, the amplitude of the Rossby wave field in observations is also found to be larger in the western part of the North Atlantic than the eastern part (Ozychny and Cornillon, 2004). Here, I will show that the timescale at which the spectrum of the WBB index flattens out in Bowl is given by the time it takes for a first baroclinic Rossby wave to cross the
Figure 3-8: (Top Panels) First baroclinic Rossby radius ($R_1$) zonally averaged over the small basin for Flat (left) and Bowl (right). (Bottom Panels) Predicted first baroclinic long Rossby wave phase speeds zonally averaged over the small basin for Flat (left) and Bowl (right). Three different estimates of the phase speed are included: the predicted phase speed for a resting ocean (blue lines) and the predicted phase speed when the mean flow and PV gradients are included for the case $l = 0$ (black lines) and $l = k$ (green lines). The red bars in the bottom panels show the observed phase speed of the waves in Flat and Bowl.
Frankignoul et al. (1997) studied the decadal response of the ocean to stochastic windstress forcing. The baroclinic variability is governed by the equation for long first baroclinic Rossby waves forced by wind stress curl fluctuations:

\[ \frac{\partial p}{\partial t} + c_R \frac{\partial p}{\partial x} = -\frac{\rho f^2}{H_{bc}} R_1^2 w_e, \]

where \( p \) is the baroclinic pressure field, \( R_1 \) is the deformation radius, and \( c = -\beta R_1^2 \) is the phase speed of long, nondispersive Rossby waves. \( H_{bc} \) is a baroclinic depth scale that depends on the stratification. \( f \) is the Coriolis parameter, \( \beta \) is the meridional gradient of the Coriolis parameter, and \( \rho \) is the density. \( w_e \) is the Ekman pumping which is given by the windstress \( \tau \):

\[ w_e = \mathbf{\hat{z}} \cdot \frac{1}{\rho} \left( \nabla \times \frac{\tau}{f} \right) \]

At each frequency, the baroclinic response consists of a forced response plus a free Rossby wave generated at the eastern boundary. If the wind stress forcing is white, the model predicts a spectrum that is red at high frequencies and levels off to a constant value at low frequencies. The timescale at which the spectrum levels off is given by the time it takes long baroclinic Rossby wave to cross the basin.

Using the propagation velocity of buoyancy anomalies at 60°N estimated from the Hovmoller plot (0.60 – 0.68 cm/s), I find that the timescale for a baroclinic Rossby wave to cross the basin is approximately 24 years. The timescale of 24 years is the timescale at which the powerspectrum of the WBB index transitions from a red spectra at high frequency to a flat spectrum at low frequency (see Figure 3-6). Thus, in Bowl the dominant frequency is also set by the timescale for first baroclinic Rossby waves to cross the basin.
3.2.3 Role of Air-Sea Buoyancy Fluxes

Numerous observational (Deser and Blackmon, 1993; Kushnir, 1994; Dong and Kelly, 2004; Dong et al., 2007) and modeling (Dong and Sutton, 2003; Grist et al., 2010; Shaffrey and Sutton, 2006) studies suggest that while interannual ocean surface buoyancy (or temperature) anomalies are forced by air-sea buoyancy fluxes, the ocean circulation plays a role in creating decadal buoyancy anomalies. These decadal buoyancy anomalies are then damped by air-sea buoyancy fluxes, leading to potentially important atmospheric temperature anomalies. I will now show that the decadal buoyancy anomalies seen in Flat and Bowl are damped by air-sea buoyancy fluxes, and thus the ocean circulation must play a role in creating these anomalies.

Figure 3-9 shows the spatial pattern of buoyancy anomalies and air-sea buoyancy flux anomalies through 60°N. The patterns were computed by projecting the buoyancy and buoyancy flux anomalies onto the WBB index at lag=0 and lag=-1 years, respectively. Air-sea buoyancy fluxes damp the buoyancy anomalies at all stages of the evolution of the buoyancy anomalies. An additional analysis of a 100 year timeseries of monthly data confirms that monthly buoyancy flux anomalies (seasonal cycle removed) damp the buoyancy anomalies. Air-sea buoyancy fluxes on the order of $4 \times 10^{-9} m^2/s^3$ are observed. The buoyancy fluxes are dominated by heat fluxes (as will be shown below), which have a maximum magnitude of $8 W/m^2$ in Flat and $6 W/m^2$ in Bowl.

Figure 3-10 shows the buoyancy budget in the box along the western boundary (box shown in black in Figures 3-3 and 3-4) for Flat (left) and Bowl (right). The top panels show the lagged covariance of the WBB index and the buoyancy in the box along the western boundary (WBB timeseries), as well as the contributions of temperature and salinity to the buoyancy anomalies. Temperature and salinity anomalies compensate each other, but temperature anomalies dominate the buoyancy anomalies in this near surface box. The bottom panels show the lagged covariance between the

$I$ show the buoyancy flux at lag=-1 years and the buoyancy anomaly at lag=0 in order to show that the buoyancy fluxes which occur before the maximum buoyancy anomaly are acting to damp the anomaly.
WBB index and the buoyancy tendency (black), buoyancy convergence (blue), and air-sea buoyancy flux due to temperature (red) and salinity (magenta) in the box along the western boundary. The red (blue) shading indicates the lags for which the buoyancy anomaly in the box is positive (negative). When the buoyancy anomaly is positive (negative), anomalous air-sea heat fluxes out of (into) the ocean are observed, damping the buoyancy anomaly. Freshwater fluxes play a negligible role in the buoyancy budget. Thus, the buoyancy tendency in the box is driven by convergence of buoyancy by the ocean circulation and damped by air-sea heat fluxes.

### 3.3 Relationship between buoyancy and MOC variability

As discussed in the previous two sections, both buoyancy in the subpolar gyre and the MOC exhibit variability on decadal timescales. In this section I will describe the relationship between buoyancy anomalies in the subpolar gyre and the MOC. Figure 3-11 shows the correlation between the MOC index and the WBB index at various lags for Flat (top panel) and Bowl (bottom panel). Although the spectrum of MOC index and WBB index are quite different in Flat and Bowl, in both models a minimum of the MOC occurs 6 years after the maximum buoyancy anomaly on the western boundary.

Figure 3-12 and 3-13 show anomalies of the MOC, buoyancy through 60°N, and buoyancy on the western and eastern boundaries projected onto the WBB index at the lags labeled in Figure 3-11 for Flat and Bowl, respectively. Lag 0 corresponds to the maximum buoyancy anomaly in the box on the western boundary (max WBB index) and positive (negative) lags are after (before) the maximum of the WBB index. In Flat, the buoyancy anomalies originate near the eastern boundary and propagate westward. In Bowl, the buoyancy anomalies appear to originate in the interior of the gyre quite near to the western boundary. In both models, when the buoyancy anomalies hit the western boundary (see lag=-2 yrs for Flat and lag=0 for Bowl), they
Figure 3-9: Spatial pattern of buoyancy anomalies and air-sea buoyancy flux anomalies through 60°N for (a) Flat and (b) Bowl. The spatial patterns were computed by projecting the buoyancy and buoyancy flux anomalies onto the WBB index at lag=0 and lag=−1 years, respectively. Positive (negative) air-sea buoyancy fluxes indicate a buoyancy flux into (out of) the ocean.
Figure 3-10: Buoyancy budget in the box along the western boundary for Flat (left) and Bowl (right). The top panels show the lagged correlation of the WBB index and the buoyancy in the box along the western boundary and the contributions of temperature (red) and salinity (magenta) to the buoyancy anomaly. The bottom panels show the lagged covariance between the WBB index and the buoyancy tendency (black), buoyancy convergence (blue), and air-sea buoyancy flux due to temperature (red) and salinity (magenta). The red (blue) shading indicates the lags for which the buoyancy anomaly in the box is positive (negative). Positive (negative) fluxes are into (out of) the ocean.
Correlation: MOC and WB buoyancy anomalies

Figure 3-11: Correlation between the WBB index and the MOC index at various lags for Flat (top) and Bowl (bottom). Lag=0 corresponds to the maximum buoyancy anomaly on the western boundary and positive (negative) lags are after (before) the maximum in the WBB index. Light blue lines show the correlation for all lags and the heavy lines indicate where the correlation is different from zero at the 95% confidence level. Open circles and labels A-G (top) and A-F (bottom) indicate the lags for which spatial fields are shown in Figure 3-12 and 3-13, respectively.
move southward at depth along the western boundary following the mean isopynals.

Hovmoller plots of the buoyancy anomaly on the western boundary on the $\sigma = 1 kg/m^3$ isopynal ($\sigma$ calculated from the $T$ and $S$ field relative to $T_o = 16^\circ$C and $S_o = 34.8$ psu) as a function of latitude and lag for Flat and Bowl are shown in the top panels of Figure 3-12 and 3-13, respectively. Lag=0 corresponds to the maximum buoyancy anomaly on the western boundary. The buoyancy anomalies travel southward along the mean isopynals at a velocity of approximately 2 cm/s, which is in accord with the speed of the mean DWBC (see Figure 2-5). Buoyancy anomalies take 10 years to travel from 60°N to the equator.

Via the zonally integrated thermal wind relation (Equation 1.1), buoyancy anomalies on the boundaries result in anomalies in the shear of the zonally integrated meridional velocity. Buoyancy anomalies on the eastern boundary are observed to be negligible, except in the subpolar gyre, implying that negative (positive) buoyancy anomalies on the western boundary lead to a spin up (down) of the MOC. The maximum negative MOC anomaly occurs six years after the maximum buoyancy anomaly on the western boundary of the subpolar gyre. As time progresses, the subsurface buoyancy anomalies on the western boundary decay from the north, resulting in a weakening of the MOC anomaly (see lag=10-15 for Flat and 8-13 for Bowl).

Hovmoller plots of MOC anomalies on the $\sigma = 1 kg/m^3$ isopynal as a function of latitude and lag are shown for Flat and Bowl in the bottom panels of Figure 3-12 and 3-13, respectively. South of 40°N the buoyancy anomalies on the eastern boundary are observed to be negligible. Hence MOC anomalies are the thermal wind response to buoyancy anomalies on the western boundary. In the subpolar gyre there are large buoyancy anomalies on the eastern boundary as well as on the western boundary, and the response of the MOC is more complicated.

It is worth stressing that the mode of buoyancy variability seen in Flat is highly unrealistic. However, it is included here since it provides an excellent test case for understanding the response of the MOC to regular buoyancy anomalies which strike the western boundary. In the ocean-only framework a number of models have studied the response of the MOC to periodic variability on the western boundary (Kawase,
Figure 3-12: Flat: MOC anomalies (top left), east-west section of buoyancy anomalies at 60°N (top right) and buoyancy anomalies along the western (bottom left) and eastern (bottom right) boundaries of the small basin projected onto the WBB index at (a) lag=-2 yrs (b) lag=0, (c) lag=2 yrs, (d) lag=6 yrs, (e) lag=10 years, (e) lag=12 yrs, and (f) lag=15 yrs. Only covariances which are significant at the 95% confidence level are plotted. Superimposed on the buoyancy anomalies on the western and eastern boundaries are the mean isopycnals (thin black lines).

1987; Johnson and Marshall, 2002b,a; Deshayes and Frankignoul, 2005). In Flat, the coupled model internally generates almost periodic buoyancy variability, allowing us to examine the response of the MOC to buoyancy anomalies striking the western boundary. Due to the coarse resolution and highly idealized geometry, the buoyancy variability in Bowl may also be unrealistic. The important point, however, is that in both models, which have drastically different buoyancy variability, the MOC responds to buoyancy anomalies on the western boundary of the subpolar gyre in the same way.
Figure 3-12: (cont.)
Figure 3-12: (cont.)

(d) Flat: lag=6 yrs

(e) Flat: lag=10 yrs

Figure 3-12: (cont.)
Figure 3.12 (cont.)

Plate: 18-12 yrs

Figure 3.12 (cont.)

Plate: 18-12 yrs
Figure 3-13: Bowl: MOC anomalies (top left), east-west section of buoyancy anomalies at 60°N (top right) and buoyancy anomalies along the western (bottom left) and eastern (bottom right) boundaries of the small basin projected onto the WBD index at (a) lag=0 (b) lag=2 yrs, (c) lag=6 yrs, (d) lag=8 yrs, (e) lag=10 years, and (e) lag=13 yrs. Only covariances which are significant at the 95% confidence level are plotted. Superimposed on the buoyancy anomalies on the western and eastern boundaries are the mean isopycnals (thin black lines).
(b) Bowl: lag=2 yrs

Figure 3-13: (cont.)

(c) Bowl: lag=6 yrs

Figure 3-13: (cont.)
(d) Bowl: lag=8 yrs

Figure 3-13: (cont.)

(e) Bowl: lag=10 yrs

Figure 3-13: (cont.)

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3.4 Parallels to other model studies: Rossby waves, the MOC, and the thermal wind relation

The main result of the previous sections is that buoyancy variability on the western boundary is due to baroclinic Rossby waves impinging on the boundary. These buoyancy anomalies then travel down the western boundary in the DWBC, leading to MOC variability according to the thermal wind relation. Here I will highlight other studies that have related MOC variability to baroclinic Rossby waves and buoyancy anomalies along the boundaries.

Huck et al. (1999); Colin de Verdiere and Huck (1999) and Arzel et al. (2007) studied MOC variability in idealized ocean models forced by fixed fluxes and idealized coupled models, respectively. They found two types of oscillatory behavior: (1) temperature anomalies propagating geostrophically westward in the eastward jet in the northern part of the basin and (2) larger stationary temperature anomalies in the
Figure 3-14: Flat: Hovmoller plot of (Top) buoyancy anomalies on the western boundary and (Bottom) MOC anomalies at the depth of the mean $\sigma = 1km/m^3$ isopycnal as a function of latitude and lag.
Figure 3-15: Bowl: Hovmoller plot of (Top) buoyancy anomalies on the western boundary and (Bottom) MOC anomalies at the depth of the mean $\sigma = 1 km/m^3$ isopynal as a function of latitude and lag.
north-west quadrant of the basin that respond to changes in the western boundary current transport. MOC variability was shown to be related to these decadal temperature anomalies. These anomalies bear an uncanny resemblance to the buoyancy anomalies observed in Bowl.

Te Raa and Dijkstra (2002) described an oscillation involving westward propagating Rossby waves and variability of the MOC in an idealized single hemisphere ocean model. They argued that the crucial mechanism of the oscillation is the phase difference between the zonal and meridional surface flow perturbations and the westward propagation of temperature anomalies. The timescale of the oscillation is set by the timescale it takes for the Rossby waves to cross the basin. In te Raa et al. (2004) and Frankcombe et al. (2008); Frankcombe and Dijkstra (2009) they argue that the basic features of this mode of variability are seen in the coupled GCMs. However, the westward propagation of temperature anomalies in these models are only observed when the data is averaged over all the latitudes in the basin, which is quite different from my model where the largest buoyancy anomalies are restricted to the subpolar gyre. Additionally, since the phase speed of baroclinic Rossby waves is a strong function of latitude, averaging over all latitudes in the North Atlantic seems problematic.

While the thermal wind relation has primarily been utilized to diagnose AMOC variability from the RAPID array data, a handful of studies have used the thermal wind relation to understand decadal MOC anomalies. Zanna et al. (2011) showed that MOC anomalies in an idealized ocean model can be understood by examining the evolution of deep density anomalies on the boundaries according to the thermal wind relation. Similar to my studies, Zanna finds that these density anomalies originate in the subpolar gyre, but she focuses on deep anomalies that lead to optimal excitation of MOC variability.

### 3.5 Role of Convection

Both observational (Joyce and Zhang, 2010; Pena-Molino, 2010) and modeling (Dong and Sutton, 2005; Danabasoglu, 2008; Zhang, 2008; Msadek et al., 2010) studies
have linked AMOC variability to variability in high latitude convection and water mass formation. Lead/lag relations are often used to hypothesize that convective variability is driving AMOC variability. Whether this is indeed the case or both convection and the AMOC are responding to some other process, such as the North Atlantic Oscillation, is unclear. In this section I will examine potential vorticity and convective index anomalies that are associated with the decadal mode of MOC and buoyancy variability. I will argue that decadal changes in convection are a response to preconditioning of water masses due to decadal buoyancy anomalies.

Here PV is approximated as the vertical component of the Ertel potential vorticity, and is calculated from the potential density field $\sigma$ (or buoyancy field $b$) as:

$$PV = -f \frac{\partial \sigma}{\partial z} = f \frac{\rho_0}{g} \frac{\partial b}{\partial z},$$

where $f$ is the Coriolis parameter.

The top panels of Figure 3-16 and 3-17 show anomalies of buoyancy (left panel, contours), PV (left panel, colors), and convective index (right panels) through 60°N projected onto the WBB index at various lags for Flat and Bowl, respectively. The bottom panels show PV anomalies on the western and eastern boundaries projected onto the WBB index at various lags. Lag 0 corresponds to the maximum buoyancy anomaly in the box on the western boundary (max WBB index) and positive (negative) lags are after (before) the maximum of the WBB index. Negative PV anomalies occur when the stratification is decreased ($\frac{\partial b}{\partial z} < 0$) and positive PV anomalies occur when the stratification is increased ($\frac{\partial b}{\partial z} > 0$).

Shallow negative (positive) PV anomalies at the surface accompany buoyant (dense) anomalies. These shallow PV anomalies are forced by air-sea heat fluxes as follows: air-sea heat fluxes out of the ocean damp positive buoyancy anomalies, leading to a PV flux out of the ocean and the negative PV anomaly at the surface. Similarly, air-sea heat fluxes into the ocean damp negative buoyancy anomalies, leading to a PV flux into the ocean and a positive PV anomaly at the surface. However, these PV anomalies are very shallow and do not affect deep convection.
Since the buoyancy anomalies decay with depth, positive (negative) subsurface PV anomalies accompany buoyant (dense) anomalies. Additionally, since the buoyancy anomalies tilt westward with height, negative (positive) PV anomalies are found on the eastward flank of positive (negative) buoyancy anomalies. These deeper PV anomalies are associated with anomalies in convection. Regions with low PV are preconditioned for convection, and positive anomalies of the convective index are observed. Regions with high PV are more stratified and thus less convection occurs.

In Flat the largest anomalies in the convective index occur on the eastern boundary. Positive (negative) buoyancy anomalies that increase with depth are associated with negative (positive) PV anomalies, and a positive (negative) anomaly in convection extending from the near the surface down to around 1 km depth.

Smaller, subsurface anomalies in convection occur on the western boundary in both Flat and Bowl. Positive (negative) buoyancy anomalies decreasing with depth on the western boundary are associated with positive (negative) PV anomalies and decreased (increased) convection. These PV anomalies travel southward along the western boundary. Negative (positive) PV anomalies are associated with an increase in strength of the MOC, as can be seen by comparing corresponding lags in Figures 3-16 and 3-17 to Figure 3-12 and 3-13. This is in accord with observational results at Line W which found that negative PV anomalies are associated with a strengthened deep western boundary current (Pena-Molino, 2010).

### 3.6 Variability of Meridional Ocean Heat Transport

Figure 3-18 shows the yearly mean meridional ocean heat transport (OHT) anomalies as a function of latitude and time for a 100 year segment of the (a) Flat and (b) Bowl runs. OHT anomalies on the order of 0.2 PW, or around 20% of the mean OHT, are observed. OHT anomalies are maximal between 20°S and 50°N and the anomalies exhibit significant coherence with latitude (spanning from 20° latitude to over 70°)
Figure 3-16: Flat: (Top panels) East-west sections of buoyancy anomalies (top left, contours), potential vorticity (PV) anomalies (top left, colors), and convective index anomalies (top right) at 60°N projected onto the WBB index at (a) lag=0 (b) lag=6 yrs. (Bottom Panels) PV anomalies along the western (bottom left) and eastern (bottom right) boundaries of the small basin projected onto the WBB index at (a) lag=0 (b) lag=6 yrs. Superimposed on the PV anomalies on the western and eastern boundaries are the mean isopycnals (thin black lines). Only covariances which are significant at the 95% confidence level are plotted.
Figure 3-16: (cont.)

(b) Flat: lag=6 yrs

lag=6 yrs  
b' and PV' along y=60° N

Conv Ind' along y=60° N

kg m^-4 s^-1

kg m^-4 s^-1
Figure 3-17: Bowl: (Top panels) East-west sections of buoyancy anomalies (top left, contours), potential vorticity (PV) anomalies (top left, colors), and convective index anomalies (top right) at 60°N projected onto the WBB index at (a) lag=0 (b) lag=6 yrs. (Bottom Panels) PV anomalies along the western (bottom left) and eastern (bottom right) boundaries of the small basin projected onto the WBB index at (a) lag=0 (b) lag=6 yrs. Superimposed on the PV anomalies on the western and eastern boundaries are the mean isopynals (thin black lines). Only covariances which are significant at the 95% confidence level are plotted.
Figure 3-17: (cont.)

(b) Flat: lag=6 yrs
Decomposing the temperature and meridional velocity into their zonal means, denoted by $[V]$ and $[T]$, and the deviations from their zonal means, denoted by $V^*$ and $T^*$, OHT anomalies can be separated into the contribution by the zonally averaged circulation $\rho_oC_p[V][T]$ (middle panels) and the deviations from the zonal average circulation $\rho_oC_p[V^*T^*]$ (bottom panels). Like the mean OHT, south of $40^\circ$N OHT anomalies are dominated the heat transport by the zonal mean circulation. In the subpolar gyre, the gyre circulation plays a role in ocean heat transport, both in the mean and the variability.

Figure 3-19 shows the OHT anomalies associated with MOC variability as a function of latitude and lag for (a) Flat and (b) Bowl. OHT anomalies associated with MOC variability are computed by projecting OHT anomalies onto the MOC index at various lags, with negative (positive) lags corresponding to years before (after) the maximum MOC index. OHT anomalies on the order of 0.04 PW are associated with MOC anomalies with standard deviation of 1 Sv. As expected, positive OHT anomalies between $20^\circ$S and $50^\circ$N are associated with positive MOC anomalies. However, the anomalies of OHT that are associated with decadal variability of the AMOC are approximately 5 times smaller than the raw, annual average OHT anomalies. While variations in the strength of the MOC do lead to meridional OHT anomalies, these OHT anomalies are quite modest. Of course, these rather modest OHT anomalies may lead to significant decadal buoyancy anomalies as they are sustained for timescales on the order of 10 years.

In the bottom two panels of Figure 3-19 the OHT variability associated with MOC variability is decomposed into the contribution by the zonally averaged circulation $\rho_oC_p[V][T]$ (middle panels) and the deviations from the zonal average circulation $\rho_oC_p[V^*T^*]$ (bottom panels). In both Flat and Bowl, south of $40^\circ$N OHT anomalies are dominated the the zonal mean circulation, i.e. the MOC. In the subpolar gyre, the gyre circulation plays a significant role in OHT anomalies.
Figure 3-18: OHT anomalies in the small basin for (a) Flat and (b) Bowl. Total advective OHT anomalies (top panels) are split into OHT anomalies by the zonal mean circulation $\rho_o C_p |V||T|$ (middle panels) and by deviations from the zonal mean circulation $\rho_o C_p [V^*T^*]$ (bottom panels).
Figure 3-18: (cont.)
Figure 3-19: OHT anomalies associated with changes in the strength of the MOC for (a) Flat and (b) Bowl. Total advective OHT anomalies $\rho C_p[V T]$ (top panels) are split into OHT anomalies by the zonal mean circulation $\rho C_p[V][T]$ (middle panels) and by deviations from the zonal mean circulation $\rho C_p[V' T']$ (bottom panels)
(b) Bowl

Figure 3-19: (cont.)
Chapter 4

Mechanisms of Decadal Variability

In the last section we saw that decadal anomalies of the MOC are associated with buoyancy anomalies on the western boundary of the subpolar gyre. These buoyancy anomalies travel southward along the western boundary, leading to large-scale MOC variability via the thermal wind relation. In this section, I will address the origin of these decadal buoyancy anomalies. Specifically: (1) Does large-scale MOC variability play an active role in creating buoyancy anomalies in the subpolar gyre? (2) Is the mode of buoyancy and MOC variability a coupled mode, a damped ocean-only mode, or a self-sustained ocean-only mode? A coupled mode is defined as a mode of variability for which two way coupling between the atmosphere and ocean is essential for its existence, such as ENSO. An ocean-only mode is a mode that does not require feedback of SST on the atmospheric circulation for its existence. If the mode is damped, it requires stochastic forcing from the atmosphere to be excited; while if the mode is self-sustained it exists in the absence of stochastic atmospheric forcing.

4.1 Ocean-Only Experiments

In order to understand the origin of the variability in the coupled model, I have conducted a series of experiments using ocean-only versions of the Flat and Bowl models. In each case the ocean model is initialized with a state from the spun-up coupled model and forced with 5-day mean timeseries of heat, freshwater, and
momentum fluxes. The 5-day mean forcing timeseries from the coupled model are linearly interpolated in order to supply the ocean with hourly forcing (the model time-step). Each experiment is run for 100 years.

4.1.1 Ability of ocean-only model to reproduce timeseries from the coupled model

Before using ocean-only models to probe the origin of the buoyancy and MOC variability, I must ensure that I am able to accurately reproduce the coupled runs using an ocean-only model. Thus, I conducted an ocean-only experiment, which I will call TOTAL, in which the ocean was forced with a 5 day timeseries of the heat, freshwater, and momentum fluxes as well as a restoring of SST and SSS to that of the coupled run on a timescale of 71 days\(^1\) for SST and 1 year for SSS. Restoring of SST and SSS are needed for the ocean-only model to accurately reproduce the coupled model since a number of processes, such as rapid convective events, are lost when the ocean model is forced with 5-day averaged forcing. Figure 4-1 shows a comparison of the MOC index from the coupled model (blue curve) and the experiment TOTAL (green curve) for Flat (top panel) and Bowl (bottom panel). The MOC timeseries in TOTAL tracks that of the coupled run almost perfectly.

4.1.2 Role of MOC in creating buoyancy anomalies

In section 3.3 I argued that the MOC variability observed in the model is the thermal wind response to buoyancy anomalies on the western boundary, which originate in the subpolar gyre. However, I did not explain the origin of these buoyancy anomalies, which are damped by air-sea buoyancy fluxes. In this section I will describe an ocean-only experiment that I conducted in order to determine if large-scale MOC variability

\(^1\)71 days is the timescale calculated for my model from the canonical value for damping of SST anomalies \(\alpha = 20 W/m^2\) (Frankignoul et al., 1998). The damping timescale can be calculated from \(\alpha = \rho_0 C_p \delta x / \alpha\), where \(\rho_0\) is a reference density, \(C_p\) is the heat capacity of water, and \(\delta x\) is the thickness of the surface layer, here the thickness of the top model layer (30 m). However, here I am not damping SST anomalies, but rather restoring SSTS to the coupled model in order to nudge the ocean model towards the state of the coupled model, and I could have chosen any reasonable value for the restoring timescale.
Figure 4-1: Yearly MOC timeseries in the coupled model (black curve) and ocean-only model experiment TOTAL (green curve) for Flat (top panel) and Bowl (bottom panel). The MOC timeseries in TOTAL matches the MOC timeseries in the coupled model almost perfectly.
plays a role in creating the buoyancy anomalies observed in the subpolar gyre.

The basic design of the experiment is kill the large-scale MOC variability and see if the buoyancy anomalies in the subpolar gyre are modified. In section 3.1 I showed that south of 50°N the meridional velocity and its variability are concentrated along the western boundary. As a result, I chose to restore temperature and salinity along the western boundary south of 50°N to climatology with a restoring timescale of two months. This has the effect of killing the large-scale MOC variability since it is concentrated on the western boundary, while not restoring temperature and salinity in the subpolar gyre where the buoyancy anomalies originate. I will refer to this experiment as RESTORE-WB.

The top panels of Figures 4-2 show the MOC timeseries in the coupled run and the RESTORE-WB experiment. Restoring temperature and salinity along the western boundary greatly reduces the amplitude of the MOC variability in both Flat (left) and Bowl (right). The residual MOC variability observed in the RESTORE-WB experiment is primarily due to variability in Ekman transport forced by wind variability. The bottom panels of Figures 4-2 show the buoyancy anomalies in a box near the western boundary for Flat and Bowl. The box used is the same as the box shown in Figures 3-3 and 3-4 (40°N and 65°N, 130 to 320 m depth), but the points immediately adjacent to the western boundary have been removed since temperature and salinity are restored to climatology along the western boundary. Although the MOC variability has been damped substantially, the buoyancy anomalies near the western boundary of the subpolar gyre remain virtually unchanged. From this experiment I conclude that although large-scale MOC variability does lead to variability of ocean heat and freshwater transports (as seen in Section 3.6), these transports are not responsible for creating the buoyancy anomalies that drive the MOC variability. In both Flat and Bowl the large-scale MOC responds passively to buoyancy anomalies that originate in the subpolar gyre. Thus, if I can explain the origin of these buoyancy

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2 Of course variability of the velocity field (and hence the MOC) and buoyancy field in the subpolar gyre are tightly coupled according to the thermal wind relation. The point here is that variability of the MOC in the subtropical gyre and the resulting OHT anomalies do not play a role in creating the buoyancy anomalies seen in the subpolar gyre. These anomalies are formed by processes local to the subpolar gyre.
anomalies, I will be able to successfully explain the mode of MOC variability.

4.1.3 Role of atmospheric forcing

In order to determine if stochastic atmospheric forcing is needed to excite the mode of buoyancy and MOC variability, I conducted an ocean-only experiment in which the ocean is forced with climatological forcing. In the experiment, which I will refer to as CLIM-DAMP, the ocean model is forced with 5-day climatological (100 year average from coupled model) forcing of heat, momentum, and freshwater and damping of SST to climatology with the canonical value of 20W m\(^{-2}\) K\(^{-1}\) (Frankignoul et al., 1998).

The MOC timeseries for the experiment CLIM-DAMP is compared to the MOC timeseries from the coupled run in Figure 4-3. For Flat (top panel) the CLIM-DAMP experiment reproduces the low-frequency MOC variability of the coupled model amazingly well. Thus, the decadal mode of variability observed in Flat is a self-sustained ocean-only mode damped by air-sea heat fluxes.

For Bowl (bottom panel) the MOC variability rapidly decays for the CLIM-DAMP experiment. In the presence of realistic damping by air-sea heat fluxes, MOC variability does not exist without continuous excitation by stochastic atmospheric forcing. To further explore the mode of variability, I conducted another experiment, which I will call CLIM-WEAK-DAMP. This experiment is the same as CLIM-DAMP, but the damping of SST anomalies was set to be only 4W m\(^{-2}\) K\(^{-1}\), a value much weaker than typical estimates of damping of SST anomalies. When damping of SST anomalies is weak, MOC variability persists even in the absence of stochastic atmospheric forcing. The MOC and buoyancy variability observed in the CLIM-WEAK-DAMP experiment is much more regular than that observed in the coupled model, but the spatial pattern of buoyancy and MOC variability is almost identical to that of the coupled model. Thus, I conclude that the mode of buoyancy and MOC variability in Bowl is a damped ocean-only mode that is excited by stochastic atmospheric forcing.
Figure 4-2: (Top Panels) Yearly MOC timeseries in the coupled model (black curve) and ocean-only model experiment REST-WB (cyan curve) for Flat (left panel) and Bowl (right panel). (Bottom Panels) Yearly WBB timeseries in the coupled model (blue curve) and ocean-only model experiment REST-WB (green curve) for Flat (left panel) and Bowl (right panel).
Figure 4-3: Yearly MOC timeseries in the coupled model (black curve) and ocean-only model experiment CLIM-DAMP (blue curve) for Flat (top panel) and Bowl (bottom panel). For Bowl (bottom panel) an additional experiment, CLIM-WEAK-DAMP is shown. CLIM-WEAK-DAMP is the same as CLIM-DAMP, but a damping of $4Wm^{-2}K^{-1}$ is used rather than the canonical value of $20Wm^{-2}K^{-1}$. 
4.1.4 Role of heat fluxes and winds

In Bowl stochastic atmospheric forcing is needed to excite the mode of buoyancy and MOC variability in the presence of realistic damping by air-sea heat fluxes. I will now address the relative roles of wind and buoyancy forcing in this mode of variability by conducting two additional ocean-only experiments. In an experiment, which I will call WINDS, the ocean model is forced with winds from the coupled model and climatological heat and freshwater fluxes and damping of SST to climatology with the canonical value of $20 W m^{-2} K^{-1}$. In the absence of variable wind and heat flux forcing, the MOC variability dies (as was seen in CLIM-DAMP experiment), but stochastic wind forcing excites the mode of variability, as can be seen in Figure 4-4 (blue curve).

In an experiment which I will call HEAT the model is forced with heat fluxes from the coupled model as well as restoring of SST to that of the coupled model on a timescale 71 days, but the winds and the freshwater fluxes are climatological. Significant MOC variability is present in the HEAT experiment as well (see red curve in Figure 4-4), and the sum of the MOC variability in WINDS and HEAT matches that of the coupled model almost perfectly.

Since the HEAT experiment includes a relaxation of SST to that of the coupled model, forcing of SST by wind variability is implicitly included in this experiment. Therefore, these experiments cannot be used to cleanly separate the effects of wind-stress and heat flux forcing on MOC variability. By damping SST to climatology, WINDS may underestimate SST anomalies due to momentum forcing. Similarly, including restoring of SST to that of the coupled model in HEAT means that the HEAT experiment implicitly includes the effects of momentum fluxes in creating SST anomalies. Despite the inability of these experiments to separate the roles of heat and momentum fluxes, I believe that is is most likely that the mode of MOC variability observed in Bowl is excited by the winds since in all experiments decadal air-sea heat fluxes are observed to damp buoyancy anomalies. However, it is certainly possible that higher-frequency air-sea heat flux anomalies provide stochastic forcing that can
Figure 4-4: Yearly MOC timeseries in the coupled model (black curve) and ocean-only model experiments WINDS (blue curve) and HEAT (red curve) for Bowl. In WINDS the ocean model is forced with winds from the coupled model and climatological heat and freshwater fluxes and damping of SST to climatology with the canonical value of 20Wm\(^{-2}\)K\(^{-1}\). In HEAT the ocean is forced with heat fluxes from the coupled model as well as restoring of SST to that of the coupled model on a timescale of one year, but the winds and the freshwater fluxes are climatological.

excite the mode of MOC variability.

4.1.5 Summary of Ocean-Only Experiments

Decadal MOC anomalies are due to buoyancy anomalies on the western boundary of the subpolar gyre. Using ocean-only experiments, I demonstrated that the buoyancy anomalies are formed by processes local to the subpolar gyre, and MOC anomalies in the subtropical gyre do not significantly alter these anomalies. In Flat the buoyancy anomalies exist without stochastic atmospheric forcing, even in the presence of realistic damping by air-sea heat fluxes. Thus, in Flat, the buoyancy and MOC variability is a self-sustained ocean-only mode. In Bowl the mode of buoyancy and MOC variability exists in the ocean without stochastic atmospheric forcing, but realistic damping destroys the mode and stochastic atmospheric forcing is needed to continuously excite the mode. Thus, the mode is believed to be a damped ocean-only
4.2 Energy Sources for Oscillations

Ocean-only experiments demonstrate that the decadal mode of buoyancy and MOC variability observed in Flat is a self-sustained ocean-only mode. No stochastic atmospheric forcing is needed to excite the mode, so there must be a source of energy internal to the ocean to create buoyancy anomalies. Although in Bowl stochastic atmospheric forcing is needed to excite the mode of variability in the presence of realistic damping, the mode of variability exists without stochastic atmospheric forcing when damping by air-sea heat fluxes is weak. Thus, in Bowl there also must be a source of energy internal to the ocean, but in Bowl this source of energy is not sufficient to grow against damping. In this section I will discuss how baroclinic instability could be a source of energy for the decadal oscillations.

4.2.1 Eddy creation of buoyancy variance

Following Colin de Verdiere and Huck (1999) the time mean (or mean over one period) linearized buoyancy variance equation can be written as:

\[
\frac{1}{2} \frac{\partial \overline{\nu^2}}{\partial t} = -\frac{1}{2} \overline{\mathbf{u} \cdot \nabla \nu^2} - \overline{\mathbf{u}' b'} \cdot \nabla b + \overline{D' b'} + \overline{F'_b},
\]

(4.1)

where \( \overline{u} \) and \( \overline{b} \) are the mean velocity and buoyancy, \( u' \) and \( b' \) are the velocity and buoyancy anomalies, \( F'_b \) is the buoyancy change forced by air-sea buoyancy flux anomalies, and \( D' \) is the buoyancy change forced by anomalies in sub-gridscale mixing processes and convection. The first term on the right hand side is the transport of variance by the mean flow and cannot create or destroy variance. The last two terms on the right hand side are the destruction of variance by diffusive processes/convection and air-sea heat fluxes (which we have shown always damp the anomalies), respectively. Therefore, in order for the mode to grow against mixing and damping by air-sea buoyancy fluxes, the term \(-\overline{u' b'} \cdot \nabla b\) must be positive averaged over the domain.
\langle -\mathbf{u}'\mathbf{b}' \cdot \nabla \bar{b} \rangle > 0$, where \langle \rangle indicates the domain average. This term is the creation of buoyancy variance by eddies.

The eddy creation of buoyancy variance $-\mathbf{u}'\mathbf{b}' \cdot \nabla \bar{b}$ and its zonal component $-\mathbf{u}'\mathbf{b}' \cdot \bar{b}_x$, meridional component $-\mathbf{v}'\mathbf{b}' \cdot \bar{b}_y$, and vertical component $-\mathbf{w}'\mathbf{b}' \cdot \bar{b}_z$ are plotted in Figure 4-5. In both (a) Flat and (b) Bowl the meridional component dominates over the zonal and vertical components. The dominance of the meridional component is not surprising since mean buoyancy gradients are much larger in the meridional direction than the zonal direction.

In Flat $-\mathbf{u}'\mathbf{b}' \cdot \nabla \bar{b} > 0$ in a broad region near the eastern boundary and also along the western boundary of the subpolar gyre. In Bowl $-\mathbf{u}'\mathbf{b}' \cdot \nabla \bar{b} > 0$ only along the western boundary of the subpolar gyre. In both models is obvious that $\langle -\mathbf{u}'\mathbf{T}'\bar{b} \cdot \nabla \bar{b} \rangle > 0$, indicating that the eddies are a source of buoyancy variance. Unfortunately, it is more difficult to conclude where in the domain the eddies are actually extracting energy from the mean flow in order to grow. Locally, $-\mathbf{u}'\mathbf{b}' \cdot \nabla \bar{b} > 0$ can mean either that the eddies are extracting energy from the mean flow locally or that the eddies/waves are transporting variance from a location where they are extracting energy from the mean flow.

A notable aspect of the buoyancy anomalies observed in both Flat and Bowl is the westward tilt of the anomalies with height (see top right panels of Figure 3-12 and 3-13). Comparing the tilt of the buoyancy anomalies to the zonal mean zonal current shear at 60°N (shown in the left panel of Figure 3-7), it is apparent that the buoyancy anomalies tilt against the zonal current shear. In this section I will demonstrate that the westward tilt of the buoyancy anomalies with height allows the eddies to extract energy from the mean flow. Additionally, as shown in the right panel of Figure 3-7, the mean meridional potential vorticity gradient reverses sign with depth, a necessary condition for baroclinic instability.

As was demonstrated in the previous section, in order for eddies to create buoyancy variance, $\langle -\mathbf{u}'\mathbf{b}' \cdot \nabla \bar{b} \rangle$ must be greater than zero. Now we will see how the sign of this term depends on the phase tilt of the anomalies with height. Assume a streamfunction
of the form:

\[ \psi = \Psi(z) \sin(kx - \omega t + \phi(z)), \]  

(4.2)

where \( k \) is the (zonal or in the direction of propagation) wavenumber, \( \omega \) is the angular frequency, and \( \phi(z) \) is the phase tilt. We will proceed to calculate \(-\bar{v}' \cdot \bar{b}_y\) (subscripts denote partial derivatives): \( v' = \psi_z \) and by the hydrostatic relation \( b' = f \psi_z \), where \( f \) is the coriolis parameter. By the thermal wind relation \( \bar{u}_z = -\frac{1}{f} \bar{b}_y \). Therefore,

\[ -\bar{v}' \cdot \bar{b}_y = \frac{f^2 k}{2} \Psi^2 \bar{u}_z \phi_z. \]  

(4.3)

Thus, \(-\bar{v}' \cdot \bar{b}_y > 0\) if \( \bar{u}_z \) and \( \phi_z \) have the same sign, which occurs when the anomalies tilt in the opposite direction as the zonal current shear.

From the hydrostatic relation, we can estimate

\[ \Psi \approx \frac{1}{f} \frac{b'}{k_z}, \]

where \( k_z \) is the vertical scale of the anomalies. Therefore,

\[ -\bar{v}' \cdot \bar{b}_y \approx \frac{k}{2} \left( \frac{b'}{k_z} \right)^2 \bar{u}_z \phi_z. \]  

(4.4)

From our decadal mode, we estimate \( b' \approx 10^{-3} \text{m/s}^2 \), \( k \approx 2.5 \times 10^{-6} \text{m} \), \( k_z \approx 1/500 \text{m}^{-1} \), \( \phi_z \approx \frac{\pi}{2} \frac{\text{m}}{500 \text{m}} \), and \( \bar{u}_z \approx \frac{1 \text{cm/s}}{500 \text{m}} \). Plugging these numbers into Equation (4.4) we find:

\[ -\bar{v}' \cdot \bar{b}_y \approx 10^{-6} (\text{ms}^{-2})^2 / \text{yr}. \]

The estimate of \(-\bar{v}' \cdot \bar{b}_y\) based on the phase tilt of the anomalies matches the direct calculation quite well. While this approach of relating the tilt of the anomalies with height to the eddy creation of variance is quite elegant, it does not allow an easy calculation of the vertical term. However, the vertical term was shown to be negligible in the previous section.
Figure 4-5: The production of buoyancy variance by eddies in (a) Flat and (b) Bowl. The production of buoyancy variance by eddies $u'b' \cdot \nabla b$ (bottom right) is split into the zonal $-u'b' \cdot b_x$ (top left), meridional $-v'b' \cdot b_y$ (top right), and vertical $-w'b' \cdot b_z$ (bottom left) directions.
Flat: Eddy creation of buoyancy variance

(b) Bowl

Figure 4-5: Same as (a) but for Bowl.
4.2.2 Local Linear Stability Analysis

In the previous section we saw that eddies are able to extract energy from the mean flow, which allows buoyancy anomalies to grow despite damping by air-sea heat fluxes and mixing. Unfortunately, it is more difficult to conclude where in the domain the eddies are actually extracting energy from the mean flow in order to grow. Locally, $-\mathbf{u}^T \mathbf{b} \cdot \nabla b > 0$ can mean either that the eddies are extracting energy from the mean flow locally or that the eddies/waves are transporting variance from a location where they are extracting energy from the mean flow. In this section I will examine the local linear stability of the mean state for Flat and Bowl. I will show that in Flat the eastern and western boundaries are baroclinically unstable and in Bowl only the western boundary is baroclinically unstable.

In order determine the local linear stability at each horizontal location in the model, I will solve the linearized quasi-geostrophic potential vorticity (QGPV) equation linearized about the local climatological state. The setup of the problem proceeds exactly as the linear Rossby wave problem described in section 3.2.1. We assume a wave solution (see Equation (3.5)) to the linearized QGPV equation (3.2) and buoyancy equation (3.4). The result is the linear eigenvalue problem (3.6) for the normal modes $\hat{\psi}_n$ (the eigenfunctions) and the frequencies $\omega_n$ (the eigenvalues). Modes with eigenvalues that have a positive imaginary part are growing modes.

The vertically discrete problem (N layers) is derived in Smith (2007). The boundary conditions (3.6) are included by constructing a discrete stretching operator, $\gamma_{nm}$. For each location, the linear wave solutions are calculated for a range of $k$ and $l$. For each wavenumber, there are as many eigensolutions are there are vertical levels. The eigenvalue with the largest imaginary frequency (growth rate) is chosen. Typically only a few other eigenvalues have positive imaginary parts, most are stable. The code used for the linear stability analysis was provided by Ross Tulloch, and the same technique of solution is used as in Tulloch et al. (2011).

In this calculation the linear wave solutions are calculated for $k \in [-K_{max}, K_{max}]$ and $l \in [0, K_{max}]$, where $K_{max} = 2\pi/250$ km$^{-1}$, such that the length scale of the
largest wavenumbers is slightly smaller than the model resolution. The bottom slope was not included in the calculation, although this term may play a role in Bowl. Figure 4-6 shows the growth rate of the most unstable mode as a function of zonal \((k)\) and meridional wavenumber \((l)\) for 4 spatial locations in (a) Flat and (b) Bowl. Only scales inside the white circle are resolved by the model.

In both Flat and Bowl the grid-points adjacent to the eastern boundary (top right panels) near 60°N are strongly unstable. Growth rates faster than 100 days are observed, even at length-scales that are sufficiently large to be resolved in the coarse-resolution model. However, moving one grid-point away from the eastern boundary (bottom right panel), there is a marked difference between Flat and Bowl. In Flat, the instability also persists one grid-point away from the eastern boundary, although the wavenumber range for which the profile is unstable is narrower than for the grid-point next to the boundary. In Bowl, the instability (at scales resolved in the model) is almost entirely absent just one grid-point away from the eastern boundary. This result is in accord with the result that \(-u'b' \neq 0\) along the eastern boundary in Flat, but not in Bowl (see Section 4.2.1).

Thus the local instability analysis one grid-point away from the eastern boundary can explain the marked difference between the variability in Flat and Bowl. In Flat, this region is baroclinically unstable and radiates baroclinic Rossby waves. In Bowl, this region is stable at the length-scales resolved in the model. However, the instability immediately adjacent to the eastern boundary is indistinguishable between Flat and Bowl. It seems likely that the instability immediately adjacent to the boundary does not play a role in Bowl, since it disappears moving one grid-point away from the boundary. Certainly processes such as lateral friction along the boundary might act to damp the instability immediately adjacent to the boundary, preventing the instability next to the boundary from growing. Additionally, the presence of the bottom slope is likely to provide a stabilizing influence in Bowl, but the effects of a sloping bottom have not been included in this calculation. It is possible that including the bottom slope in the calculation for Bowl would stabilize the grid-point immediately adjacent to the eastern boundary.
However, there is another possible means of explaining the marked difference in the variability between Flat and Bowl that does not rely on different stability properties in the two models. In Flat the instability along the eastern boundary can radiate into the interior since the mean PV contours are latitude circles. In Bowl, bathymetry modifies the mean PV contours, making it difficult for the instability to radiate into the interior and it remains trapped on the boundary. This explanation is also consistent with the observed result that radiating instabilities on the eastern boundary are only observed in Flat. A brief review of theory describing radiating instabilities in presented in Section 4.2.3.

In both Flat and Bowl the grid-points adjacent to the western boundary (top left panels) near 60°N are unstable and the grid-points just inside the boundary (bottom left panels) are weakly unstable. The instability of the western boundary in both Flat and Bowl is in accord with the result that $-u'b' \cdot \nabla b > 0$ along the western boundary (see Section 4.2.1). Thus, in both Flat and Bowl, the western boundary is weakly baroclinically unstable. The instability is weak enough that it is damped away in the presence of realistic damping by air-sea heat fluxes, and it must be excited by stochastic atmospheric forcing (as shown for Bowl in section 4.1.3). While this instability is also present in Flat, it is weaker than the instability along the eastern boundary and thus is swamped by the baroclinic Rossby waves traveling from the eastern boundary.

4.2.3 Radiation of waves by Eastern boundary currents in flat bottomed models

We have seen that the eastern boundary is unstable in Flat, and it radiates baroclinic Rossby waves with large zonal and meridional wavelengths. In fact, the tendency of eastern boundary currents to radiate waves, especially in flat bottomed models, is a well-known phenomenon. Here I will summarize some of the theoretical and modeling work regarding eastern boundary currents and their stability.

Radiating instabilities occur when the phase speed and wavenumber of a distur-
Figure 4-6: Flat: The growth rate of the most unstable mode as a function of zonal \( (k) \) and meridional wavenumber \( (l) \) for 4 spatial locations near 60°N. Red colors are growth rates faster than 100 days. The left panels show points next to the western boundary (top) and one grid-point off the western boundary (bottom). The right panels show points next to the eastern boundary (top) and one grid-point off the eastern boundary (bottom).
Figure 4-6: Same as (a) but for Bowl.
bance within the unstable region match those of the freely propagating Rossby waves in the far field, allowing the local instability to propagate into the interior. The zonal group velocity of radiated waves must be away from the locally unstable region. Thus, eastern boundary currents radiate long Rossby waves \((K << K_1)\), which have a large westward group velocity and a slowly decaying amplitude envelope. On the other hand, western boundary currents radiate short Rossby waves \((K >> K_1)\), which have a small eastward group velocity and an amplitude that decays rapidly away from the boundary. Additionally, there is no long meridional wave cutoff for instability of an eastern boundary current (Hristova et al., 2008). Thus, we now can understand the reason for the large zonal scale of the baroclinic Rossby waves seen in Flat. While the eastern boundary is unstable at a large range of zonal wavenumbers, only the wavenumbers that couple with the Rossby waves in the basin can propagate into the interior.

In the absence of bathymetry, instabilities on the boundary can easily radiate into the interior, as the PV gradients are primarily in the meridional direction, enabling zonal propagation. However, when bathymetry that varies in the zonal/offshore direction is added, there are now mean PV gradients in the zonal direction, preventing the anomalies on the eastern boundary from propagating offshore and coupling with the free baroclinic Rossby waves in the interior (Wang, 2011).

4.2.4 Role of bathymetry

In this section I will examine why adding relatively modest bowl bathymetry leads to drastically different buoyancy variability in Flat and Bowl. Specifically, I will determine if it is important where the bathymetry is added or if adding any bathymetry changes the “pathological” behavior seen in Flat. By conducting several additional experiments with different bathymetry, I will show that bathymetry on the eastern boundary leads to the different behavior between the Flat and Bowl models.

I conducted four additional experiments with different bathymetries in order to better understand why the variability observed in Bowl is so different to that of Flat. The first experiment, which I will call Bowl-varNS, has bathymetry similar to that
of Bowl, but the bathymetry decreases in height northward so that the depth of the ocean is 3 km at the center of the basin for all latitudes (see the top left panel of Figure 4-7). The MOC and buoyancy variability in the subpolar gyre in Bowl-varNS is found to be almost identical to that of Bowl. A Hovmoller plot of the buoyancy anomalies in the small basin at 60°N for Bowl-varNS is shown in the leftmost panel of Figure 4-8. Just like in Bowl, the largest buoyancy anomalies are found along the western boundary.

Since in Bowl-varNS the depth of the ocean at the center of the basin is 3 km for all latitudes, the bathymetry on the eastern and western boundaries can be examined in isolation. In experiments BathymEB and BathymWB I retained only the bathymetry on the eastern or western boundaries, respectively (see bottom panels of Figure 4-7). Hovmoeller plots of the buoyancy anomalies in the small basin at 60°N for BathymEB and BathymWB are shown in the middle two panels of Figure 4-8. In BathymEB, like in Bowl, the largest buoyancy anomalies are restricted to the western boundary. In BathymWB, like in Flat, buoyancy anomalies originate near the eastern boundary and propagate westwards as baroclinic Rossby waves. Thus, it is the presence of bathymetry on the eastern boundary that is important for the change in the MOC and buoyancy variability between Flat and Bowl.

In a final experiment, which I will call Ridge, I added a ridge with a height of 1 km in the middle of the basin (see top right panel of Figure 4-7). A Hovmoeller plot of the buoyancy anomalies in the small basin at 60°N for Ridge is shown in the rightmost panel of Figure 4-8. Since there is no bathymetry on the eastern boundary of this experiment, buoyancy anomalies originate near the eastern boundary, as in Flat. However, due to potential vorticity constraints, the anomalies are modified as they travel westwards over the ridge.
Figure 4-7: Ocean geometry and depth (km) for Bowl-varNS (top left), Ridge (top right), BathymWB (bottom left), and BathymEB (bottom right). Two strips of land (white) extend from the north pole to 34°S, dividing the world ocean into a small basin, a large basin, and a zonally unblocked southern ocean.
Figure 4-8: Hovmoller plot of subsurface (depth of 265 m) buoyancy anomalies at 60°N as a function of longitude and time for experiments (from left to right) Bowl-varNS, bathymEB, bathyWB, and Ridge.
4.3 Baroclinic Instability and decadal variability in other modeling studies

In this section I will remark on other modeling studies that have cited baroclinic instability as an origin of decadal variability. Colin de Verdiere and Huck (1999) studied decadal variability of idealized ocean models with constant surface heat flux forcing. They found that interdecadal oscillations emerge without stochastic atmospheric forcing (as I found in Flat) and are driven by baroclinic instability of the mean state. In their model, the northwestern part of the basin is most unstable and eddies can extract energy from the mean flow to grow \( \langle -u'T' \cdot \nabla T' \rangle > 0 \). Huck et al. (2001) used a three-dimensional linear stability analysis to show that no unstable modes exist when the model is run with restoring boundary conditions, but for fixed flux boundary conditions a single unstable mode is observed. This mode bears a striking resemblance to the fully developed oscillations. Arzel et al. (2007) showed that an idealized coupled model exhibits similar interdecadal oscillations. The parallel between the decadal variability in my models, particularly Bowl, and the work of Huck, Arzel, and deVerdiere is striking.
Chapter 5

Conclusions

In this thesis I studied decadal MOC and buoyancy variability in coupled and ocean-only GCMs run in idealized geometries. By simplifying the geometry, I endeavored to create models that are simple enough that the mechanisms of decadal AMOC and buoyancy variability could be isolated, but are complex enough to exhibit interesting and perhaps realistic behavior. The main result of this work is that in these models decadal MOC variability is due to the thermal wind response to mid-depth buoyancy anomalies on the western boundary. This result is robust across all the models studied, despite their drastically different spatial and temporal patterns of buoyancy variability, and I believe that this result will hold in both more complex models and in the ocean. In section 5.1 I will summarize the mechanism by which buoyancy anomalies on the western boundary are created and how the MOC responds to these anomalies in my models. Then, in section 5.2 I will use the main results from my idealized GCM studies as a prism for understanding decadal MOC and buoyancy variability in more complex GCMs and data.
5.1 Summary of mechanisms of decadal MOC and buoyancy variability in idealized GCMs

In this section I will summarize the mechanisms of decadal buoyancy and MOC variability seen in my study of idealized GCMs. First, I will focus on the aspects of buoyancy and MOC variability that are robust amongst the models studied, as these results are likely to be the most relevant for understanding decadal variability in more complex GCMs and the ocean. Next I will outline the features which differ between the models, and comment on the realism of these features.

Decadal upper ocean buoyancy anomalies in the subpolar gyre are formed through ocean buoyancy transport convergence and are damped by air-sea buoyancy fluxes. Temperature and salinity anomalies are always found to compensate each other, and upper-ocean buoyancy anomalies are dominated by temperature. Negative (positive) subsurface potential vorticity (PV) anomalies accompany dense (buoyant) anomalies. Increased (decreased) convection is observed in regions of negative (positive) PV anomalies, as expected via preconditioning. While convection does not play a leading order role in creating the observed buoyancy anomalies, it does modify the T/S characteristics of the water masses.

Upon reaching the western boundary, the buoyancy anomalies are advected southward along the boundary by the deep western boundary current. Via the thermal wind relation, buoyancy anomalies on the boundaries result in anomalies in the shear of the zonally integrated meridional velocity. Buoyancy anomalies on the eastern boundary are observed to be negligible, except in the subpolar gyre, indicating that negative (positive) buoyancy anomalies on the western boundary lead to a spin up (down) of the MOC. The MOC is observed to respond passively to buoyancy anomalies on the western boundary: although variability of the MOC does lead to variability in the meridional transport of heat and salt, these transports are not responsible for creating the buoyancy anomalies on the western boundary that drive the MOC variability.

While the spatial structure of the buoyancy anomalies is found to change with model bathymetry, in all models studied the buoyancy variability is due to an ocean-
only mode. In Flat, the mode is weakly damped (large Q-factor), resulting in regular, predictable oscillations. In Bowl, the ocean-only mode is highly damped (small Q-factor) and must be excited by stochastic atmospheric variability, resulting in irregular, less predictable variability.

The origin, spatial pattern, and spectra of buoyancy variability varies with model bathymetry. In Flat the eastern boundary is baroclinically unstable and it radiates baroclinic Rossby waves. These waves travel westwards, taking 34 years to cross the basin. The time that it takes for the waves to cross the basin sets the dominant timescale of the buoyancy and MOC variability and are responsible for the highly predictable variability seen in Flat. Atmospheric variability is not needed to excite this mode. In Bowl buoyancy anomalies originate near the western boundary, which is weakly baroclinically unstable. The mode of buoyancy variability on the western boundary exists without stochastic atmospheric forcing, but it is destroyed with realistic damping. Atmospheric variability is needed to energize the mode of variability.

It is worth stressing that the mode of buoyancy variability seen in Flat is highly unrealistic. However, it is included here since it provides an excellent test case for understanding the response of the MOC to regular buoyancy anomalies which strike the western boundary. In the ocean-only framework a number of models have studied the response of the MOC to periodic variability on the western boundary (Kawase, 1987; Johnson and Marshall, 2002b,a; Deshayes and Frankignoul, 2005). In Flat, the coupled model internally generates almost periodic buoyancy variability, allowing us to examine the response of the MOC to buoyancy anomalies striking the western boundary. Due to the coarse resolution and highly idealized geometry of the model, the buoyancy variability in Bowl may also be unrealistic. The important point is that in both models, despite their drastically different buoyancy variability, the MOC responds to buoyancy anomalies on the western boundary of the subpolar gyre in the same way.

Large buoyancy anomalies are found near the western boundary, particularly in the subpolar gyre and along the boundary between the subtropical and subpolar gyres, in both more complex GCMs (Danabasoglu, 2008; Zhang, 2008) and in nature
(Kwon et al., 2010). This location is exactly where I find buoyancy anomalies to be important in changing the strength of the MOC in my idealized models. Thus, I expect that in nature, MOC variability is likely a response to buoyancy anomalies on the western boundary.

5.2 Outlook: Comparison to data and more complex GCMs

In this section I will use the main results from my idealized GCM studies as a prism for understanding decadal MOC and buoyancy variability in more complex GCMs and data. I will focus on four aspects of AMOC variability for which I believe my results can provide insight into AMOC variability in more realistic coupled models and data: (1) The relationship between AMOC and buoyancy anomalies. (2) The role of the AMOC in creating decadal buoyancy anomalies. (3) The role of the AMOC in meridional ocean heat transport. (4) The travel of AMOC anomalies down the western boundary.

5.2.1 Relationship between AMOC and buoyancy anomalies

While the spatial pattern and spectra of the buoyancy anomalies in Flat and Bowl differ substantially, in both models buoyancy anomalies along the western boundary (both in the subpolar gyre and along the boundary between the subtropical and subpolar gyres) are important in driving MOC variability. In this section I will show that in both NCAR CCSM3 and GFDL CM2.1, AMOC variability is associated with density anomalies in the subpolar gyre and along the boundary between the subtropical and subpolar gyres. As a result, when these anomalies reach the western boundary, they may travel southward in the DWBC and lead to AMOC variability according to the thermal wind relation. The analysis of decadal AMOC and density variability for CM2.1 and CCSM3 were carried out by Ross Tulloch, but the description of the potential mechanisms of variability and the relationship between the variability seen...
in CM2.1 and CCSM3 and my idealized models is primarily my own.

Both GFDL CM2.1 and NCAR CCSM3 exhibit decadal fluctuations of the AMOC. Figure 5-1 gives a brief synopsis on the AMOC variability in CCSM2 (top panels) and CM2.1 (bottom panels). The mean AMOC in each model is shown in the leftmost panels. Timeseries of the AMOC in a box between $35^\circ$N and $50^\circ$N from 400 and 2000 m depth (box shown in black in leftmost panels) are shown in the middle panels. Both models exhibit variability of the AMOC on the order of a few Sverdrups. The rightmost panels show the powerspectra of the AMOC index. AMOC variability in CCSM3 has a broad peak centered at timescale of 23 years, and AMOC variability in CM2.1 has a narrower peak centered at a timescale of 22 years. A number of studies concern AMOC variability in CM2.1 (Zhang et al., 2007; Zhang, 2008, 2010; Msadek et al., 2010) and CCSM (Danabasoglu, 2008). The goal here is not to repeat these analyses, but rather to show how the results of my idealized GCM studies might provide insight for understanding of the variability in these models.

Figures 5-2 and 5-3 show the density anomalies (averaged over the top 1000m) that are associated with AMOC variability in CCSM3 and CM2.1, respectively. HIGH and LOW MOC years are defined as years where the AMOC timeseries is more than one standard deviation above (below) its mean value (shown by the red and blue x’s in Figure 5-1). Spatial patterns of density variability are created by taking composites of HIGH MOC minus LOW MOC years, shifted by -10, -5, -1, 0, 1, and 5 years. In both models there are significant density anomalies on the boundary between the subtropical and subpolar gyres.

In CCSM relatively small scale density anomalies originate on the boundary between the subtropical and subpolar gyres near 40°W. These anomalies propagate westward into the Labrador Sea, where they are hypothesized to modify the AMOC by modifying deep convection (Danabasoglu, 2008). In CM2.1 large-scale density anomalies originate in the subpolar gyre. These anomalies also appear to propagate westwards and divide into two anomalies, a southerly anomaly that travels along the Gulf Stream path and a northerly anomaly that enters the Labrador Sea. Zhang (2008) hypothesizes that these density anomalies are created by changes in the
Figure 5-1: Mean AMOC and AMOC variability in CCSM2 (top panels) and CM2.1 (bottom panels). The mean AMOC in each model is shown in the leftmost panels. Timeseries of the AMOC in a box between 35°N and 50°N from 400 and 2000 m depth (box shown in black in leftmost panels) are shown in the middle panels. Red (blue) x’s indicate years for which the AMOC timeseries is more than one standard deviation above (below) its mean value. The rightmost panels show the powerspectra of the AMOC index. Figure from Tulloch and Marshall (in prep).
strength of the AMOC and DWBC (as will be described in the following section). However, she does not attempt to explain the origin of the AMOC variability, other than noting that it is linked to variability in deep convection.

Insight from studying MOC variability in my idealized GCMs suggests that considering the response of the AMOC to density anomalies on the western boundary would be useful when attempting to understand modes of AMOC variability. Density anomalies may influence AMOC variability both through the thermal wind response to density anomalies that travel down the western boundary and by forcing changes in deep convection through preconditioning. Furthermore, both complex GCMs, such as GFDL CM2.1 and CCSM3, and observations (see Figure 5-4) show significant decadal density variability in the western part of the subpolar gyre, precisely in the regions where I have shown the MOC to be sensitive to decadal density anomalies.

5.2.2 Role of the MOC in creating decadal buoyancy anomalies

In both Flat and Bowl the MOC passively responds to buoyancy anomalies on the boundaries. Although ocean heat and freshwater transport anomalies are associated with decadal MOC variability, these transport anomalies do not play a leading order role in creating decadal buoyancy anomalies. This conclusion is contrary to notion that AMOC variability plays an active role in climate variability on decadal timescales, a notion which pervades literature on MOC variability. However, the hypothesis that observed decadal SST anomalies are the result of changes in the strength of the AMOC is just that, a hypothesis. The observational evidence that supports this hypothesis are (1) In the mean a large portion of the ocean heat transport in the Atlantic is achieved by the deep meridional overturning circulation and (2) Large scale-decadal SST anomalies, generally of a singular polarity over the entire basin, have been observed in the North Atlantic. Both of these facts give plausibility to the idea that decadal SST anomalies might be due to changes in the strength of the AMOC, but certainly do not prove that this is the case.
Figure 5-2: Density anomalies (g/cm$^3$, averaged over the top 1000m) that are associated with AMOC variability in CCSM3. HIGH and LOW MOC years are defined as years where the AMOC timeseries is more than one standard deviation above (below) its mean value. Spatial patterns of density variability are created by taking composites of HIGH MOC minus LOW MOC years, shifted by -10, -5, -1, 0, 1, and 5 years. Black contours are zeros of the mean windstress curl. Figure from Tulloch and Marshall (in prep).
Figure 5-3: Density anomalies (kg/m³, averaged over the top 1000m) that are associated with AMOC variability in CM2.1. HIGH and LOW MOC years are defined as years where the AMOC timeseries is more than one standard deviation above (below) its mean value. Spatial patterns of density variability are created by taking composites of HIGH MOC minus LOW MOC years, shifted by -10, -5, -1, 0, 1, and 5 years. Black contours are zeros of the mean windstress curl. Figure from Tulloch and Marshall (in prep)
Figure 5-4: Standard deviation of JanMar SST from NCEPNCAR reanalysis I for 1948-2006. SST is low-pass filtered to retain periods greater than 5 years. Contour intervals are 0.1°C. From Kwon et al. (2010).

As a result of the lack of MOC and OHT observations on decadal timescales, models have been used to probe the relationship between decadal buoyancy anomalies and AMOC variability. While many studies invoke the idea that decadal buoyancy (or temperature) anomalies are due to AMOC variability, only a few attempt to isolate the role that ocean heat and freshwater transports that are associated with the AMOC play in buoyancy budgets. Not surprisingly, whether or not the AMOC plays an active role in creating decadal buoyancy anomalies differs between models.

In HadCM3 and GFDL CM2.1 the AMOC is argued to play an active role in creating decadal buoyancy (or temperature) anomalies. HadCM3 shows large-scale temperature anomalies with a single polarity over the entire North Atlantic associated with AMOC variability (see Figure 1-9), similar to the SST anomalies seen in observations (see Figure 1-3). Knight et al. (2005) asserts (although does not prove) that these SST anomalies are due to ocean heat transport anomalies associated with AMOC variability. In GFDL CM2.1, cooling along the Gulf Stream path and warming in the subpolar gyre is associated with a positive anomaly in the AMOC (see temperature anomalies in left panel of Figure 1-11 or corresponding density anomalies in bottom left panel of Figure 5-3). Zhang (2008) argues that the temperature
(or density) anomalies are related to changes in the AMOC. Increased deep water formation leads to a strengthening of the deep western boundary current (DWBC) and the AMOC. The increased DWBC leads to a strengthening of the northern recirculation gyre (NRG) and a southward displacement of the Gulf Stream, leading to a divergence of ocean mass and heat transport and hence cooling (increased density) along the Gulf Stream path. At the same time, a denser DWBC leads to a reduction of the mixed layer depth in the subpolar gyre, leading to higher sea surface heights, a weaker subpolar gyre, and a convergence of ocean heat transport, hence warming (decreased density) in the subpolar gyre.

In NCAR CCSM3 MOC variability does not play an active role in creating the observed decadal density anomalies. In CCSM3 the spatial pattern of density anomalies associated with the AMOC is an east-west dipole along the boundary between the subtropical and subpolar gyres (see Figure 5-2). The density anomalies have a much smaller scale and larger magnitude at the surface than the large-scale patterns found in HadCM3 or CM2.1, with maximum SST anomalies of 6 – 7°C. Danabasoglu (2008) argues that the decadal density anomalies are not associated with ocean heat and freshwater transport anomalies due to AMOC variability. Instead, density anomalies are created by oscillations in the subtropical-subpolar gyre boundaries driven by windstress curl anomalies related to the NAO. These density anomalies are associated with the AMOC because when they propagate into the Labrador Sea, they lead to changes in convection and deep water formation and hence changes in the strength of the DWBC.

Despite the many studies that suggest that AMOC variability leads to decadal temperature/density anomalies, whether the AMOC plays an active role in creating decadal density anomalies has not been firmly established, either through observations or modeling results. As was shown here, some models suggest the AMOC plays an active role in creating decadal density anomalies (Knight et al., 2005; Zhang, 2008), while others suggest it does not (Danabasoglu, 2008). Thus, the relevance of studies that isolate the role (or lack thereof) that the MOC plays in creating decadal density anomalies are essential for a better understanding of decadal variability.
5.2.3 Role of the AMOC in Ocean Heat Transport Variability

On interannual timescales and spatial scales on the order of the meridional variation of the winds (20°), much of the ocean heat transport variability in both the Atlantic and the Pacific is due to overturning cells forced by Ekman transport variability (Jayne and Marotzke, 2001). Variability of the deep, inter-hemispheric AMOC is believed to play a role in OHT variability on longer timescales and larger spatial scales (Dong and Sutton, 2003; Shaffrey and Sutton, 2004, 2006; Danabasoglu, 2007; Zhang, 2008). In particular, OHT anomalies that are coherent across a large range of latitudes are often assumed to be forced from the northern latitudes by deep convection and associated with the deep AMOC.

However, in both Flat and Bowl I found that the OHT anomalies associated with the AMOC are actually quite modest when compared to the full annual mean OHT anomalies. Perhaps more surprisingly, annual OHT anomalies are quite coherence with latitude, even on timescales as short as a few years. Thus, my results call into question the often made assumption that AMOC and OHT anomalies that are coherent with latitude are due to variability in the deep-interhemispheric AMOC. In this section I will propose an alternative explanation for these latitudinally coherent OHT anomalies. In a future work, I would like to test this hypothesis in both my idealized GCMs and more realistic models.

The leading mode of low-frequency variability in the tropical Atlantic is the co-varying fluctuation of tropical SST gradients and trade winds (Chang et al., 1997; Marshall et al., 2001b), which has been termed the “Atlantic Meridional Mode”\(^1\). A negative anomaly in the cross-equatorial temperature gradient is associated with increased trade winds in the Northern Hemisphere and decreased trade winds in the Southern Hemisphere, leading to northward Ekman transport and hence northward OHT in the tropics of both hemispheres. If this state is combined with a positive

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\(^1\)This mode of variability has also been called “Tropical Atlantic Variability (TAV)” and the “Atlantic SST Dipole”
state of the North Atlantic Oscillation\(^2\), which is associated with increased and pole-ward shifted easterlies and westerlies in the Northern Hemisphere, northward OHT anomalies that are coherent from 20°N to 40°N latitude would result. Interestingly, these OHT anomalies are associated with inter-hemispheric MOC anomalies due to Ekman transport anomalies and a barotropic return flow, but these MOC anomalies are unrelated to water mass formation in the polar regions classically thought to drive large-scale MOC and OHT variability on decadal timescales (Delworth et al., 1993; Dong and Sutton, 2005; Mignot and Frankignoul, 2005; Msadek and Frankignoul, 2009).

### 5.2.4 Travel of AMOC anomalies down the western boundary

In Flat and Bowl, buoyancy anomalies that reach the western boundary are advected southward by the DWBC. It takes the buoyancy anomalies approximately 8 years to travel from 40°N to the equator, consistent with an advective timescale. Since the buoyancy anomalies on the eastern boundary are negligible south of 40°N, via the thermal wind relation, MOC anomalies also take approximately 8 years to reach the equator.

The advective timescale for travel of buoyancy (or PV) anomalies down the western boundary is consistent with observational studies. Curry et al. (1998) find that anomalies in the temperature and thickness of the Labrador Sea Water (LSW) layer take approximately 6 years to reach the deep water of the subtropical basin near Bermuda. Additionally, temperature and thickness anomalies associated with LSW formation spread to the tropics primarily along the western boundary. Analysis of PV anomalies at Line W along the continental slope south of New England (near 40°N, 70°W) demonstrates that PV anomalies take 6-9 years to travel from the Labrador Sea to Line W (Pena-Molino, 2010), implying a spreading rate of 1.5 to 2.5 cm/s, somewhat slower than that observed by Curry et al. (1998).

\(^2\)A number of studies have linked decadal variability of the Atlantic Meridional Mode and the NAO (Czaja et al., 2002; Marshall et al., 2001b; Xie and Tanimoto, 1998)
Despite the observational consensus that PV anomalies travel down the DWBC at timescales that are consistent with advection, a number of theoretical and modeling studies suggest that Kelvin waves play a role of communicating anomalies in deep water formation southward along the western boundary. (Kawase, 1987) considered the response of a two layer model to flow prescribed across the interface between the layers. In the limit of weak damping, the flow spins up to the Stommel-Arons state with a DWBC crossing the equator and poleward flow in the interior. The flow is set up by Kelvin waves which travel southward along the western boundary, eastward along the equator, and polewards along the eastern boundary in both hemispheres. In the second stage of the flow evolution, the eastern boundary currents disperse and the equatorial boundary current broadens as a result of long Rossby waves radiating from the eastern boundary into the interior.

Johnson and Marshall (2002b,a) and Deshayes and Frankignoul (2005) applied Kawase's ideas to study the response of the AMOC to changes in deep water formation. These studies used reduced gravity models to represent the upper limb of the AMOC and parameterized the effects of deep water formation by imposing a variable volume flux out of the model’s surface layer at high latitudes. They found that the anomaly of the surface layer thickness propagates southward along the western boundary as a Kelvin wave and eastward along the equator as an Equatorial Kelvin wave. The subsequent poleward propagation of Kelvin waves on the eastern boundary leads to the generation of westward propagating long Rossby waves that drive the interior flow. Thus, the equator acts as a low-pass filter because Kelvin waves on the western boundary cannot cross the equator, and the flow in the Southern Hemisphere can only be set in motion by long Rossby waves, which have decadal timescales.

Zhang (2010) studied the latitudinal dependence of AMOC anomalies in GFDL CM2.1. She argued that AMOC anomalies travel at the advective speed from the Flemish Cap to Cape Hatteras, but travel at the coastal Kelvin wave speed between Cape Hatteras and the equator.

Thus, the role of advection and waves in communicating AMOC anomalies southward is not yet resolved. My model results suggest that examining the travel of
buoyancy anomalies along the western boundary, rather than a zonally integrated quantity such as the AMOC, might lead to a better understanding of the means by which AMOC signals get communicated.

5.3 Implications

As outlined in the previous section, the decadal variability observed in my idealized GCMs has many similarities with observations and studies of more complex GCMs. MOC variability is associated with decadal buoyancy variability in the subpolar gyre. In GFDL CM2.1 and NCAR CCSM3 significant buoyancy variability in the subpolar gyre and on the boundary between the subtropical and subpolar gyres is associated with AMOC variability. Additionally, low-frequency SST anomalies from observations are also maximal in this region Kwon et al. (2010). The MOC responds to buoyancy anomalies on the western boundary in accord with the thermal wind relation, as has been demonstrated by the RAPID array. Negative PV anomalies on the western boundary are associated with a stronger deep western boundary current and an enhanced MOC, in accord with observational and modeling studies.

However, in my idealized model studies, the causal relationship between buoyancy anomalies and MOC anomalies is reversed. The prevailing idea which has permeated studies of decadal MOC and buoyancy variability is that the MOC plays an active role in creating decadal buoyancy anomalies. AMOC variability leads to ocean heat transport variability, whose convergence leads to decadal buoyancy anomalies. The origin of the MOC variability differs between studies, but is often linked to variability in convection and water mass formation. However, in my study, decadal buoyancy anomalies lead to MOC variability. Buoyancy anomalies on the western boundary are advected southward by the deep western boundary current, and lead to MOC variability according to the thermal wind relation. Buoyancy anomalies also lead to changes in stratification and hence PV anomalies, which lead to changes in deep convection as expected from preconditioning. The MOC responds passively to buoyancy anomalies on the western boundary: while MOC variability leads to changes in
ocean heat and freshwater transport, these transports do not play a significant role in creating the buoyancy anomalies in the subpolar gyre that are responsible for the changes in the MOC.

Schematics of the “traditional” active AMOC paradigm and the passive AMOC paradigm suggested by this study are shown in Figure 5-5. If the AMOC is truly passive, knowledge of MOC variability in the subtropical gyre will not enable the prediction of decadal SST anomalies.\(^3\) In nature, the AMOC may play a more active role in creating decadal buoyancy anomalies than it does in this model. However, my model studies highlight the need for studies that examine the role (or lack thereof) that meridional ocean heat transport anomalies associated with the AMOC play in creating decadal buoyancy anomalies. Additionally, via the thermal wind relation, buoyancy anomalies on the boundaries lead to MOC variability. While this fact has been appreciated in studies which attempt to diagnose MOC variability, such as RAPID, the response of the MOC to buoyancy anomalies has rarely been highlighted in studies of decadal variability. Since decadal buoyancy anomalies in the subpolar gyre appear to be essential to understanding decadal variability, observing systems for making decadal predictions should monitor upper ocean buoyancy anomalies in the subpolar gyre in addition to directly observing MOC variability.

\(^3\)For a highly periodic mode of variability such as seen in Flat, knowledge of decadal MOC variability does appear to allow prediction of decadal buoyancy anomalies, but this is just because the buoyancy anomalies that the MOC responds to are very regular.
"Active" MOC

Convective variability $\rightarrow$ PV anomalies $\rightarrow$ MOC variability $\rightarrow$ Decadal Buoyancy anomalies

"Passive" MOC

PV anomalies $\rightarrow$ Convective variability

Decadal Buoyancy anomalies

Thermal wind $\rightarrow$ MOC variability

Figure 5-5: Schematic of the causal relationship between buoyancy variability, MOC variability, and convection in which (top) the MOC variability is "active" in driving buoyancy anomalies and (bottom) the MOC is "passive" and responds to buoyancy anomalies impinging on the western boundary.
Bibliography


