A Simple Model of the Response of the Atlantic to the North Atlantic Oscillation

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The response of an idealised Atlantic ocean to wind and thermohaline forcing associated with the North Atlantic Oscillation (NAO) is investigated both analytically and numerically in the framework of a reduced-gravity model. The NAO-related wind forcing is found to drive a time-dependent “leaky” gyre circulation that integrates basin-wide stochastic wind Ekman pumping and initiates low-frequency variability along the western boundary. This is subsequently communicated, together with the stochastic variability induced by thermohaline forcing at high latitudes, to the remainder of the Atlantic via boundary and Rossby waves. At low frequencies, the basin-wide ocean heat content changes owing to NAO wind forcing and thermohaline forcing are found to oppose each other. The model further suggests that the recently reported opposing changes of the meridional overturning circulation in the Atlantic subtropical and subpolar gyres between 1950-1970 and 1980-2000 may be a generic feature caused by an interplay between the NAO wind and thermohaline forcing.
1. Introduction

The North Atlantic Oscillation (NAO) is the dominant mode of atmospheric variability in the North Atlantic sector (e.g. Hurrell 1995), and has a significant impact on the North Atlantic ocean circulation and hydrographic properties through its modulation of air-sea momentum, heat and freshwater fluxes. During the last 50 years, the NAO has exhibited large interannual and decadal variability, switching from its negative phase in the 1960s to strong positive phase in the late 1980s and early 1990s.

Hindcast simulations using ocean general circulation models (OGCMs) have been widely used to study the oceanic response to changes of the NAO (e.g. Häkkinen 1999; Eden and Willebrand 2001; Eden and Jung 2001; Eden and Greatbatch 2003; Dong and Sutton 2005; Böning et al. 2006; Lozier et al. 2008; Deshayes and Frankignoul 2008; Robson et al. 2012; Zhai and Sheldon 2012). During positive NAO years, the anomalous wind stress is generally found to spin up the subtropical and subpolar gyre circulation, whereas the enhanced heat loss at high latitudes, particularly in the Labrador Sea, leads to a strengthening of the meridional overturning circulation (MOC) (see Visbeck et al. 2003, for a review). Although significant progress has been made over the last decade, detailed adjustment mechanisms remain unclear due, in part, to the complex nature of the OGCMs as well as to the lack of observations.

In a theoretical study, Marshall et al. (2001) argued that the meridional shift in the jet stream associated with the NAO, through anomalous wind stress curl, drives an anomalous “intergyre gyre” at midlatitudes between the climatological subtropical and subpolar gyres. The anomalous wind forcing of this intergyre gyre is coherent in space but stochastic in time, which initiates upper ocean buoyancy anomalies that propagate westward as baroclinic Rossby waves. As a result, the strength of the intergyre gyre fluctuates in time, albeit at much lower frequencies than that of the imposed forcing. However, it is not clear what happens

\(^1\)We hereafter describe the NAO as a stochastic forcing, although the spectrum of the NAO index is slightly red (e.g. Wunsch 1999).
to these buoyancy anomalies when they arrive at the western boundary. It is possible that
boundary waves may be excited that propagate southward along the western boundary, and
after that communicate the buoyancy anomalies into the ocean interior in a similar fashion as
described by Johnson and Marshall (2002a) for the upper ocean adjustment to thermohaline
forcing.

Zhai et al. (2011) studied the Atlantic heat content and sea level change in response
to stochastic deep-water formation at high latitudes in the framework of a reduced-gravity
model. They identified the “Rossby buffer” effect: high-frequency basin-wide anomalies in
heat content are confined to low latitudes whereas low-frequency anomalies extend to mid
and high latitudes in both hemispheres. However, the ocean heat content change in response
to the wind forcing was left unexplored. There are a number of theoretical investigations on
oceanic adjustment to idealised wind forcing (e.g. Frankignoul et al. 1997; Cessi and Louazel
2001; Primeau 2002; Cessi and Otheguy 2003; Sirven et al. 2007), but the focus has been on
the thermocline variability, with little attention being paid to the response of the MOC and
ocean heat content, in particular to the wind forcing associated with the NAO. Furthermore,
there is the issue of the interplay between the NAO wind forcing and thermohaline forcing
in determining the ocean heat content and transport variability.

The present study aims to investigate the response of an idealised Atlantic ocean to
wind and thermohaline forcing associated with the NAO in the context of a reduced-gravity
model. The paper is organised as follows. In Section 2, a quantitative theory is developed
for the ocean response to NAO-related wind forcing. In Section 3, the reduced-gravity
model experiments and results are described. In Section 4, we use the theory to explore
the possible mechanism behind the recently-reported opposing changes of the MOC in the
Atlantic subtropical and subpolar gyres. Finally, in Section 5, we summarize our key findings.
2. Theory

In this section we focus on providing a simple linear theory of the ocean response to the NAO-related wind and high-latitude thermohaline forcing in the framework of a reduced-gravity model. Readers are referred to Johnson and Marshall (2002a) and Zhai et al. (2011) for related theory of ocean heat content and MOC variability induced purely by high-latitude thermohaline forcing.

We consider a semi-enclosed rectangular domain that is open to the south, with $0 < x < L_x$ in the zonal direction and $0 < y < L_y$ in the meridional direction (Fig. 1). As the NAO wind forcing is dominated by the zonal wind stress anomaly that is largely uniform in the zonal direction over the North Atlantic, we hereafter consider only the zonally-averaged zonal wind stress anomaly $\tau_x(y, t)$, and we further assume $\tau_x(y, t)$ vanishes approaching the northern and southern boundaries. Ekman pumping associated with this simplified NAO wind forcing is thus uniform in space in the zonal direction but stochastic in time.

Outside the western boundary layer, the linear dynamics are described by the momentum and continuity equations:

$$-fv + g\frac{\partial h}{\partial x} = \frac{\tau_x}{\rho_0 H},$$

$$fu + g\frac{\partial h}{\partial y} = 0,$$

$$\frac{\partial h}{\partial t} + H \left( \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} \right) = 0.$$

Here $u$ and $v$ are the zonal and meridional velocities, $h$ is the anomaly in the layer thickness from its initial value $H$, $f(y)$ is the Coriolis parameter, $g'$ is the reduced gravity and $\rho_0$ is a reference density.

Although presented in Cartesian coordinates and a rectangular basin for pedagogical simplicity, the theory is easily generalized to spherical geometry and realistic basin geometries. The solutions presented in Section 3 are obtained in a rectangular basin on the sphere, and in Section 4 in a realistic Atlantic basin geometry.

\(^2\)Anomalies hereafter refer to deviations from time-means.
a. Volume budget

Substituting for $u$ and $v$ in (3) using (1) and (2) gives the forced long Rossby wave equation:

$$\frac{\partial h}{\partial t} - c(y) \frac{\partial h}{\partial x} = -w_E(y, t),$$

(4)

where

$$w_E(y, t) = -\frac{\partial}{\partial y} \left( \frac{\tau_x(y, t)}{\rho_0 f} \right)$$

(5)

is the Ekman upwelling velocity,

$$c(y) = \frac{\beta(y) g' H}{f(y)^2} = -\frac{\partial}{\partial y} \left( \frac{g'H}{f(y)} \right)$$

(6)

is the long Rossby wave speed, and $\beta(y) = df(y)/dy$ is the meridional gradient in the Coriolis parameter.

The solution to (4) is obtained by integrating westward along the Rossby wave characteristics (e.g. Frankignoul et al. 1997; Cessi and Louazel 2001):

$$h(x, y, t) = h_e \left( t - L_x - \frac{x}{c(y)} \right) - \int_{t - L_x - \frac{x}{c(y)}}^{t} w_E(y, t') dt',$$

(7)

where $x = L_x$ is the longitude of the eastern boundary. In (7) we assume that the upper layer thickness anomaly along the eastern boundary, $h_e(t)$, is uniform due to fast boundary wave propagation (Johnson and Marshall 2002a; Marshall and Johnson 2013).

Following Johnson and Marshall (2002a) and zonally-integrating (4) from just outside the western boundary current region, $x = \delta$, to the eastern boundary, we obtain

$$\frac{\partial}{\partial t} \int_{\delta}^{L_x} h dx = c(y) [h_e(t) - h_b(y, t)] - w_E(y, t)L_x,$$

(8)

where

$$h_b(y, t) = h_e \left( t - \frac{L_x}{c(y)} \right) - \int_{t - \frac{L_x}{c(y)}}^{t} w_E(y, t') dt'.$$

(9)
is the upper layer thickness anomaly just outside the western boundary current and $L_x/c(y)$ is the Rossby wave basin-crossing time. Physically (8) indicates that the change of zonally-integrated layer thickness at any given latitude is caused by wind Ekman pumping integrated across the basin as well as the difference between the thickness anomalies propagated by Rossby waves from the eastern boundary into the ocean interior, and propagated from the ocean interior into the western boundary current.

A further relation for the rate of change of layer thickness integrated across the basin is obtained by zonally integrating (3):

$$\frac{\partial}{\partial t} \int_0^{L_x} h \, dx = - \frac{\partial T(y, t)}{\partial y},$$  \hspace{1cm} (10)

where

$$T(y, t) = \int_0^{L_x} vH \, dx$$  \hspace{1cm} (11)

is the net northward volume transport.

Now combining (8) and (10), and neglecting the volume anomaly of the narrow western boundary region, gives:

$$\frac{\partial T(y, t)}{\partial y} = - c(y) [h_e(t) - h_b(y, t)] + w_E(y, t) L_x.$$  \hspace{1cm} (12)

Integrating (12) over the latitudinal extent of the domain gives:

$$\int_0^{L_y} c(y) [h_e(t) - h_b(y, t)] \, dy = T_S(t) - T_N(t),$$  \hspace{1cm} (13)

where $T_N$ is the prescribed northward transport anomaly at the northern boundary associated with thermohaline forcing,

$$T_S(t) = \frac{g' H}{f_s} h_e(t)$$  \hspace{1cm} (14)

is the northward transport anomaly across the southern boundary, $f_s = f(0)$ is the Coriolis parameter at the southern boundary and we have used that

$$\int_0^{L_y} w_E(y, t) \, dy = 0$$  \hspace{1cm} (15)
since $\tau_x$ vanishes at $y = 0$, $L_y$. In deriving (13), we have assumed that the layer thickness anomaly in the southwestern corner of the domain vanishes, which is justified on the time scales considered in this paper due to the long adjustment time scales of the Pacific and Indian basins (Johnson and Marshall 2004).

Finally, combining (9), (13) and (14), we obtain an equation for the eastern boundary layer thickness that represents the generalization of equation (14) in Johnson and Marshall (2002a) to include wind forcing:

$$h_e(t) = \frac{1}{\alpha} \left[ \int_0^{L_y} \tilde{c}(y) h_e \left( t - \frac{L_x}{c(y)} \right) dy + \int_0^{L_y} \tilde{c}(y) \int_{t-L_x/c(y)}^t w_E(y, t') dt' dy \right]$$

$$- \frac{1}{\alpha} T_{\text{ND}},$$

where

$$\alpha = \int_0^{L_y} \tilde{c}(y) dy - \frac{g' H}{f_s}.$$  

The Rossby wave speed,

$$\tilde{c}(y) = \min \left( c(y), \frac{\sqrt{g' H}}{3} \right)$$

is capped at the equatorial Rossby wave speed to prevent the integrals from diverging in (16).³ The terms on the right hand side of (16) represent radiation of Rossby waves from the eastern boundary providing the memory of past wind and thermohaline forcing, basin-wide wind-forced Ekman pumping, and outflow associated with thermohaline forcing at

³As in Johnson and Marshall (2002a), the solution is insensitive to the precise value at which $c$ is capped. To understand this result, note that when $c(y)$ becomes large, the integrand on the left-hand side of (13) is
high latitudes, respectively. Notwithstanding the simplicity of the reduced-gravity model, (16) can be used to explore the relative importance of, and interplay between, wind and thermohaline forcing associated with the NAO in driving MOC and heat content anomalies, as detailed in the following sections.

b. MOC anomaly

The northward transport can be decomposed into Ekman and geostrophic components,

\[ T(y, t) = T_{Ek}(y, t) + T_g(y, t) \]
\[ = \frac{\tau_x(y, t)L_x}{\rho_0 f(y)} + \frac{g'H}{f(y)}[h_e(t) - h_w(y, t)]. \]

(17)

Integrating (12) southward from the northern boundary, the geostrophic northward transport is given by

\[ T_g(y, t) = T_N(t) + \int_{L_y-y}^{L_y} \tilde{c}(y') [h_e(t) - h_b(y', t)] \, dy'. \]

(18)

The MOC variability at any latitude can thus be determined from the basin-wide wind forcing, high-latitude thermohaline forcing, and the history of the eastern boundary layer thickness anomaly. Notice that the wind forcing influences the MOC transport through both modifying the layer thickness anomaly communicated by Rossby waves into the western boundary current and directly driving an Ekman transport. Substituting for \( h_b \), we obtain approximately

\[ c(y) [h_e(t) - h_b(y, t)] = c(y) \left[ \left( \frac{\partial h_e(t)}{\partial t} + w_E(y, t) \right) \frac{L_x}{c(y)} + \cdots \right] \]
\[ = \left( \frac{\partial h_e(t)}{\partial t} + w_E(y, t) \right) L_x + \cdots, \]

independent of \( c(y) \). Thus, the solution of (13) is unaffected by replacing \( c(y) \) by \( \tilde{c}(y) \), even though the individual contributions associated with \( h_e(t) \) and \( h_b(y, t) \) in the integrand both diverge unless \( c(y) \) is capped.
the expressions for the northward geostrophic transport across any latitude,

\[ T_g(y, t) = T_N(t) + \int_{L_y-y}^{L_y} \tilde{c}(y') \left[ h_e(t) - h_e \left( t - \frac{L_x}{\tilde{c}(y')} \right) \right] dy' + \int_{t-\tilde{c}(y')}^{t} w_E(y', t') dt' \]

(19)

and the western boundary thickness anomaly,

\[ h_w(y, t) = h_e(t) - \frac{f(y) T_g(y, t)}{g'H} \] 

(20)

Here we have derived a similar expression for \( h_w \) to Cessi and Louazel (2001), but independent of the details of the momentum balance in the narrow frictional western boundary layer.

It is worth pointing out that Equation (19) shows that the geostrophic transport, \( T_g(y, t) \), at any given latitude depends on the entire history of the wind forcing integrated over the whole model domain through the dependence of \( h_e \) on this basin-wide forcing in (16). In particular, \( T_g(y, t) \) depends strongly on the history of wind Ekman pumping integrated over the whole area poleward of the latitude under consideration. As such, the wind-induced geostrophic transport variability at any given latitude cannot be diagnosed from a linear Rossby wave model driven by wind forcing at that single latitude alone. The same is true for the thermocline depth on the western boundary.

Furthermore, since the sea level anomaly is proportional to the upper layer thickness anomaly in the reduced-gravity model, (20) shows that the sea level variability along the east coast of the United States and Canada depends on the history of wind forcing integrated over the whole ocean basin at higher latitudes to the north and thus cannot be predicted using models driven by wind forcing at a single local latitude such as Sturges and Hong (1995) and Frankignoul et al. (1997).

There are two geographic regimes associated with (17): regions under direct NAO wind forcing and regions to the south. At latitudes under direct influence of stochastic NAO wind forcing, the Ekman contribution in (17) is also stochastic at any given latitude, and vanishes.
at the latitude where $\tau_x = 0$, forming opposing meridional transport anomalies north and south of that latitude. The geostrophic transport anomaly, $T_g$, at any given latitude, on the other hand, depends on stochastic Ekman pumping integrated over the whole area further to the north and thus varies at much lower frequencies.

In regions to the south of the NAO wind forcing, the ocean feels only the low-frequency $h_w$ that integrates stochastic Ekman pumping over the whole intergyre gyre area. The low-frequency $h_w$ then propagates equatorward along the western boundary, eastward along the equator and poleward along the eastern boundary, followed by the slow radiation of the Rossby waves off the eastern boundary. This pattern is in contrast to the ocean’s response to stochastic deep-water formation at high latitudes, where the upper layer thickness anomaly along the western boundary exhibits both high-frequency and low-frequency variability, both of which spread into the interior of the model domain through boundary and Rossby wave adjustment (Zhai et al. 2011). Consequently, the MOC in the thermohaline-forced case varies at all frequencies, although the high-frequency component tends to be confined in the hemisphere in which it is generated (Johnson and Marshall 2002b).

c. Ocean heat content anomaly

In the reduced-gravity model, the ocean heat content change at each latitude, $\mathcal{H}$, is proportional to the zonally-integrated layer thickness anomaly,

$$\mathcal{H}(y, t) = \rho_0 c^0_p \Delta \Theta \int_0^{L_x} h(x, y, t) \, dx,$$

(21)

where $\Delta \Theta$ is the Conservative Temperature difference between the abyssal and surface layer and $c^0_p$ is a constant close to the specific heat capacity at the sea surface of the present ocean
Substituting (7) into (21) gives

\[
\mathcal{H}(y, t) = \rho_0 c_p^0 \Delta \Theta \int_{t}^{t} \frac{h_{e}(t')}{c(y)} dt'
- \rho_0 c_p^0 \Delta \Theta \int_{0}^{t} \frac{t}{c(y)} \int_{t-t''}^{t} w_{E}(y, t') \frac{\tilde{c}(y)}{c(y)} dt' dt''
\]

(22)

For further discussion, in the limit of no wind forcing, the reader is referred to Zhai et al. (2011).

3. The reduced-gravity model experiment

We now examine the anomalies in the MOC and ocean heat content in response to recent NAO-related wind and thermohaline forcing in the idealized reduced-gravity sector model.

a. Numerical model description

The nonlinear, numerical reduced-gravity model used in this study is similar to that described in Johnson and Marshall (2002a) and Zhai et al. (2011). The model domain is an idealised sector ocean 40° wide and stretching from 45°S to 75°N, with vertical sidewalls and a resolution of 0.25°. The background model layer thickness is 750 m, with a reduced gravity of 0.02 m s⁻². No-slip and no-normal flow boundary conditions are applied. Sponges are applied at the northernmost and southernmost regions of the model domain to damp out any waves approaching these boundaries.
b. Experiment design

1) NAO wind forcing

In the first experiment, the reduced-gravity model is forced solely by the wind stress anomaly associated with the NAO. The wind stress anomaly is obtained in a similar way as in Eden and Jung (2001). First, the spatial pattern of the wind stress associated with the NAO is estimated by regressing the monthly NCEP reanalysis wind stress (Kalnay et al. 1996) on the NAO index (Hurrell 1995) for the period 1950-2010. The regressed wind stress is concentrated in the North Atlantic Ocean and reveals an anticyclonic wind stress curl situated at the gyre boundary that drives an “intergyre gyre” (not shown; Marshall et al. 2001). The regressed zonal wind stress in the Atlantic Ocean is then zonally-averaged (Fig. 2a) and multiplied by the monthly NAO index from 1950 to 2010 to drive the reduced-gravity model. In other words, the reduced-gravity model is forced by the zonal-mean “intergyre gyre” wind stress anomaly that pulses in time with the NAO index.

2) NAO thermohaline forcing

In the second experiment, the NAO thermohaline forcing is simply prescribed as a northward transport anomaly, $T_N$, at the northern boundary of the model, following Johnson and Marshall (2002a), representing the deep-water formation process at high latitudes (outside the model domain). The intensity of deep ocean convection at high latitudes in the North Atlantic and associated dense water renewals have been found to be closely linked to the phase of the NAO (e.g. Dickson et al. 1996; Curry et al. 1998; Marshall and Schott 1999). During positive NAO years, cold and dry air outbreaks from the nearby continent to the Labrador Sea in late winter tend to be more frequent and more severe, which can lead to enhanced convective activity and deep-water formation, while the opposite is true during negative NAO years. In the present process study, $T_N$ is, for simplicity, assumed to vary linearly with the NAO index for the period 1950-2010, with a standard deviation of $\sim 2$ Sv.
(Fig. 2b). However, it should be noted that convective activities and associated deep-water formation in the Greenland Sea are often observed to occur in anti phase with those in the Labrador Sea (e.g. Dickson et al. 1996; Marsh 2000).

3) NAO WIND AND THERMOHALINE FORCING

In the final experiment, the reduced-gravity model is forced by both NAO wind and thermohaline forcing in order to investigate the interplay between the two.

c. Results

1) NAO WIND FORCING

Fig. 3 shows the time evolution of the upper layer thickness anomaly during years of negative NAO (1956, 1962 and 1972) and positive NAO (1992, 1994 and 1996) in the experiment with only NAO wind forcing. The anomalous positive wind stress curl associated with the negative phase of the NAO before 1956 reduces the upper layer thickness between 35°N and 55°N, creating an anomalous cyclonic intergyre gyre circulation there. There is also indication of a weak anticyclonic gyre circulation to the south of the intergyre gyre owing to the relatively weak negative wind stress curl there before 1956. The negative layer thickness anomaly generated in the intergyre gyre area in the 1950s is shown to leak equatorward along the western boundary, eastward along the equator and then poleward along the eastern boundary, followed by the slow radiation of Rossby waves into the ocean interior. Over the following two decades of negative NAO, the leakage of negative anomalies from the intergyre gyre intensifies, and eventually fills the whole model domain by the year 1972. During the late 1980s and early 1990s when the NAO is switched to its strong positive phase, the anomalous negative wind stress curl deepens the upper layer and drives an anomalous anticyclonic intergyre gyre. The positive layer thickness anomaly, again, leaks equatorward and eventually spread over the whole model domain by the year 1996. Our results thus
suggest that the NAO wind stress forces a basin-wide upper layer thickness/ocean heat content change via a time-dependent “leaky” intergyre gyre circulation.

The zonally-averaged layer thickness at latitudes under direct influence of the intergyre wind forcing varies at both high and low frequencies, but is dominated by the large negative anomalies before the early 1970s and positive anomalies in the late 1980s and early 1990s (Fig. 4a). In contrast, in regions south of about 15°N where there is almost no direct wind forcing, only a multidecadal signal emerges. This multidecadal signal has a phase lag that varies with latitude, owing to the decrease in Rossby wave speed with increasing latitude (Zhai et al. 2011). There is a noticeable amount of layer thickness variability in the tropics (Figs. 3 and 4), suggesting that the low-frequency ocean variability generated by the intergyre gyre NAO wind forcing in the extra-tropics can potentially, via oceanic teleconnections, modulate the thermocline depth and sea surface temperature in the tropics where the atmosphere is particularly sensitive to small sea surface temperature anomalies. Fig. 4b shows the zonally-averaged layer thickness predicted from the theory using (16) and (22), which broadly agrees with that in the numerical model experiment.

The meridional transport anomaly, $T$, reveals a complex pattern with high-frequency variability in the North Atlantic, especially at latitudes under direct wind forcing, and low-frequency variability in the South Atlantic (Fig. 5a). Furthermore, $T$ appears somewhat disconnected at about 42°N. In order to understand the mechanism behind these features, $T$ is decomposed into its Ekman component, $T_{E_k}$, and its geostrophic component, $T_g$. The Ekman component, existing only in the North Atlantic, varies at high frequencies with a distinct discontinuity at 42°N (Fig. 4b). This pattern is consistent with the discussion in Section 2b: $T_{E_k}$ varies stochastically in response to stochastic NAO wind forcing, and vanishes at the latitude where $\tau_x = 0$, forming opposing meridional transport anomalies to the north and south of this latitude. The geostrophic component, taken as the residual, i.e., $T - T_{E_k}$, shows predominantly the multidecadal variability that spreads from the North Atlantic into the South Atlantic (Fig. 4c), and agrees well with that calculated using boundary
layer thickness anomalies through geostrophy \( T_g = g' H (h_e - h_w)/f \); Fig. 5d), providing justifications for the linear approximation used in deriving (14) and (16) in Section 2.

It is instructive to further investigate the boundary layer thickness anomalies that balance the geostrophic MOC anomalies. At latitudes under direct influence of stochastic wind forcing, \( h_w \), integrating the high-frequency forcing to the east and north, shows mostly multidecadal variability, while \( h_e \) has a more pronounced high-frequency component on top of the background multidecadal variability (Fig. 6). At latitudes to the south of the wind forcing, the ocean basin, including the lateral boundaries, feels only the multidecadal signal of \( h_w \). Note again that the strong tilting of \( h_w \) southward with latitude is associated with the slow Rossby wave propagation in the zonal direction, rather than the fast boundary wave adjustment in the meridional direction. It is clear from Fig. 6b that \( h_e \) is uniform along the eastern boundary.

Fig. 7a compares the eastern boundary layer thickness anomalies, \( h_e \), predicted using (16) from the linear theory and simulated in the numerical model experiment. The two curves broadly agree with each other, particularly well at low frequencies where \( h_e \) predicted from the theory captures the negative eastern boundary thickness anomalies before the late 1980s and positive anomalies thereafter. The theory seems to overestimate the magnitude of the interannual variability of \( h_e \), which could be due to numerical damping and the sponge layers that exist in the reduced-gravity model but not in the theory. Generally speaking, the theoretically-predicted \( h_b \), \( h_w \) and \( T_g \) from (9), (20) and (19) also agree reasonably well with those simulated in the numerical model experiment (Figs. 7b, c and d).

Note that the magnitude of the layer thickness anomalies on the western boundary (Fig. 7c) are roughly half those just outside the western boundary current (Fig. 7b). A significant reduction in variability is observed adjacent to the western boundary of the Atlantic, as discussed in Kanzow et al. (2009) and physically interpreted in terms of linear Rossby wave theory. Here, however, the reduction is less dramatic since the NAO-induced layer thickness anomalies occur on large scales, also consistent with the analysis of Cessi and Otheguy.
Finally we note that the magnitude of the layer thickness anomalies in the basin interior are smaller than those inferred from observations by Leadbetter et al. (2007). A possible explanation is that the present calculations are forced only the fraction of the Ekman upwelling anomaly associated with the NAO whereas Leadbetter et al. (2007) used the full wind stress anomaly.

2) NAO THERMOHALINE FORCING

The response of the reduced-gravity model to thermohaline forcing, $T_N$, at high latitudes has been studied previously (e.g. Johnson and Marshall 2002a, 2004; Deshayes and Frankignoul 2005; Zhai et al. 2011). Here we focus on ocean heat content and MOC variability when $T_N$ co-varies with the NAO index (Fig. 2b). Fig. 8 shows the time evolution of the upper layer thickness anomaly during the same years as in Fig. 3. There are two noticeable differences. First, the basin-wide layer thickness/heat content is anomalously positive during negative NAO years and anomalously negative during positive NAO years in the experiment with NAO thermohaline forcing, in the opposite sense to what happens in the experiment with NAO wind forcing. This difference arises due to the fact that during positive NAO years, stronger outflow, $T_N$, at the northern boundary acts to suck water out of the model, the influence of which gradually spreads to the ocean interior through boundary wave and Rossby wave adjustment processes. The same argument works in reverse for negative NAO years. Second, the stochastic variability excited along the western boundary with NAO thermohaline forcing gradually propagates into the interior of the ocean. As a result, the layer thickness anomaly at any given longitude and latitude is also largely stochastic, manifested by the positive and negative stripes of layer thickness anomalies in the interior of the model domain. In contrast, the layer thickness anomaly at any given longitude and latitude away from the direct influence of the intergyre wind forcing in the case with NAO wind forcing varies only on multidecadal time scales.
Consistent with Zhai et al. (2011), the high-frequency zonal-mean layer thickness/heat content variability in the experiment with NAO thermohaline forcing is confined to low latitudes while the low-frequency variability extends to mid- and high latitudes due to the “Rossby buffer” effect when the model is forced by a stochastic $T_N$ (Fig. 8a). The MOC anomaly in this experiment exhibits large variability throughout the northern hemisphere of the basin, but its high-frequency component is significantly reduced south of the equator (Fig. 8b), consistent with the prediction of Johnson and Marshall (2002a). The boundary layer thickness anomalies (Fig. 10) also show a very different behaviour from those in the experiment with NAO wind forcing: the high-frequency variability at the western boundary is greatly reduced upon arriving at the eastern boundary such that it is now easy to detect the multidecadal variability of $h_e$.

3) NAO WIND AND THERMOHALINE FORCING

The results from the experiment with both NAO wind and thermohaline forcing largely represent the linear combination of those from individual experiments with only NAO wind forcing and only NAO thermohaline forcing. Similar behaviour has also been reported from studies using OGCMs (e.g. Biastoch et al. 2008).

Fig. 11 shows time evolution of the upper layer thickness anomaly in the experiment with both NAO wind and thermohaline forcing. During negative NAO years, the reduced outflow at the northern boundary tends to deepen the upper layer of the ocean, which is counteracted by the positive wind stress curl anomaly that reduces the upper layer thickness and drives an anomalous cyclonic intergyre gyre. The thermohaline forcing appears to play a more important role in regions to the south of the intergyre gyre in the beginning of the negative NAO decade, while the wind forcing becomes more dominant towards the end of it. The same reasoning applies to the positive NAO decade. This phase difference between the wind-forced and thermohaline-forced low-frequency variability is readily seen in Figs. 4a and 9a, and can be explained largely by the time it takes the Rossby waves to cross the basin at

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mid-latitudes, integrating thickness anomalies induced by stochastic NAO wind forcing. In contrast, thickness anomalies induced by the NAO thermohaline forcing originate directly on the western boundary. One consequence of this phase difference is that the zonal-mean layer thickness/heat content anomaly appears to vary at somewhat higher-frequency south of the intergyre gyre (south of ~ 35°N) in the experiment with both NAO wind and thermohaline forcing than in the experiments with either NAO wind or thermohaline forcing alone (Fig. 12a). It is interesting to note that the low-frequency wind-forced and thermohaline-forced layer thickness anomalies nearly cancel each other out in the subtropics in 1962, leaving the ocean to the south of the intergyre gyre with only the high-frequency anomalies generated by stochastic thermohaline forcing (Fig. 11). With the addition of thermohaline forcing, the latitude at which the MOC variability appears disconnected is shifted northward from the experiment with wind forcing alone (Fig. 12b) because the thermohaline-forced MOC anomaly is in phase with the wind-driven Ekman transport anomaly south of 42°N, but out of phase north of 42°N.

Averaged over the whole Atlantic basin (excluding the sponges), the layer thickness and heat content anomalies in all three experiments exhibit pronounced multidecadal variability, with wind-forced anomalies largely opposing thermohaline-forced anomalies, especially at low frequencies (Fig. 13). While the basin-wide heat content anomaly generated by the wind forcing dominates in our experiments, the extent to which the wind- and thermohaline-forced anomalies compensate for each other depends on the parameters chosen.

4. Opposing changes of the MOC in the Atlantic

The meridional overturning circulation in the Atlantic Ocean plays an important role in our climate system through its transport of heat to high latitudes and transfer of atmospheric carbon dioxide to the deep ocean. The traditional view of the Atlantic MOC as a single coherent overturning cell has been called into question by recent modelling studies that have
produced gyre-specific decadal and multidecadal MOC changes (e.g. Bingham et al. 2007; Biastoch et al. 2008; Lozier et al. 2010). For example, Lozier et al. (2010) recently reported that the overturning circulation weakened by 1.5 ± 1 Sv in the subtropical gyre between 1950-1970 and 1980-2000, but strengthened by 0.8 ± 0.5 Sv in the subpolar gyre over the same period. Furthermore, changes in geostrophic transport associated with changes in the east-west density contrast appear to be responsible for these gyre-specific overturning changes.

Here we examine the changes of the geostrophic transport, $T_g$, between 1950-1970 and 1980-2000 based on the theory presented in Section 2, while taking into account of the realistic geometry of the Atlantic Ocean, i.e., $L_x = L_x(\phi)$ where $\phi$ is latitude (see Fig. 14). Fig. 15a shows $T_g$ diagnosed using (19) with the NAO wind forcing and then averaged over each 20 year period. The anomalous cyclonic intergyre wind stress curl during 1950-1970 is found to result in an overall increase in $T_g$ (blue curve), or equivalently a strengthening of the MOC, that extends from about 55°N all the way into the South Atlantic, whereas the anomalous anticyclonic wind stress curl during 1980-2000 results in an overall weakening of the MOC (red curve). There are opposing changes of $T_g$ north of 55°N, roughly where the anomalous wind stress curl changes sign, but the magnitude is much smaller in comparison. Therefore, with the NAO wind forcing alone, our analytical model predicts an overall reduction of the overturning circulation in the Atlantic Ocean by about 1 Sv from 1950-1970 to 1980-2000 (black curve).

Fig. 15b shows that the thermohaline forcing at high latitudes during 1950-1970 results in a southward transport anomaly over the whole Atlantic basin, albeit with amplitude decreasing slightly towards the south. A northward transport anomaly with a similar spatial pattern is found during 1980-2000. Therefore, with the NAO thermohaline forcing alone, the theory predicts an overall strengthening of the overturning circulation in the Atlantic Ocean by about 0.4 Sv from 1950-1970 to 1980-2000.

With both the NAO wind and thermohaline forcing, opposing changes of the meridional...
transport (or MOC) emerge, not unlike what has been reported by Lozier et al. (2010): weakening of the overturning in the subtropical gyre by about 0.7 Sv and strengthening of the overturning in the subpolar gyre by about 0.5 Sv from 1950-1970 to 1980-2000 (Fig. 15c). Note that the latitude where $T_g$ changes sign has now shifted southward to about 45°N. Our simple analytical model thus suggests that it is the interplay between the MOC induced by the NAO-related wind forcing and the MOC induced by the NAO-related thermohaline forcing that results in the opposing changes of the MOC in the subtropical and subpolar gyres. However, the magnitude of these gyre-specific overturning changes and the latitude at which these changes switch sign depend on the precise time periods chosen, and on our assumption about the magnitude of the thermohaline forcing associated with the NAO.

5. Conclusions

The response of the Atlantic to NAO-related wind and thermohaline forcing has been studied in the framework of a reduced-gravity ocean model. A quantitative theory has been developed, the results of which compare favorably with those obtained from numerical model calculations.

Our main conclusions are:

- NAO-related wind forcing is found to drive a time-dependent “leaky intergyre gyre” that integrates stochastic wind forcing and spreads low-frequency ocean heat content and MOC variability to the rest of the ocean basin, whereas NAO-related thermohaline forcing excites stochastic variability along the western boundary that is subsequently communicated into the ocean interior via boundary and Rossby waves.

- Basin-wide ocean heat content changes owing to NAO-related wind forcing and NAO-related thermohaline forcing are found to oppose each other, especially at low frequencies.
The interplay between the MOCs generated by NAO-related wind forcing and thermohaline forcing appears to cause the opposing changes of the MOC in the subtropical and subpolar gyres found between 1950-1970 and 1980-2000.

There are, of course, limitations with our simple reduced-gravity model approach. For example, no mean flow advection has been taken into account, and the model assumes vertical sidewalls and neglects all topographic influences. Nevertheless, the simple model used in this study does have the advantage of providing traceable analytical solutions to the problem at hand, and many of its qualitative features, we believe, are likely to carry over to the ocean.

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REFERENCES


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