Topographic Coupling of the Atlantic Overturning and Gyre Circulations

STEPHEN YEAGER
National Center for Atmospheric Research, Boulder, Colorado

(Manuscript received 29 May 2014, in final form 6 February 2015)

ABSTRACT

The vorticity dynamics associated with the mean and time-varying gyre and overturning circulations of the Atlantic Ocean are examined in a realistic ocean model hindcast simulation of the late twentieth century. Abyssal flow interaction with sloping bottom bathymetry gives rise to the bottom pressure torque (BPT) term of the vertically integrated vorticity equation. The dominance of this term in the closure of the barotropic gyre circulation noted in previous studies is corroborated here for both non-eddy-resolving and eddy-resolving versions of the Parallel Ocean Program (POP) model. This study shows that BPT is also a dominant term in the vorticity balance of the Atlantic meridional overturning circulation (AMOC) and therefore represents a key dynamical link between the overturning and gyre streamfunctions. The interannual variability of the Atlantic circulation over the last several decades, viewed in terms of time-varying integral vorticity balances, demonstrates the fundamental role played by BPT in mediating the large-scale barotropic and baroclinic flows. Forcing perturbation experiments show how flow–bathymetry interactions mediate buoyancy-driven changes in the gyre circulation and momentum-driven changes in the AMOC. Examples of topographic coupling of the overturning and gyre circulations that this analysis elucidates include the covariation of the high-latitude AMOC and subpolar gyre flows on decadal time scales, buoyancy-forced variance of the Gulf Stream, and large wind-driven variations in AMOC at subtropical latitudes.

1. Introduction

The Atlantic meridional overturning circulation (AMOC) plays an important role in the mean energy balance of Earth by transporting upward of 1 PW (1 PW = 10^{15} W) of heat northward between subtropical and subpolar latitudes (Johns et al. 2011; Msadek et al. 2013). Changes in the strength of AMOC have been linked to slow variations in North Atlantic sea surface temperature (SST) (e.g., Delworth and Mann 2000; Sutton and Hodson 2005; Knight et al. 2005; Danabasoglu et al. 2012b; Robson et al. 2012; Yeager et al. 2012) that drive far-ranging surface climate variations on multidecadal time scales [see, e.g., Delworth et al. (2007) for a review]. While the overturning component of the Atlantic circulation accounts for most of the oceanic poleward heat transport in the subtropics (Johns et al. 2011), the gyre component assumes a greater role at extratropical and subpolar latitudes (e.g., Böning and Herrmann 1994; Dong and Sutton 2002; Tiedje et al. 2012). There is evidence both from ocean general circulation models (OGCMs; e.g., Gerdes and Köberle 1995; Häkkinen 2001; Böning et al. 2006; Yeager et al. 2012; Yeager and Danabasoglu 2014) as well as from recent observations (Häkkinen and Rhines 2004; McCarthy et al. 2012) that changes in the large-scale Atlantic overturning and gyre circulations are fundamentally linked, but the dynamical connections remain obscure. Understanding the relationship between the overturning and gyre circulations, and their coordinated response to surface wind and buoyancy forcing perturbations, is of fundamental physical oceanographic interest, and it has broad implications for efforts to understand and predict the climate variations associated with changes in poleward oceanic heat transport.

The classic works that laid the foundations of our understanding of the large-scale ocean circulation emphasized wind and thermohaline forcing as the fundamental drivers of the barotropic gyre circulation (BGC; Sverdrup 1947) and AMOC (Stommel and Arons 1960; Stommel 1961). The idea that the gyre and overturning
circulations are associated with distinct mechanical and thermohaline surface forcings persists to some degree even to this day, although subsequent work has shed light on a more complex reality in which surface exchanges of both momentum and buoyancy can contribute significantly to both circulations, depending on location and time scale. For example, the observational studies of Luyten et al. (1985) and Wunsch and Roemmich (1985) called into question the validity of traditional “Sverdrupian” dynamics in the subpolar and subtropical Atlantic gyres, respectively. Both studies highlighted the importance of abyssal flow over sloping bottom topography in the vorticity balance of the BGC, and in Luyten et al. (1985) this deep flow was explicitly associated with a heat flux forcing. It is now generally understood that surface winds play an essential role in setting the strength of the mean overturning circulation by providing a key power source for the diapycnal mixing required to maintain abyssal stratification (Munk and Wunsch 1998). Furthermore, a variety of observational and modeling studies have made it clear that wind variability is the source of much of the high-frequency variance in AMOC, through adjustment to large-scale Ekman transport anomalies (Jayne and Marotzke 2001; McCarthy et al. 2012) or wind stress curl forcing (e.g., Kanzow et al. 2010; Duchez et al. 2014b).

The aim of the present study is to contribute to a deeper understanding of historical Atlantic Ocean circulation changes related to AMOC by exploring the topographic effects that give rise to coupling between the AMOC and the BGC in a realistic OGCM hindcast simulation of the late twentieth century. The initial focus will be on the time-mean dynamical balances of the large-scale barotropic and baroclinic flows; later sections will examine how these balances change in response to surface-forcing perturbations. Where and how does buoyancy forcing drive gyre circulation? Where and how does momentum forcing generate overturning motion? The answers relate to where and how these different surface forcings give rise to bottom flows that interact with spatially varying bottom topography \( [H = H(x, y) > 0] \). It is therefore fitting to briefly review some of the relevant classical ideas and results from the large body of work on the topic of rotating flow over topography.

The effects of topography on the large-scale gyre circulation are most easily conceptualized in the limit of a uniform density ocean. In such a framework, inviscid flow is constrained to follow geostrophic contours given by \( f/H \) (Pedlosky 1979), where \( f \) is the Coriolis parameter. For a purely wind-driven (frictionless) gyre in a uniform density ocean. In such a framework, inviscid flow is constrained to follow geostrophic contours given by \( f/H \) (Pedlosky 1979), where \( f \) is the Coriolis parameter. For a purely wind-driven (frictionless) gyre in a uniform density ocean. In such a framework, inviscid flow is constrained to follow geostrophic contours given by \( f/H \) (Pedlosky 1979), where \( f \) is the Coriolis parameter. For a purely wind-driven (frictionless) gyre in a uniform density ocean. In such a framework, inviscid flow is constrained to follow geostrophic contours given by \( f/H \) (Pedlosky 1979), where \( f \) is the Coriolis parameter. For a purely wind-driven (frictionless) gyre in a uniform density ocean. In such a framework, inviscid flow is constrained to follow geostrophic contours given by \( f/H \) (Pedlosky 1979), where \( f \) is the Coriolis parameter. For a purely wind-driven (frictionless) gyre in a uniform density ocean. In such a framework, inviscid flow is constrained to follow geostrophic contours given by \( f/H \) (Pedlosky 1979), where \( f \) is the Coriolis parameter. For a purely wind-driven (frictionless) gyre in a uniform density ocean. In such a framework, inviscid flow is constrained to follow geostrophic contours given by \( f/H \) (Pedlosky 1979), where \( f \) is the Coriolis parameter. For a purely wind-driven (frictionless) gyre in a uniform density ocean. In such a framework, inviscid flow is constrained to follow geostrophic contours given by \( f/H \) (Pedlosky 1979), where \( f \) is the Coriolis parameter. For a purely wind-driven (frictionless) gyre in a uniform density ocean. In such a framework, inviscid flow is constrained to follow geostrophic contours given by \( f/H \) (Pedlosky 1979), where \( f \) is the Coriolis parameter. For a purely wind-driven (frictionless) gyre in a uniform density ocean. In such a framework, inviscid flow is constrained to follow geostrophic contours given by \( f/H \) (Pedlosky 1979), where \( f \) is the Coriolis parameter. For a purely wind-driven (frictionless) gyre in a uniform density ocean. In such a framework, inviscid flow is constrained to follow geostrophic contours given by \( f/H \) (Pedlosky 1979), where \( f \) is the Coriolis parameter. For a purely wind-driven (frictionless) gyre in a uniform density ocean. In such a framework, inviscid flow is constrained to follow geostrophic contours given by \( f/H \) (Pedlosky 1979), where \( f \) is the Coriolis parameter. For a purely wind-driven (frictionless) gyre in a uniform density ocean. In such a framework, inviscid flow is constrained to follow geostrophic contours given by \( f/H \) (Pedlosky 1979), where \( f \) is the Coriolis parameter. For a purely wind-driven (frictionless) gyre in a uniform density ocean. In such a framework, inviscid flow is constrained to follow geostrophic contours given by \( f/H \) (Pedlosky 1979), where \( f \) is the Coriolis parameter. For a purely wind-driven (frictionless) gyre in a uniform density ocean. In such a framework, inviscid flow is constrained to follow geostrophic contours given by \( f/H \) (Pedlosky 1979), where \( f \) is the Coriolis parameter. For a purely wind-driven (frictionless) gyre in a uniform density ocean. In such a framework, inviscid flow is constrained to follow geostrophic contours given by \( f/H \) (Pedlosky 1979), where \( f \) is the Coriolis parameter. For a purely wind-driven (frictionless) gyre in a uniform density ocean. In such a framework, inviscid flow is constrained to follow geostrophic contours given by \( f/H \) (Pedlosky 1979), where \( f \) is the Coriolis parameter. For a purely wind-driven (frictionless) gyre in a uniform density ocean. In such a framework, inviscid flow is constrained to follow geostrophic contours given by \( f/H \) (Pedlosky 1979), where \( f \) is the Coriolis parameter. For a purely wind-driven (frictionless) gyre in a uniform density ocean. In such a framework, inviscid flow is constrained to follow geostrophic contours given by \( f/H \) (Pedlosky 1979), where \( f \) is the Coriolis parameter. For a purely wind-driven (frictionless) gyre in a uniform density ocean. In such a framework, inviscid flow is constrained to follow geostrophic contours given by \( f/H \) (Pedlosky 1979), where \( f \) is the Coriolis parameter. For a purely wind-driven (frictionless) gyre in a uniform density ocean. In such a framework, inviscid flow is constrained to follow geostrophic contours given by \( f/H \) (Pedlosky 1979), where \( f \) is the Coriolis parameter. For a purely wind-driven (frictionless) gyre in a uniform density ocean. In such a framework, inviscid flow is constrained to follow geostrophic contours given by \( f/H \) (Pedlosky 1979), where \( f \) is the Coriolis parameter. For a purely wind-driven (frictionless) gyre in a uniform density ocean. In such a framework, inviscid flow is constrained to follow geostrophic contours given by \( f/H \) (Pedlosky 1979), where \( f \) is the Coriolis parameter. For a purely wind-driven (frictionless) gyre in a uniform density ocean. In such a framework, inviscid flow is constrained to follow geostrophic contours given by \( f/H \) (Pedlosky 1979), where \( f \) is the Coriolis parameter. For a purely wind-driven (frictionless) gyre in a uniform density ocean. In such a framework, inviscid flow is constrained to follow geostrophic contours given by \( f/H \) (Pedlosky 1979), where \( f \) is the Coriolis parameter. For a purely wind-driven (frictionless) gyre in a uniform density ocean. In such a framework, inviscid flow is constrained to follow geostrophic contours given by \( f/H \) (Pedlosky 1979), where \( f \) is the Coriolis parameter. For a purely wind-driven (frictionless) gyre in a uniform density ocean. In such a framework, inviscid flow is constrained to follow geostrophic contours given by \( f/H \) (Pedlosky 1979), where \( f \) is the Coriolis parameter. For a purely wind-driven (frictionless) gyre in a uniform density ocean. In such a framework, inviscid flow is constrained to follow geostrophic contours given by \( f/H \) (Pedlosky 1979), where \( f \) is the Coriolis parameter.
waves, leading to strong coupling between surface- and bottom-intensified flows. They pointed out that JEBAR arises because topography causes the PV contours at difference depths (densities) to diverge, and so circulation that is essentially baroclinic (with PV-conserving flow in each layer) can acquire a barotropic component (which does not follow barotropic PV contours).

Many other recent studies have highlighted the important role of bottom topography in the dynamics of the large-scale Atlantic circulation by focusing on BPT effects. The works of Hughes (2000) and Hughes and de Cuevas (2001) argue that western boundary current (WBC) flow is inviscid and balanced by BPT based on theoretical considerations and numerical modeling evidence. Jackson et al. (2006) address the paradox of the role of friction, which is important in the balance of closed mean PV contours but which can nevertheless be negligible in the local BV balance of boundary current flow. Spence et al. (2012) show that, when model viscosity is reduced, North Atlantic Deep Water (NADW) flows through interior pathways with planetary vorticity advection balanced by BPT. Zhang and Vallis (2007) have shown that the BPT associated with deep western boundary current (DWBC) flow offshore of the Grand Banks is a key factor in setting the strength of the northern recirculation gyre (NRG)—the cyclonic, barotropic flow that has been observed between the northern flank of the GS and the Grand Banks (Hogg et al. 1986). They showed that BPT-related changes in the NRG can influence the GS path after separation from Cape Hatteras.

The relationship between JEBAR and BPT has been clarified by, among others, Mertz and Wright (1992), Greatbatch et al. (1991), and Bell (1999); the former arises in the (potential) vorticity equation of the vertically averaged horizontal flow, whereas the latter arises in the vorticity equation of the vertically integrated horizontal flow (see appendix A). JEBAR represents the component of BPT associated with the baroclinic (buoyancy-dependent) part of the pressure gradient, and therefore it vanishes in the absence of stratification (Mertz and Wright 1992; Salmon 1998). BPT can be nonzero regardless of stratification because it represents the projection of horizontal geostrophic bottom flow normal to isobaths. With the condition of no normal flow at the ocean bottom, BPT can be understood as a geostrophic bottom vortex stretching (refer to appendix A for notation):

$$
\frac{1}{\rho_o} \mathbf{J}(p_b,H) = \frac{1}{\rho_o} \left[ \frac{\partial p_b}{\partial x} \frac{\partial H}{\partial y} - \frac{\partial p_b}{\partial y} \frac{\partial H}{\partial x} \right]
= f \left[ v_g \frac{\partial H}{\partial y} + u_g \frac{\partial H}{\partial x} \right]
= f v_{gbg} \cdot \nabla H = -f w_{gb},
$$ (1)

where $g$ indicates geostrophic and $b$ indicates evaluation at the ocean bottom. In an ocean with variable bottom topography, inviscid, small Rossby number flow is characterized by the following barotropic vorticity relations based on (A13) and (A9):

$$\mathbf{U} \cdot \nabla \left( \frac{f}{H} \right) = \text{JEBAR} + \frac{1}{\rho_o} \mathbf{V} \times \left( \frac{\Delta \mathbf{r}}{H} \right)$$

$$\beta \mathbf{V} = \frac{1}{\rho_o} \mathbf{J}(p_b,H) + \frac{1}{\rho_o} \mathbf{V} \times \Delta \mathbf{r}. \quad (2)$$

In regions removed from direct wind forcing, the vertical stress terms are negligible. If we further postulate that the ocean is homogeneous, then JEBAR = 0, and the barotropic flow follows planetary PV contours ($q = f/H$), with BPT providing the necessary balance for any meridional advection across $f$ contours. In the more realistic baroclinic framework (with a nonzero JEBAR term), the barotropic flow has a component normal to planetary PV contours even where wind (and bottom drag) forcing is negligible, and BPT cannot be interpreted simply as the topographic steering associated with $f/H$-following flow. It follows from (2) that in the unforced interior

$$\frac{1}{\rho_o} \mathbf{J}(p_b,H) = f v_{gb} \cdot \nabla H = H(\text{JEBAR}) + f v_g \cdot \nabla H, \quad (3)$$

where the overbar indicates a vertical average. As discussed at length in Mertz and Wright (1992), the above relation shows that the geostrophic horizontal bottom velocity that appears in the expression for BPT is the sum of a barotropic (vertically averaged) and baroclinic component, with the latter directly related to the JEBAR term. It should be noted that (3) reflects the barotropic balance of unforced mean flow that does conserve layerwise PV (A11) with RHS equal to zero for isopycnal layers) but whose vertical sum deviates from $f/H$ contours according to the magnitude of JEBAR.

Vorticity features ubiquitously in studies of gyre circulation, but it is less common to analyze AMOC in the conceptual framework of vorticity dynamics. It is enlightening to do so since AMOC reflects the net basinwide flow across $f$ contours at each point in latitude–depth space. AMOC is clearly controlled by forcings that balance the advection of planetary vorticity. We will demonstrate how historical surface forcing perturbations (decomposed into momentum and buoyancy components) alter the mean vorticity balances of both the gyre (B1) and overturning (B3) streamfunctions. This approach sheds light on how (and where) topographic effects contribute to a coupled gyre/overturning response to what are essentially baroclinic (buoyancy) or barotropic (momentum)
surface forcings. The choice is made to focus on AMOC in depth space because it is the most common representation of the overturning. When AMOC is computed in density space rather than depth space, there is little distinction between overturning and gyre circulation at subpolar latitudes (Kwon and Frankignoul 2014), but the depth-space perspective is particularly useful for present purposes because the vertically integrated vorticity in 2 space \([A10]\) suggests a natural decomposition in terms of components familiar from gyre dynamics. Other studies have analyzed AMOC in terms of separate barotropic and baroclinic velocity fields, with the latter decomposed further into Ekman and (largely) geostrophic shear subcomponents (e.g., Lee and Marotzke 1998; Hirschi et al. 2003; Hirschi and Marotzke 2007; Cabanes et al. 2008).

We present a complementary perspective that clarifies the role of topography while avoiding the common (and questionable) assumption that surface Ekman transport is compensated by a uniform, barotropic return flow.

The primary tool employed in this study is the non-eddy-resolving (nominal 1° resolution) ocean component of the Community Earth System Model, version 1 (CESM1), which has been run in hindcast mode over the period 1948–2011. Some preliminary results from an eddy-resolving (nominal 0.1° resolution) version are also included for comparison. After describing the model experiments in section 2, we examine the climatological vorticity balances that characterize the Atlantic gyre and overturning streamfunctions in section 3. Interannual variations in the strengths of the BGC and AMOC are interpreted from a vorticity perspective in section 4, with a particular focus on how and where large buoyancy-forcing perturbations drove changes in gyre flow and large momentum-forcing perturbations drove changes in overturning in the recent past. A brief summary is given in section 5, and the two appendices contain derivations of the equations referenced throughout.

2. Methods

The model used is the coupled ocean–sea ice configuration of the CESM1 whose general description is given in Gent et al. (2011). The ocean model is the Parallel Ocean Program version 2 (POP2; Smith et al. 2010). We consider two different resolutions: a non-eddy-resolving version at nominal 1° horizontal resolution with 60 vertical levels and an eddy-resolving version at nominal 0.1° horizontal resolution with 62 vertical levels. The ocean model is coupled to the Los Alamos Sea Ice Model, version 4 (CICE4; Hunke and Lipscomb 2008), which runs on the same horizontal grid as the ocean. Detailed information about the POP2 and CICE4 models as implemented in CESM1 can be found in Danabasoglu et al. (2012a) and Holland et al. (2012), respectively, and the 0.1° simulation is an upgraded (and as yet undocumented) version of the model analyzed in Maltrud et al. (2010).

The focus will be on a set of forced historical simulations at 1° resolution that have been documented in Yeager and Danabasoglu (2014). The experiments are hindcasts of the 1948–2011 period with surface forcings based on the Co-ordinated Ocean–Ice Reference Experiments phase II (CORE-II) atmospheric datasets (Large and Yeager 2009; Danabasoglu et al. 2014). Yeager and Danabasoglu (2014) describe in detail the experimental setup of a “CONTROL” hindcast simulation of the late twentieth century, in which all surface forcing fields are interannually varying to the extent permitted by the CORE-II dataset, as well as perturbation experiments that isolate the oceanic interannual variability associated with surface momentum and buoyancy forcing variations (simulations M and B, respectively). The CONTROL simulation has been shown to compare favorably with North Atlantic observations (Yeager et al. 2012; Danabasoglu et al. 2014), and Yeager and Danabasoglu (2014) show that, to a very good approximation, large-scale, interannual Atlantic circulation anomalies from CONTROL can be recovered by summing the separate momentum- and buoyancy-forced anomalies (i.e., CONTROL’ = M’ + B’) is found to hold generally for both overturning- and gyre-related flow fields). The analytic vorticity components are related to finite-difference expressions following Bell (1999), with details of the computations given in Yeager (2013).

To address, at least in a preliminary way, the question of how robust the model mean BV balance is to whether or not the ocean mesoscale is resolved, the analysis of the aforementioned 1° CORE-II simulations is supplemented by analysis of a 15-yr-long 0.1° POP2 ocean–sea ice simulation [high resolution (HIRES)] forced with a repeating annual cycle of atmospheric surface fields [normal-year forcing (NYF); Large and Yeager 2004]. The different simulations considered are summarized in Table 1.

3. The mean vorticity balance of the large-scale Atlantic circulation

a. The barotropic gyre circulation

The climatological North Atlantic BV balance \([A9]\) from CONTROL (Fig. 1) clearly shows that BPT provides the first-order local balance for the strongest

---

2 The original 1948–2007 experiments have been extended through 2011 using climatological forcings for the radiation fluxes for 2010–11 and climatological precipitation for 2011. Climatology is used for those fields in those years because historical forcing data are unavailable.
meridional flows: the GS and North Atlantic Current (NAC), southward interior NADW transport [evident on either side of the Mid-Atlantic Ridge (MAR) at \(\approx 35^\circ W\)], the DWBC [evident at the western boundary (WB) south of \(\approx 30^\circ N\)], and subpolar gyre (SPG) currents. The fields in Fig. 1 have been spatially smoothed in order to focus attention on the large-scale balance. The nonlinear (Fig. 1e) and viscous torque (Fig. 1d) terms are nonnegligible in the vicinity of continental shelves, particularly along the western boundary of the subtropical gyre (STG) and around the perimeter of the subpolar gyre, but they show much more spatial variation than either \(\beta V\) or BPT. These terms may be important in explaining local flow features, but they do not contribute as systematically as BPT in the balance of strong barotropic flows. There are significant departures from a local Sverdrup balance \(\beta V = (1/\rho_o)\nabla \times \tau\), where BPT is large, most notably along the MAR north of \(\approx 30^\circ N\), the western boundary region, and throughout the SPG. The bottom drag (Fig. 1g) is only a weak term in the barotropic vorticity balance, with the largest impact along the east coast of Greenland. This balance of terms in the nominal 1° POP model corroborates similar findings in other OGCM simulations (Hughes and de Cuevas 2001; Thomas et al. 2014) and is in line with observational

<table>
<thead>
<tr>
<th>Experiment</th>
<th>Resolution</th>
<th>Forcing</th>
</tr>
</thead>
<tbody>
<tr>
<td>CONTROL</td>
<td>1°, 60L</td>
<td>CORE-II all forcings</td>
</tr>
<tr>
<td>M</td>
<td>1°, 60L</td>
<td>CORE-II momentum</td>
</tr>
<tr>
<td>B</td>
<td>1°, 60L</td>
<td>CORE-II buoyancy</td>
</tr>
<tr>
<td>HIRES</td>
<td>0.1°, 62L</td>
<td>NYF</td>
</tr>
</tbody>
</table>

Table 1. Guide to the experiments analyzed. CONTROL, M, and B are described in detail in Yeager and Danabasoglu (2014).

![Fig. 1. Terms in the climatological (1958–2007) vertically integrated barotropic vorticity balance \([A9]\) from the 1° CONTROL simulation \((10^{-9} \text{ m s}^{-2})\). The fields have been smoothed to remove features smaller than (roughly) \(2^\circ \times 2^\circ\). The breakdown of the (c) vertical stress term into (f) surface and (g) bottom stress components is indicated.](image)
studies of subtropical (Wunsch and Roemmich 1985; Roemmich and Wunsch 1985) and subpolar (Luyten et al. 1985) Atlantic dynamics.

To what extent does the BPT term of Fig. 1 simply reflect PV-conserving flow along $\beta H$ contours, as in a homogeneous ocean? Such a topographic Sverdrup balance would correspond to (2) with $\text{JEBAR} = 0$. However, the climatological vertically averaged BV balance [(A13)] from CONTROL (Fig. 2) shows a significant JEBAR term (Fig. 2b), far greater than the vertical stress term (Fig. 2c) over most of the domain, which dominates the nonzero advection of planetary PV (Fig. 2a). BPT is in general a small residual of large terms in (3), corresponding to almost compensating barotropic and baroclinic (i.e., JEBAR) components of the abyssal geostrophic horizontal velocity field, as shown in Holland (1972). Thermal wind tends to greatly reduce what would otherwise be very large (barotropic) velocities at the seafloor, particularly along the western boundary (cf. Figs. 1b and 2b). The balance of strong barotropic flow in Fig. 1 by the BPT associated with relatively weak abyssal flow (i.e., $\beta H \approx f v_{gb} \cdot V H$ even though $|v_{gb}| \ll |V|$ because of JEBAR) is explained by the fact that the “topographic beta” effect can become extremely large in regions characterized by steep bathymetric slopes:

$$f \frac{V H}{H} \gg \beta.$$  \hspace{1cm} (4)

The large JEBAR term in the (A13) balance implies that it is incorrect to interpret Fig. 1 as a topographic

---

**Fig. 2.** Terms in the climatological (1958–2007) vertically averaged barotropic vorticity balance [(A13)] from the 1° CONTROL simulation. Each term is multiplied by the ocean depth to transform into the units of vertically integrated vorticity ($10^{-9} \text{ m s}^{-2}$). Note the $5 \times$ increase in scale compared to Fig. 1. The fields have been smoothed to remove features smaller than (roughly) $2^\circ \times 2^\circ$. The breakdown of the (c) vertical stress term into (f) surface and (g) bottom stress components is indicated.
Sverdrup vorticity balance. Rather, it strongly suggests that JEBAR contributes significantly to the mean Atlantic BGC, as has been demonstrated by Greatbatch et al. (1991) by integrating (A13) along \( f/H \) contours. Large JEBAR suggests that there is a strong coupling of the baroclinic and barotropic modes in the model, that the barotropic flow is not simply steered by BPT to follow \( f/H \) contours, and that BPT is fundamentally linked to the baroclinic (buoyancy forced) velocity field.

The vertically integrated BV balance in the eddy-resolving POP2 simulation (HIRES; Fig. 3) is qualitatively similar to that in our CONTROL simulation, suggesting only a modest sensitivity to the model eddy field. The larger nonlinear term (Fig. 3e) is an expected result given the fact that, at this resolution, mesoscale eddy energy is quite realistic both in terms of magnitude and geographic distribution (Smith et al. 2000). This term is responsible for the considerable meandering of the GS after separation from the Atlantic coast, and it appears to contribute to an extension of the NAC into the “northwest corner” region off of the Grand Banks. Conversely, viscous torque (Fig. 3d) is much weaker in HIRES than in CONTROL, showing negligible influence except in the shelf currents around Greenland and in the Labrador Sea. As in CONTROL, BPT provides the main balance for the strong, barotropic WBC flows, much of the SPG flow, and southward NADW flow along interior pathways south of \( \approx 45^\circ \)N.

The mean barotropic gyre streamfunction \( \Psi_g \) in CONTROL is decomposed into vorticity components by zonally integrating the terms in the vertically integrated BV balance [(B1); Fig. 4]. For practical reasons, the integration in Fig. 4 has been performed along model grid lines that deviate from true latitude circles (albeit, only slightly south of about \( 50^\circ \)N), but the decomposition is robust insofar as the RHS terms do sum to give the \( \Psi_g \) field shown in Fig. 4a.3 The streamfunction perspective clarifies how the cumulative vorticity forcing of the ocean interior spins up the large-scale gyre circulation. The injection of negative vorticity by relatively weak, but uniform, wind stress curl (WSC) forcing over large swaths of ocean in the STG accounts for most of the mean anticyclonic gyre circulation south of Cape Hatteras \( (\approx 37^\circ \)N; Fig. 4c; note that by convention, \( \Psi_g \) is positive where the circulation is clockwise, reflecting negative mean relative vorticity within streamline contours). The strength of the anticyclonic STG is considerably enhanced by BPT forcing, particularly in the latitude band of \( \approx 33^\circ – 43^\circ \)N (Fig. 4b).

---

3 There is an unavoidable loss of mass continuity associated with the rotation from grid-oriented \( V \) to the true north-oriented \( V \) that appears in (A9). Hence, \( \Psi_g \) computed in this way does not quite close at the western boundary, and there are small local deviations from the climatological barotropic streamfunction computed by the model elliptic solver. Our conclusions are unaffected by the imperfect reproduction of the model \( \Psi_g \) from (B1).
positive $\Psi_g^{BPT}$ in this latitude range is associated with the southward, upslope flow of NADW on both the western and eastern flanks of the MAR (Figs. 1a,b). Similarly, the cyclonic circulation between $\approx 40^\circ$ and $55^\circ$N (comprising both the NRG, on the northern flank of the GS, and the central SPG) is almost entirely attributable to the BPT component (Figs. 4a,b). This cyclonic barotropic flow can be traced to strong bottom downwelling (positive BPT) that occurs near the MAR at around $45^\circ$N as well as on the flanks of the Rockall Plateau farther north (Fig. 1b). The mean WSC contributes very little to the strength of the cyclonic gyre circulation south of about $55^\circ$N, but negative $\Psi_g^c$ does dominate the cyclonic flow north of $55^\circ$N, in the Irminger and northern Labrador Seas (Fig. 4c). A strong conclusion is that the model’s climatological BGC is considerably enhanced by the interaction of abyssal flow with bathymetry over what it would be with just WSC forcing alone, in line with the mean JEBAR effect described in Greatbatch et al. (1991). BPT control of $\Psi_g^c$ is most pronounced in the latitude range of $\approx 33^\circ$–$55^\circ$N; it results in a strengthening of cyclonic circulation in the southern SPG and NRG regions and a strengthening of anticyclonic circulation in the northern STG.

The mass closure requirement of the gyre streamfunction [(B2)] implies a quasi-instantaneous relationship between interior and WB vorticity forcing that must hold at each latitude [we neglect here any net mass transport associated with Bering Strait throughflow that is of order 1 Sverdrup (Sv; 1 Sv = $10^6$ m$^3$ s$^{-1}$)]. Consistent with Fig. 1, the streamfunction decomposition in Fig. 4 shows that the WB closure of the largely wind-driven STG is accomplished in CONTROL primarily by the BPT term, which provides the main balance for GS flow along the North American upper continental shelf. To first order then, STG closure implies the following integral balance between WSC and BPT:

$$0 \approx \int_{x_i}^{x_f} \left[ \frac{1}{\rho_o} \mathbf{V} \times \mathbf{\tau} + \frac{1}{\rho_o} J(p_b, H) \right] dx. \quad (5)$$

This vorticity balance follows from the basinwide zonal momentum balance between wind stress and bottom form stress as discussed in Hughes (2000) and Hughes and de Cuevas (2001). This mass closure balance has important ramifications for understanding the circulation response to time-dependent surface forcing, as will be discussed below.

b. The meridional overturning circulation

The climatological AMOC in $\z$ space from CONTROL is characterized by a panhemispheric cell of 15–18 Sv that is centered around 1000-m depth, a stronger (>21 Sv)
recirculation between about 35° and 45°N and strong downwelling at middepths north of 40°N (Fig. 5a). South of about 35°N, the sign and spatial structure of the mean AMOC is largely set by the BPT term of the streamfunction decomposition [(B3)], with viscous torque $C_{\text{BPT}}$ and vortex stretching $C_{\text{fw}}$ generally tending to damp the BPT contribution to the total streamfunction. North of 30°N, AMOC is a small residual of large, counteracting stretching and BPT terms, with $C_{\text{BPT}}$ dominating south of $\approx 37$°N (Cape Hatteras) and $C_{\text{fw}}$ dominating farther north. The nonlinear term ($\Psi_{\text{o}}^A$; Fig. 5b) contributes very little to the mean AMOC balance except in the upper few hundred meters where WBCs are strongest, and the vertical shear stress term ($\Psi_{\text{o}}^t$; Fig. 5f) is negligibly small except in the surface Ekman layer. These surface-intensified vorticity terms are not visually apparent in the “bottom-up” integral shown in Fig. 5, but they do contribute to the (approximate) closure of the $\Psi_{\text{o}}$ streamfunction at the surface.4

South of about 35°N, the mean AMOC is largely associated with coastal boundary currents that flow northward in shallow layers and southward at depth, hugging the continental shelf of the Americas. Figure 5 shows that viscous torque does not balance the basinwide net meridional flow associated with these WBCs. There are, however, large viscous torques of opposite sign on the land and interior sides of the strong coastal jets that

4 A top-down computation of $\Psi_{\text{o}}$ [(B5)] generates a more complicated vorticity decomposition (not shown) because of the large surface terms, but it still highlights the importance of the BPT term in controlling the sign and structure of AMOC south of $\approx 37$°N.
layer thickness geostrophic, northward GS flow acquires an increase in zonal integral required for Fig. 5. A complementary that characterize the mean AMOC is obscured by the ultimate source of vorticity required to balance the overturning circulation over most of the Atlantic.\(^5\)

The mean AMOC decomposition is consistent with the classical picture of a highly geostrophic overturning. The vertically integrated vorticity from the ocean bottom to an interior z surface \([A10]\) can be decomposed into geostrophic and ageostrophic components with the geostrophic part given by

$$
\beta V_g(z) = \frac{1}{\rho_0} \int (p_b, H) + f w_g \left|_z \right. .
$$

The above equation is the vertical integral of the (planetary) geostrophic vorticity balance \((\beta v_g = f \alpha_g)\) with the bottom vortex stretching term given by BPT \([A11]\). The overturning streamfunctions associated with the terms above are plotted in Fig. 5 as \(\Psi_{a,e}, \Psi_{a,BPT}, \) and \(\Psi_{a,sw},\) with \(g\) indicating geostrophic. The geostrophic AMOC is almost indistinguishable from the total AMOC \(\Psi_o,\) showing that the net meridional flow can be understood in terms of net geostrophic stretching of planetary vorticity. The dominance of BPT south of 35°N can now be interpreted as an indication that the net, basinwide geostrophic stretching is largely set by bottom downwelling/upwelling \([i.e., BPT is the dominant term in (6)],\) while at higher latitudes the geostrophic stretching is largely set by vertical motion at middepths \([f w_g, C_g] \text{dominates in (6)]}.\) Of course, surface forcing (either of momentum or buoyancy) sets the ocean in motion, and BPT reflects the bottom boundary conditions on the resulting flow that arranges itself to materially conserve \(PV, f \alpha h,\) on large scales, where \(h\) represents the thickness of isopycnal layers \([A11]\). From this perspective, PV-conserving, geostrophic, northward GS flow acquires an increase in layer thickness \(h\) through downslope motion (positive BPT), while PV-conserving DWBC flow moves upslope (negative BPT).

The geographical distribution of the vorticity terms that characterize the mean AMOC is obscured by the zonal integral required for Fig. 5. A complementary perspective on AMOC steady-state dynamics can be gained by considering separate spatial maps of vertically integrated vorticity in the upper (northward) and lower (southward) branches of AMOC. The partial vertical integrals \([A10]\) include a middepth stretching term that vanishes in the barotropic integral. The resulting mean decompositions are shown in Figs. 6 and 7, where a depth of 1171 m is chosen as the interface between upper and lower branches based on the approximate depth of the \(\Psi_o\) maximum \((Fig. 5a).\) The AMOC decomposition discussed above can now be related to particular flow features in the shallow and deep layers. For example, the difference in the AMOC balance poleward and equatorward of about 35°N \((Fig. 5)\) is clearly associated with the transition from BPT-balanced GS flow along the continental shelf \((Fig. 6b)\) to NAC flow at higher latitudes that is balanced by intense geostrophic downwelling in the vicinity of 42°N, 35°W and extending to the northeast \((Figs. 6d,g).\) This region of strong downwelling, reflecting the model NAC \((which is misplaced too far to the east), accounts for much of the high-latitude closure of the mean AMOC \((Fig. 5).\) The vertical velocity here is largely diabatic, reflecting PV-conserving flow along sloping isopycnals \([A4];\) not shown. The AMOC from CONTROL in density coordinates \((see Fig. 4 of Danabasoglu et al. 2014)\) confirms that most of the diabatic transport in this model occurs much farther north, near 60°N. The negative contribution of \(\Psi_{BPT}^o\) to high-latitude AMOC \((Fig. 5d)\) is related to BPT-balanced Labrador and Grand Banks shelf transports that flow southward in the upper layer \((Figs. 6a,b).\)

The lower branch balance \((Fig. 7)\) highlights similar features with the sign reversed: BPT balances the DWBC along the deep continental shelf south of Cape Hatteras \((Fig. 7b),\) and geostrophic vortex compression under the NAC accounts for much of the southward interior flow between 37° and 50°N \((Fig. 7d).\) The middepth stretching term shows that the southward NADW flow is strongly coupled to the NAC in this region. We will show that the highly baroclinic flow in the NAC region is most sensitive to buoyancy forcing. BPT plays an important role over a much more expansive geographic region in the deep layer, where it provides the vorticity balance for interior flows in addition to shelf currents. In particular, the deep southward flow of NADW on the eastern flank of the MAR is almost entirely balanced by BPT, and large BPT signals are also seen on the western flank of the MAR and throughout the SPG \((Fig. 7).\) As already noted in the discussion of Fig. 5, the upper and lower branch transports are highly geostrophic \((Figs. 6h and 7),\) as given by sum of the BPT and geostrophic stretching terms \([Figs. 6b,g, 7b,g; (6)].\) PV-conserving DWBC flow is evident even in this vertical integral over multiple density layers. The large negative BPT \((strong bottom upwelling) along the

\(^5\) It is not clear how the AMOC vorticity balance would change under slip boundary conditions, but Yeager (2013) argues that BPT is dominated by the no normal flow condition.
continental shelf south of 37°N (Fig. 7b) is much larger than the middepth upwelling (Fig. 7d), implying upslope NADW flow that is being compressed into thinner layers to conserve $f/h$. Theoretical considerations imply an associated widening of the DWBC (Stommel and Arons 1972).

The mean dynamical balances discussed above can already offer insight into the time-dependent coupling between the overturning and gyre circulations. First, the dominance of surface buoyancy forcing in driving interannual SPG variance in CONTROL as shown in Yeager and Danabasoglu (2014) is evidently related to the fact that BPT acts as the primary vorticity balance for the mean BGC at high latitudes. Buoyancy forcing can drive gyre circulation in the presence of spatially varying bottom topography via the BPT term of the BV equation, and so the decadal covariation of high-latitude AMOC with SPG strength in CONTROL (Yeager et al. 2012) can be understood as a coupled overturning/gyre response to low-frequency, high-latitude density changes. We also showed that BPT provides a first-order vorticity balance for the mean gyre circulation as far south as 37°N, and so variations in the strength of (buoyancy driven) abyssal flow on the Grand Banks shelf and flanks of the MAR will tend to induce changes in the boundary between cyclonic SPG and anticyclonic STG circulation in that latitude band. It is to be expected therefore that BPT dynamics will greatly clarify the link between NADW
strength/path and GS latitude in this model, as suggested by Zhang and Vallis (2007). In the STG, the relationship between zonally integrated BPT and WSC [(5)] required for gyre closure will entail a strong covariation of these terms on interannual time scales. The AMOC vorticity equation [(B3)] yields the same mass closure balance [(B2)] when integrated to the ocean surface. We can begin to understand, therefore, how a wind-driven spinup of the anticyclonic STG might coincide with an AMOC spindown, in line with the recently observed “shift from overturning to gyre circulation” at 26.5°N (McCarthy et al. 2012). A stronger Sverdrup gyre associated with anomalously negative WSC (WSC\textsuperscript{−}, integrated over the basin) must be balanced by anomalously positive BPT (BPT\textsuperscript{+}), according to (5). To the extent that the BPT\textsuperscript{+} occurs in the abyssal ocean, below the depth of the AMOC maximum (1100 m), then (B3) suggests that flow–bathymetry interactions will contribute to a weakening of the AMOC.

4. Momentum- and buoyancy-forced interannual variability

In this section, we demonstrate the coupled response of the gyre and overturning circulations to interannual surface buoyancy and momentum forcing perturbations. Annual circulation anomalies in CONTROL can be decomposed as superpositions of anomalies from experiments M and B (Table 1). Figures 8 and 9 show how mechanical (M) and thermohaline (B) surface-forcing perturbations contribute to the variance of the overturning and gyre streamfunctions, respectively. The relative contributions depend on location and time scale with AMOC showing distinctly different temporal characteristics to the north and south of about 37°N, in line with Bingham et al. (2007). Experiment M dominates C\textsubscript{BPT} variance south of 37°N, where we have seen that C\textsubscript{BPT} dominates the mean (Fig. 5) as well as the variance (not shown) of AMOC. Experiment B dominates north of
37°N, where $\Psi_{fw}$ is paramount. The covariance of CONTROL with $M$ diminishes substantially when decadal time scales are isolated\(^6\) (Fig. 8e), while the covariance with $B$ is amplified on longer time scales (Fig. 8f). As discussed in Yeager and Danabasoglu (2014), the decadal variance explained by $M$ south of the equator is associated with low-frequency changes in the strength of Southern Ocean wind stress. The above results suggest that the pronounced change in the spectral character of AMOC near 40°N noted by Bingham et al. (2007) in a range of different ocean models relates to the latitudinal dependence of the efficacy of surface momentum and buoyancy forcing in driving AMOC.

The decomposition of $C_g$ variance (Fig. 9) leads to similar conclusions as for $C_o$: most of the gyre variance south of about 37°N is interannual and momentum driven, while it is primarily decadal and buoyancy driven at higher latitudes. The influence of buoyancy forcing is particularly strong in the intergyre–gyre (IGG) region that straddles the mean STG and SPG (Marshall et al. 2001b). The variance of $\Psi_g$ in the IGG region is often ascribed to WSC changes (e.g., Marshall et al. 2001b; Czaja and Marshall 2001), but our experiment B suggests that buoyancy forcing exerts a larger influence on $\Psi_g$ in the latitude band between ≈37° and 55°N (Figs. 9c,f).

\[a. \text{ Buoyancy-driven circulation}\]

A composite analysis of flow changes associated with strong/weak North Atlantic buoyancy forcing illustrates the mechanisms associated with decadal, buoyancy-driven modulations of the gyre circulation (Fig. 10). The index used for the composite is the high-latitude AMOC strength diagnosed from experiment B (Fig. 10a). This time series is characterized by large positive anomalies in the late 1980s and 1990s and very negative anomalies circa 1970 and 1980. Yeager and Danabasoglu (2014) show that the signal is largely associated with turbulent buoyancy forcing over the Labrador Sea region, and it explains most of variability in the high-latitude AMOC in CONTROL.

Stronger, buoyancy-driven overturning (Fig. 10f) is associated with stronger barotropic gyre circulation in B (Fig. 10b). Both the cyclonic SPG and the anticyclonic STG are enhanced, with the largest impacts seen in the Labrador and Irminger Seas, the IGG region, and the NRG. The vorticity balance required for these changes in barotropic circulation must come from changes in BPT because WSC does not vary in this experiment. The BPT change (Fig. 10d) resembles an enhancement of the

---

\(^6\)Here and throughout, decadal time scales are isolated using a 15-point Lanczos filter (Duchon 1979) with a cutoff period of 7 yr. This filter passes almost all of the variance at periods longer than 10 yr and half of the variance at a period of 7 yr.
mean field (Fig. 1b); as a result, the anomalous $\Psi_{\text{BPT}}^\text{c}$ is cyclonic in the SPG as it is in Fig. 4b. The enhanced NRG is associated with stronger bottom downwelling (BPT$^-$) on the western flank of the MAR and to the south of the Grand Banks, in line with Zhang and Vallis (2007). This is related to a westward shift in the branch of southward NADW that in the mean flows upslope on the western flank of the MAR (Figs. 7a,b) toward a position offshore of Grand Banks where it flows down-slope. Stronger anticycloonic circulation in the central STG is induced by stronger upslope flow (BPT$^+$) on the eastern flank of the MAR (Figs. 10b,d) and likewise along the Atlantic coast where the GS transport is enhanced by a few Sverdrups because of the strengthening of the (upslope) DWBC.

Many of the changes described above are also evident in the CONTROL composites (Figs. 10c,e,g), but the buoyancy-forced signal is somewhat obscured by the momentum-forced signal that is included in CONTROL. Decadal WSC changes enhance the buoyancy-forced strengthening of the SPG, and they dominate over buoyancy-forced gyre changes south of 40°N. Still, the BPT signal in CONTROL bears a close resemblance to the buoyancy-forced signal over most of the Atlantic, and a remnant of buoyancy-driven GS spinup associated with stronger upslope DWBC flow is evident in CONTROL. The composite difference of $\Psi_\epsilon$ shows that buoyancy-driven AMOC variations of subpolar origin propagate southward and have maximum impact on subtropical AMOC at depths between 1000 and 3000 m, below the depth of the climatological AMOC maximum (Fig. 10f). The similarity of the $\Psi_\epsilon$ composite to the covariance of B with CONTROL (Fig. 8f) indicates that the change from the 1970s to the 1990s captures a significant fraction of the buoyancy-driven AMOC variance over the 1958–2011 period. In CONTROL, there is also a large increase in AMOC strength south of the equator, which is momentum driven (in line with Fig. 8e), that Yeager and Danabasoglu (2014) attribute to the spinup of Southern Ocean winds over this time period.

b. Momentum-driven circulation

Changes in surface winds can give rise to abyssal flow variations that change the BPT term of the BV balance [(A9)], leading to a topographically modified gyre response to WSC forcing. Figure 11 shows the vorticity decomposition of $\Psi_\epsilon$ variance obtained in experiment M. Changes in the WSC-driven Sverdrup component $\Psi_t$ dominate on both interannual and decadal time scales. This implies that wind-driven decadal gyre variability in CONTROL (Fig. 9e) derives primarily from decadal variations in the WSC field itself [presumably, strongly linked to decadal North Atlantic Oscillation (NAO) variability]. The BPT component of $\Psi_\epsilon$ variance in M (Figs. 11c,f) reflects wind-driven adjustments in the abyssal flow field and its interaction with topographic gradients, which can reflect a highly nonlocal response to wind stress perturbations. This term results in significant departures from a purely Sverdrupian response, even in this purely wind-driven experiment. On interannual time scales
Fig. 10. Atlantic circulation change associated with extremes in buoyancy-driven AMOC.
(a) An index of high-latitude AMOC strength (averaged between 500- and 1500-m depth; 30°–50°N; see (f)) from the CONTROL and B experiments. Years of anomalously strong/weak buoyancy-driven AMOC are indicated with circles/squares (these correspond to >± one std dev in the AMOC index from experiment B). (b),(d),(f) The composite differences (strong – weak) in $\Psi_g$ (Sv), BPT ($10^{-9}$ m$^2$), and $\Psi_c$ (Sv) from experiment B. (c),(e),(g) As in (b),(d), and (f), but showing the composite differences from CONTROL (using the identical strong/weak years).
(Fig. 11c), the SPG response to WSC forcing is strongly damped by BPT effects, as are BGC fluctuations just north of Cape Hatteras. The gyre variance is amplified by BPT effects in the central STG, presumably because WSC forcing can induce deep flow over the MAR such that the BPT effect amplifies the vorticity forcing coming from the wind field. On decadal time scales, bathymetric slopes to the northeast and southeast of the Grand Banks stand out as important features that tend to augment the gyre response to low-frequency WSC changes.

As shown in Fig. 10 and in Yeager and Danabasoglu (2014), there is a high degree of meridional coherence of buoyancy-driven, decadal AMOC anomalies, with same-signed anomalies spanning the entire North Atlantic. In contrast, annual AMOC strength anomalies diagnosed from M tend to show coherence over $O(10^\circ-20^\circ)$ of latitude and are often of opposite sign to the north and south of 37°N (Yeager and Danabasoglu 2014) because of opposing Ekman effects (Bingham et al. 2007). The most important buoyancy forcing of the large-scale Atlantic circulation is geographically confined to the deep-water formation regions (e.g., the Labrador Sea) where large isopycnal layer thickness anomalies are generated that propagate southward. Thus, circulation anomalies in B, and the covariance of B with CONTROL, can be largely understood in terms of adjustment to the large (and nonlocal) turbulent buoyancy flux forcing perturbations that occur at high latitudes. In contrast, momentum forcing is important at all latitudes, and momentum-driven circulation variations (Figs. 8b,e, 9b,e) will in general be a complex convolution of local (i.e., same latitude) and nonlocal forcing perturbations (Polo et al. 2014).

We now examine the role played by bottom topography in translating wind stress variations into anomalous overturning motion, focusing on subtropical latitudes. This geographical focus is motivated by our finding that wind stress variability dominates interannual AMOC variance south of 37°N (Fig. 8) and by the fact that multiyear observations of AMOC and its components at 26.5°N (RAPID; Cunningham et al. 2007) exhibit substantial surface-forced (as opposed to internal ocean) variations (McCarthy et al. 2012; Roberts et al. 2013) that are not yet fully understood.

Large-scale circulation changes in experiment M associated with extremes in STG WSC forcing are shown in Fig. 12 using the composite technique. The regionally averaged WSC in the STG box (Fig. 12a) is negative in the time mean, corresponding to anticyclonic $\Psi_g^T$ (Fig. 4c). Interannual fluctuations in STG WSC do not exhibit a clear and consistent relation to winter NAO conditions ($r = 0.22$). The composite WSC signal over the whole North Atlantic associated with negative–positive STG WSC (Fig. 12b) resembles the WSC pattern associated with NAO$^-$ (e.g., Marshall et al. 2001a) at subpolar latitudes, but the composite shows a much stronger STG signal by construction. As expected, the STG WSC composite corresponds to a spinup of the anticyclonic subtropical gyre.
FIG. 12. Atlantic circulation change in experiment M associated with extremes in STG WSC forcing. (a) The time series of regionally averaged WSC [averaged over the box 22°–32°N, 70°–20°W; see (b)] from experiment M. Years of anomalously positive/negative WSC are indicated with circles/squares (these correspond to ± one std dev in the STG WSC). For reference, the winter [December–March (DJFM)] NAO index is also shown. The remaining panels show composite differences (negative–positive WSC) of (b) \( \frac{1}{r_0} \) \( \frac{3}{t_s} (WSC) \), (c) \( \frac{f}{r_0} \) \( \frac{3}{t_s} (Ekman pumping) \), (d) \( 2 \frac{b}{r_0} V_{\text{Ek}} \), (e) BPT, (f) \( \frac{f w_{\text{g}}(-1171 \text{m})}{r_0} \) (middepth geostrophic stretching), (g) \( b V_{g, \text{lower}} \) (vertically integrated geostrophic transport below 1171 m), (h) \( \psi_g \) (Sv), and (i) \( \psi_o \) (Sv). Panels (b)–(g) are in units of vertically integrated vorticity (10^{-9} \text{ ms}^{-1}), and Ekman transport (\( b V_{\text{Ek}} \)) is multiplied by a factor of 10 to show on that scale.
circulation (Fig. 12h). It also corresponds to a modest (0.4 Sv) decrease in AMOC at the latitude of the STG box (Fig. 12i) and a larger (2 Sv) decrease just to the north of the box (30°–40°N). A simple conceptual model of the relationship between barotropic gyre circulation and AMOC might suggest the opposite relationship: if the enhanced barotropic southward flow in the STG is returned as enhanced northward WBC flow at shallow depths, then an increase in \( \Psi_g \) should correspond to an increase in \( \Psi_o \).

The vorticity terms help to shed light on the anti-correlation of STG \( \Psi_g \) and \( \Psi_o \) seen in the model and in RAPID observations (McCarthy et al. 2012). The WSC composite can be decomposed into terms associated with ageostrophic surface Ekman transport and Ekman pumping of the geostrophic interior (see Pedlosky 1996):

\[
\frac{1}{\rho_o} \nabla \times \tau_s = -\frac{\beta}{\rho_o f} \tau_s + \frac{f}{\rho_o} \nabla \times (\tau_s/f)
\]

\[
\beta V_S = \beta V_{ek} + \beta V_g,
\]

where \( \beta V_S \) is the vertically integrated Sverdrup transport, \( \beta V_{ek} \) is the Ekman transport associated with zonal wind stress at the surface, and \( \beta V_g \) is the geostrophic transport associated with convergent Ekman transport (all expressed in terms of vertically integrated vorticity). Anomalous Ekman pumping of fresh Sverdrup transport is by far the largest component of the anomalous southward Sverdrup transport associated with negative STG WSC (Fig. 12c); the anomalous Ekman transport is an order of magnitude smaller (Fig. 12d). We note that the zonal wind stress anomaly implied by Fig. 12d is again reminiscent of a NAO- pattern (Visbeck et al. 2003; compare to Fig. 14d). Years in which the strength of the wind-driven \( \Psi_g \) is maximum in the STG box correspond to strongly negative \( \beta V_{ek} \) anomalies farther north between 30° and 40°N. Thus, anomalous Ekman transport would appear to explain the large negative \( \Psi_o \) anomalies seen in the AMOC composite over that latitude range (Fig. 12i).

The \( \Psi_o \) anomaly is broken down into lower branch components in Figs. 12e–g, which show the composites of terms in the vertically integrated geostrophic transport between the ocean bottom and \( z = 1171 \text{ m} \) [(6)]. These terms show how the vorticity balance of the lower branch of AMOC adjusts in response to enhanced STG WSC. Note that a positive AMOC anomaly corresponds to a negative \( \beta V \) anomaly (enhanced southward transport) in the deep layer. As anticipated from (5), the enhanced southward gyre transport (WSC') necessitates an enhanced closure term (BPT'), as is apparent all along the Atlantic shelf in Fig. 12c. The change in BPT is nearly synchronous with the change in WSC, with the latter communicated to the western boundary via fast barotropic Rossby waves. Importantly, much of the BPT' occurs at depths below 1100 m, and this leads to a reduction in DWBC flow along the deep continental shelf (Fig. 12g). Although ultimately caused by surface WSC forcing over a broad region, the weakening of the DWBC off the coast of Florida is locally “forced” by the interaction of barotropic waves with deep shelf bathymetry that results in anomalously negative bottom vortex stretching (BPT').

Anomalous geostrophic stretching at middepths (Fig. 12i) is clearly related to the Ekman pumping at the surface (Fig. 12c). The difference between these fields gives the upper-layer geostrophic transport anomaly (i.e., the net geostrophic stretching in the upper layer; see Fig. 6g), for interior regions. We interpret this term (Fig. 12f) as a synchronous snapshot of the westward-propagating, downwelling baroclinic Rossby wave response to WSC-. It is associated with a positive meridional flow anomaly in the upper layer (near 26°N, west of the MAR) and a negative meridional flow anomaly in the deep layer. Its contribution to \( \beta V_{g,\text{lower}} \) is to enhance southward flow to the west of the MAR, tending to enhance AMOC. Thus, the BPT' signal and the \( f \omega \) signal tend to cancel, so that the net AMOC reduction at STG latitudes is weaker than it would be with just the DWBC reduction (BPT') alone. The efficacy of \( \beta V_{ek} \) in driving the large AMOC anomaly farther north (30°–40°N) is related to the fact that large BPT' is still generated on the western boundary for closure, but there is little offsetting baroclinic wave response to the large-scale Ekman transport anomaly.

The circulation response in M to extreme STG Ekman transport anomalies is shown in Fig. 13. There is only a weak spinup of the anticyclonic STG circulation (Fig. 13h) when \( \beta V_{ek} \) is minimized, associated with a region of negative WSC in the western STG (Fig. 13b). The correlation of the STG Ekman transport index with the winter NAO index is fairly high (\( r = 0.53 \)), and we note that the strong NAO- of 2010 coincided with a very negative \( \beta V_{ek} \) anomaly in the STG. The large, full-depth negative AMOC response seen at 30°–40°N in Fig. 12i is now shifted farther south to the latitude range of the STG box (22°–32°N), yielding a >1-Sv reduction in AMOC near 26°N. As seen near 35°N in the WSC composite, the AMOC response to anomalous STG Ekman transport is not uniform in depth but exhibits particularly strong reductions in the upper branch (above 1000 m) and in the abyss (below 2000 m). The slowdown of the lower branch of AMOC (Fig. 13g) near 26°N is again associated...
primarily with BPT$^+$ (Fig. 13e), but now there is a noticeable BPT$^+$ signal on the eastern flank of the MAR (near 40°W) that was not seen in Fig. 12e. This is consistent with the idea that Ekman transport anomalies project strongly onto the barotropic mode (e.g., Jayne and Marotzke 2001; Lee and Marotzke 1998; Cabanes et al. 2008), thereby promoting wave interactions with very deep bathymetric features in addition to shallower continental
shelf bathymetry. This is perhaps the reason why the AMOC response to Ekman transport anomalies exhibits such nonuniformity with depth, with the compensating return flow also nonuniform in longitude.

Finally, the combined gyre/overturning wind-driven response to extreme winter NAO conditions is shown in Fig. 14. The response in the STG is a mixture of the previous two composites, with a modest spinup of the anticyclonic gyre and weakening of the AMOC associated with NAO$^-$ conditions. This results from the fact that extreme NAO$^-$ (NAO$^+$) tends to correspond to moderately negative (positive) WSC and $\beta V_{ek}$ anomalies.
at STG latitudes. As already mentioned, however, the relationship between NAO and STG WSC is quite weak. North of 40°N, positive WSC (with a strong, basinwide positive Ekman transport component) induces intense BPT along the Labrador and Grand Banks shelf for closure. The result is enhanced southward flow in the deep layer, producing an AMOC component to NAO− wind forcing at subpolar latitudes.

Understanding how the lower branch of AMOC, and in particular the DWBC, responds to wind stress perturbations is critical for interpreting recent observed changes at 26°N, and it complements recent studies that have focused on the wind-driven response of the upper branch at this latitude (Duchez et al. 2014a,b). McCarthy et al. (2012) show a pronounced weakening of lower NADW transport associated with two extreme, negative Ekman transport anomalies that occurred in early and late 2010 (the double event appears as a single, negative, annual-mean anomaly, computed between successive 1 Januarys, in our Fig. 13). Although there is not a consistent relationship between winter NAO conditions and STG Ekman transport, both reached extreme negative values in 2010. The observed impact on lower NADW, but not upper NADW, may be attributable to the depth of the Ekman-induced BPT signal (Fig. 13e). Although not observed, Fig. 13g suggests that a considerable fraction of the reduced southward transport in the lower layer of AMOC in 2010 may have been associated with anomalous deep flow on the eastern flank of MAR (positive \( \beta V_g \), lower near 26°N, 40°W). We can also speculate that the observed slow intensification of upper midocean (gyre) transport and the weakening of lower NADW that began as early as 2006 (McCarthy et al. 2012) was related to the advent of anomalously negative STG WSC in 2006 after several years of anomalously positive WSC in this region (Fig. 12a). The resulting gyre spinup would tend to induce enhanced BPT− and a weakened DWBC on the deep continental shelf (Fig. 12g).

5. Conclusions

The large-scale dynamics of the Atlantic Ocean cannot be properly understood without consideration of deep flow interaction with bottom topography because the abyssal ocean is not motionless. Bottom pressure torque is a leading-order term in the vorticity balance of the strong meridional flows that feature prominently in the mean overturning and gyre circulations in a widely used OGCM, even at eddy-resolving resolution. BPT plays an important role in driving the barotropic circulation insofar as it represents a proximate vorticity forcing of the overlying water column. In a realistic hindcast simulation of the late twentieth century, the BPT field is not simply an artifact of topographic Sverdrup balance, but rather shows a strong dependence on the baroclinicity of the flow. It is key to understanding the western boundary closure of the gyre circulation, despite the fact that the model is parameterized with substantially enhanced viscosity along western boundaries in order to resolve a viscous Munk boundary layer (Large et al. 2001; Jochum et al. 2008). These results build on and corroborate the findings of many earlier studies (e.g., Greatbatch et al. 1991; Hughes 2000; Hughes and de Cuevas 2001; Zhang and Vallis 2007; Spence et al. 2012) but go further to clarify the topographic control of the AMOC and its time-dependent coupling with the horizontal gyre circulation.

AMOC in the CESM model is characterized by distinct dynamical regimes to the north/south of about 37°N, which is the approximate latitude where strong, interior, baroclinic meridional flows (the NAC and NADW flow over the MAR) transition to strong, barotropic boundary current flows (the GS and DWBC). The two regimes are characterized by different mean vorticity balances, dominated by interior stretching (to the north) and BPT (to the south). In both regimes, the AMOC can be well understood in terms of net geostrophic stretching of planetary vorticity, but they differ in terms of the baroclinic/barotropic character of the stretching. Strong buoyancy forcing at high latitudes maintains zonal density gradients that allow for baroclinic overturning balanced by intense geostrophic downwelling at middepths. The upper branch is composed of a detached, northward NAC that is strongly coupled to the southward lower branch via a large (mostly adiabatic) \( f \omega \) term near 1000-m depth. In the WBC regime of AMOC, large viscous torques play a key role in setting the interior AMOC into motion by generating geostrophic stretching in regions adjacent to the boundary currents, but large torques of opposite sign prevail on the land side of the boundary currents. Thus, in the net AMOC vorticity balance south of 37°N, meridional flow is balanced by pressure forces from the solid Earth. Whether or not there is a net effect of AMOC variations on the vorticity of the planet, implying measurable variations in the length of day (LOD) is an interesting and open question. Tantalizing hints have emerged of a correlation between LOD and intrinsic multidecadal variations of global air temperature (Dickey et al. 2011), but a clear link to ocean dynamical changes has not been established.

Flow interaction with bathymetry plays a significant role in the transient circulation response to changes in surface momentum and buoyancy forcing. Large buoyancy-forced changes in the gyre circulation are accomplished through the BPT term of the barotropic vorticity equation. This
explains the covariation of AMOC and cyclonic SPG flow on decadal time scales (such as the spinup of AMOC and subpolar gyre circulation between the 1970s and 1990s), and it also appears to be the key mechanism associated with variability of the NRG and IGG. Strong, buoyancy-forced AMOC is associated with greatly enhanced cyclonic flow in the SPG and NRG, but also with enhanced anticyclonic flow in the STG because of stronger NADW flowing upslope along the MAR. High-latitude buoyancy forcing would appear to influence GS flow through NADW flow interaction with bathymetry, but this signal is small compared to wind-forced changes at subtropical latitudes. BPT also significantly modulates the gyre response to pure Sverdrup forcing, generally tending to damp the Sverdrup response to WSC anomalies (as it must, for gyre closure) but amplifying the Sverdrup response in certain regions (such as the central STG).

Buoyancy forcing changes AMOC by altering the baroclinicity of the flow on decadal time scales, but momentum forcing projects more strongly onto the barotropic mode of AMOC. By decomposing both the overturning and gyre circulations in terms of vorticity, the role of BPT as a coupling term becomes evident. BPT provides the primary closure for the wind-driven gyre, and therefore variations in the basinwide wind-driven meridional transport (either geostrophic, via Ekman pumping, or ageostrophic via surface Ekman transport) give rise to corresponding BPT anomalies at each latitude. BPT anomalies can, in turn, influence AMOC by changing the depth distribution of barotropic return flow. We have explored how interannual wind variations generate overturning anomalies at subtropical latitudes by showing how extreme WSC and Ekman transport events impact the deep southward branch of AMOC. Our annual-mean analysis focusing on the fast response to (local) forcing suggests that, in several respects, recently observed interannual AMOC variability could be interpreted as a topographically coupled gyre/overturning response to wind stress changes, but many details remain to be explored.

Acknowledgments. This work was supported by the NOAA Climate Program Office under Climate Variability and Predictability Program Grants NA09OAR4310163 and NA13OAR4310138 by the National Science Foundation (NSF) Collaborative Research EaSM2 Grant OCE-1243015. I gratefully acknowledge the work and dedication of the team of scientists and engineers who have developed the CESM model. This work could not have been accomplished without the computing resources provided by NCAR’s Computational and Information Systems Laboratory. Thanks to Peter Gent and Baylor Fox-Kemper for helpful comments on an early draft of the paper. I’m grateful for the in-depth critiques provided by Peter Rhines and another anonymous reviewer that helped to improve the paper.

APPENDIX A

Integrated Vorticity Relations

We are interested in the large-scale, integral properties of the primitive equations for a thin fluid on a rotating sphere:

$$\begin{align*}
\partial_t \mathbf{u} + f \mathbf{k} \times \mathbf{u} &= -\frac{1}{\rho_o} \nabla p + \frac{1}{\rho_o} \partial_z \mathbf{r} + \mathbf{a} + \mathbf{b} \\
\partial_t \mathbf{u}_g + f \mathbf{k} \times \mathbf{u}_g &= -\frac{1}{\rho_o} \nabla p \\
\partial_t \mathbf{u}_\text{ag} + f \mathbf{k} \times \mathbf{u}_\text{ag} &= \frac{1}{\rho_o} \partial_z \mathbf{r} + \mathbf{a} + \mathbf{b} \\
\partial_t p &= -g \rho \\
\mathbf{v} \cdot \nabla \mathbf{v} &= 0,
\end{align*}$$

where $\mathbf{v} = (u, v, w) = (\mathbf{u}, w)$ is the full three-dimensional velocity field; $\mathbf{u} = \mathbf{u}_g + \mathbf{u}_\text{ag}$ represents the horizontal velocity split into geostrophic and ageostrophic components; $f \mathbf{k} = (0, 0, 2\Omega \sin \phi)$ is the projection of planetary vorticity in the local vertical direction as a function of latitude $\phi$; $\mathbf{a}$ represents the nonlinear advective momentum tendency; $\mathbf{b}$ represents the horizontal viscous forcing; $\partial_z \mathbf{r}$ is the viscous forcing associated with the vertical shear stress $\mathbf{r}$; $p$ is pressure; $g$ is the gravitational acceleration; $\rho$ is the ocean density; and $\rho_o$ is the background ocean density.

In steady state, the vertical integral over a layer of thickness $h = h_1 - h_2$ can be written (neglecting for now the nonlinear and horizontal viscosity terms):

$$\rho_o f \mathbf{k} \times \mathbf{U} = -\mathbf{V} \int_{h_2}^{h_1} p \, dz + p_1 \mathbf{V} h_1 - p_2 \mathbf{V} h_2 + \tau_1 - \tau_2,$$

where $\mathbf{U} = \int_{h_2}^{h_1} \mathbf{u} \, dz$ represents the net horizontal transport within the layer and $\tau_1 = \tau_1_{z=h_1}$ and so on. The vertical integral of the continuity relation gives

$$\mathbf{V} \cdot \mathbf{U} - \mathbf{u}_1 \cdot \mathbf{V} h_1 + \mathbf{u}_2 \cdot \mathbf{V} h_2 = -(w_1 - w_2).$$

If the surfaces defined by $z = h_1(x, y, t) = 0$ and $z = h_2(x, y, t) = 0$ represent material surfaces [i.e., if $D_i(z = h_1) = 0$ and similarly for $h_2$], then the kinematic boundary conditions for no normal flow apply:

$$w_1 = \partial_z h_1 + \mathbf{u}_1 \cdot \mathbf{V} h_1.$$

These conditions apply for both the ocean free surface [$z = \eta(x, y, t)$] and the bottom [$z = -H(x, y)$] and to
very good approximation for surfaces of constant potential density (isopycnals). The vertically integrated continuity equation then relates horizontal divergence and thickness tendency:
\[
\nabla \cdot \mathbf{U} = \nabla \cdot \mathbf{U}_g + \nabla \cdot \mathbf{U}_{ag} = -\partial h,
\]
where the breakdown into geostrophic and ageostrophic divergence follows from the linearity of the continuity equation. If the bottom surface is the ocean bottom and the top layer is a surface of constant \( z \), then the vertically integrated continuity equation \( \int_{-H}^z \, dz \) becomes
\[
\nabla \cdot \mathbf{U} = -\partial h|_z.
\]
(A6)

The bulk vorticity of the layer can be expressed either as a vertically integrated vorticity or as a vertically averaged vorticity. The former is obtained by applying the vertical curl operator \( \mathbf{k} \cdot \nabla \times (\cdot) \) to (A2):
\[
\rho_o \mathbf{U} \cdot \nabla f + \rho_o f \nabla \cdot \mathbf{U} = J(p_1, h_1) - J(p_2, h_2) + \nabla \times \Delta \tau,
\]
where \( J(a, b) = \partial_z a \partial \eta b - \partial_z a \partial \eta b = \nabla a \times \nabla b \) and \( \Delta \tau = \tau_1 - \tau_2 \). The interfacial pressure gradient terms on the RHS can be rewritten in terms of horizontal geostrophic flow, for example, \( J(p_1, h_1) = f \rho_o \mathbf{u}_g \cdot \nabla h_1 \). If \( h_1 \) is a material surface, then it follows from (A4) that \( J(p_1, h_1) = f \rho_o w_\eta \) (where \( w_\eta \) is the geostrophic vertical velocity at surface \( h_1 \)) to ensure that \( \mathbf{v}_g \) be tangential to the material surface in steady state.

The steady-state barotropic vorticity balance is obtained when \( h_1 \) corresponds to the sea surface \( z = \eta(x, y, t) \) and \( h_2 \) corresponds to the bottom depth \( z = -H(x, y) \):
\[
\beta V = \frac{1}{\rho_o} J(p_b, H) + \frac{1}{\rho_o} \nabla \times \Delta \tau,
\]
(A8)

where we have used \( \nabla \cdot \mathbf{U} = 0 \), \( \nabla \cdot \mathbf{V} = \beta V \), \( p_b \) is the bottom pressure at \( z = -H(x, y) \), and we have neglected variations in pressure at the sea surface (\( \nabla p_b = 0 \)). The first term on the RHS is the BPT term of the vertically integrated vorticity equation. The nonlinear and horizontal viscous forcing terms from (A1) can be added to the balance as follows:
\[
\beta V = \frac{1}{\rho_o} J(p_b, H) + \frac{1}{\rho_o} \nabla \times \Delta \tau + \nabla \times \mathbf{A} + \nabla \times \mathbf{B},
\]
(A9)

where \( \mathbf{A} = \int_{-H}^z a \, dz \) and \( \mathbf{B} = \int_{-H}^z b \, dz \). It is straightforward to show that this barotropic vorticity equation can alternatively be derived by first taking the curl of the momentum equation and then integrating over depth (e.g., Hughes and de Cuevas 2001). We also note here that a partial vertical integral from the ocean bottom to a given \( z \) level \( (\int_{-H}^z \, dz) \) yields
\[
\beta V(z) = \frac{1}{\rho_o} J(p_b, H) + \frac{1}{\rho_o} \nabla \times \Delta \tau + \nabla \times \mathbf{A} + \nabla \times \mathbf{B} + f w|_z,
\]
(A10)

where we have used \( \nabla h = 0 \) and (A6), and the limited vertical extent of the integrals is implied for the \( \tau, \mathbf{A}, \) and \( \mathbf{B} \) terms on the RHS.

The (vertical) vorticity of the vertically averaged momentum equation results in an expression for material conservation of planetary geostrophic (PG) potential vorticity: \( q = f/h \). Taking the curl of (A2) after normalizing by the layer thickness \( h \) gives
\[
\rho_o \mathbf{U} \cdot \nabla \left( \frac{f}{h} \right) = \nabla \times \left( -\frac{1}{h} \int_{h_2}^{h_1} \nabla p \, dz \right) + \nabla \times \left( \frac{\Delta \tau}{h} \right),
\]
(A11)

where we have used (A5) corresponding to material surfaces \( h_1 \) and \( h_2 \). The first term on the RHS vanishes when the pressure gradient is constant within the layer, which corresponds to a layer of constant density and uniform horizontal geostrophic velocity through (A1). In that case, the horizontal flow in the layer follows contours in the absence of forcing \( \Delta \tau = 0 \). In general, however, the pressure gradient term in (A11) can be nonzero in a layer of nonconstant density, and so the bulk planetary PV will not be materially conserved. If the vertical integral is taken over the full ocean depth, then pressure gradient term becomes the JEBAR term discussed by numerous authors (Mertz and Wright 1992; Greatbatch et al. 1991; Salmon 1998; Bell 1999):
\[
\nabla \times \left( -\frac{1}{H} \int_{-H}^{0} \nabla p \, dz \right) = J(\gamma, H^{-1}) = \text{JEBAR},
\]
(A12)

where \( \gamma = g \int_{-H}^{0} \rho z \, dz \) represents the potential energy of the water column. The steady-state barotropic PV equation becomes (adding in nonlinear and horizontal viscosity terms)
\[
\mathbf{U} \cdot \nabla \left( \frac{f}{H} \right) = \text{JEBAR} + \frac{1}{\rho_o} \nabla \times \left( \frac{\Delta \tau}{H} \right)
+ \nabla \times \left( \frac{\mathbf{A}}{H} \right) + \nabla \times \left( \frac{\mathbf{B}}{H} \right).
\]
(A13)

APPENDIX B

**Gyre and Overturning Streamfunctions**

Equation (A9) relates the local depth-integrated meridional flow across planetary vorticity contours to the barotropic forcing associated with (in order on the
right-hand side) BPT, surface wind stress curl (minus the torque associated with any bottom drag), nonlinear effects, and lateral friction. The nondimensionalization of the barotropic flow [(A5)] means that (A9) can be recast as the well-known streamfunction equation of the barotropic gyre circulation, traditionally integrated from the eastern boundary:

$$\psi(x) = \frac{1}{\rho_o} \int_{x_w}^{x_e} \left[ - \frac{1}{\beta} J(p_b, H) + \frac{1}{\rho_o} \nabla \times \Delta \tau + \nabla \times A + \nabla \times B \right] dx,$$

(B1)

where $U = k \times \nabla \psi$ (the subscript here denotes gyre, not geostrophic). Closure of the gyre streamfunction at the western boundary [$\psi(x_w) = 0$] implies the following integral balance of terms:

$$0 = \int_{x_w}^{x_e} \left[ - \frac{1}{\beta} J(p_b, H) + \frac{1}{\rho_o} \nabla \times \Delta \tau + \nabla \times A + \nabla \times B \right] dx.$$

(B2)

In similar fashion, one can derive the steady-state vorticity balance for the large-scale overturning streamfunction. Equation (A10) gives an expression for the vertically integrated meridional transport between the ocean bottom and an interior depth surface ($\int_{z}^{H} dz$). The zonal integral of (A10) from the western boundary to the eastern boundary yields an expression for the net basinwide meridional transport below a given depth $z$. Thus, it implies the following vorticity decomposition of the meridional overturning streamfunction in depth space:

$$\psi_o(z) = \frac{1}{\beta} \int_{x_w}^{x_e} \left[ - \frac{1}{\beta} J(p_b, H)(z) + \frac{1}{\rho_o} \nabla \times \Delta \tau + \nabla \times A + \nabla \times B - f w \right] dx$$

(B3)

where the overturning streamfunction (subscript $o$) is defined in the $y-z$ plane as $(-\partial_z \psi_o, \partial_y \psi_o) = (\int_{x_w}^{x_e} v dx, \int_{x_w}^{x_e} w dx)$. We include $(z)$ in (B3) to stress that terms represent partial vertical integrals up to depth $z$ (e.g., the BPT term only reflects flow interaction with bottom topography that is deeper than $z$). Closing the overturning streamfunction at the surface [$\psi_o(n) = 0$] is mathematically equivalent to the closure of the gyre streamfunction at the western boundary [$\psi(x_w) = 0$] as implied by the same balance of terms obtained earlier [(B2)].

The expression for the overturning streamfunction given by (B3) is a bottom-up perspective that highlights the role of abyssal flow interaction with bottom topography in controlling the deep, southward limb of the AMOC. If we consider a $z$ that intersects the geostrophic interior (say, $z = 1000$ m), then $\psi_o$ will primarily reflect the bottom drag, and $\psi_o^A$, $\psi_o^B$, and $\psi_o^{BPT}$ will reflect the steering of abyssal flow by deep ridges and continental shelves. One could equivalently consider a “top-down” decomposition of the AMOC vorticity balance by performing the vertical integral in (A2) from an interior depth $z$ to the ocean surface $\int_{z}^{H} dz$. Instead of (A10), one gets

$$\beta V(z) = \frac{1}{\rho_o} J(p_b, H) + \frac{1}{\rho_o} \nabla \times \Delta \tau + \nabla \times A + \nabla \times B - f w \bigg|_z,$$

(B4)

since the stretching term in (A6) changes sign when $z$ is the lower boundary of the integration. BPT appears in (B4) because $\int_{z}^{H} dz \to \int_{z}^{H} dz$ in regions where the ocean bottom depth is shallower than $z$. The top-down AMOC vorticity decomposition is then

$$\psi_o(z) = \frac{1}{\beta} \int_{x_w}^{x_e} \left[ - \frac{1}{\beta} J(p_b, H)(z) + \frac{1}{\rho_o} \nabla \times \Delta \tau + \nabla \times A + \nabla \times B - f w \bigg|_z \right] dx.$$

(B5)

In (B5), the surface wind stress curl $\nabla \times \tau_o$ dominates $\psi_o$, since now $\Delta \tau = \tau_o - \tau(z)$, and $\psi_o^A$, $\psi_o^B$, and $\psi_o^{BPT}$ will largely reflect the steering of upper-ocean flow by shallow (western boundary) continental shelves. The equivalence of these two perspectives [(B3) = (B5)] again yields the closure balance [(B2)].


