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Special Section:
Atlantic Meridional Overturning Circulation: Reviews of Observational and Modeling Advances

Key Points:
- Theoretical models have provided a dynamical framework for understanding recent AMOC observations and numerical model results
- We review recent advances in conceptual understanding of the processes setting the strength, structure, and variability of the AMOC
- Recent models go beyond the zonal average, include interaction between wind and buoyancy forcing, and acknowledge the importance of eddies

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Recent Contributions of Theory to Our Understanding of the Atlantic Meridional Overturning Circulation

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Abstract Revolutionary observational arrays, together with a new generation of ocean and climate models, have provided new and intriguing insights into the Atlantic Meridional Overturning Circulation (AMOC) over the last two decades. Theoretical models have also changed our view of the AMOC, providing a dynamical framework for understanding the new observations and the results of complex models. In this paper we review recent advances in conceptual understanding of the processes maintaining the AMOC. We discuss recent theoretical models that address issues such as the interplay between surface buoyancy and wind forcing, the extent to which the AMOC is adiabatic, the importance of mesoscale eddies, the interaction between the middepth North Atlantic Deep Water cell and the abyssal Antarctic Bottom Water cell, the role of basin geometry and bathymetry, and the importance of a three-dimensional multiple-basin perspective. We review new paradigms for deep water formation in the high-latitude North Atlantic and the impact of diapycnal mixing on vertical motion in the ocean interior. And we discuss advances in our understanding of the AMOC's stability and its scaling with large-scale meridional density gradients. Along with reviewing theories for the mean AMOC, we consider models of AMOC variability and discuss what we have learned from theory about the detection and meridional propagation of AMOC anomalies. Simple theoretical models remain a vital and powerful tool for articulating our understanding of the AMOC and identifying the processes that are most critical to represent accurately in the next generation of numerical ocean and climate models.

1. Introduction

A net northward flow of warm thermocline and intermediate waters occurs in the upper kilometer throughout the Atlantic basin. This is compensated by a net southward flow of colder North Atlantic Deep Water (NADW) at depths between approximately 1 and 3 km. Water is transformed from the upper to the lower isopycnal layer in the high-latitude North Atlantic where strong cooling results in significant buoyancy loss, locally eroding the stratification maintained by various processes. Transformation from the lower to the upper isopycnal layer occurs via wind-driven and tidally driven diapycnal mixing throughout the global ocean interior as well as in the surface mixed layer of the Southern Ocean after adiabatic wind-driven upwelling. This *Atlantic Meridional Overturning Circulation* (or AMOC) is responsible for transporting significant heat northward in both hemispheres and also plays a key role in the uptake and distribution of other important tracers such as carbon and oxygen. As such, the AMOC is a fundamental component of the climate system.

Figure 1 illustrates the key processes which determine the strength, structure, and variability of the AMOC. Over the last decade or two, groundbreaking basin-wide observational arrays, such as the RAPID-MOCHA array at 26°N (Cunningham et al., 2007; Srokosz & Bryden, 2015), together with other ocean observations (e.g., Cunningham & coauthors, 2019) and a new generation of ocean and climate models, have provided new and intriguing insights into the mean AMOC and its variability. Theoretical models have also contributed to changing our view of the AMOC, providing a dynamical framework for understanding the new observations and some of the complexity described by comprehensive numerical models. In this paper we review recent advances in the theoretical modeling of the AMOC and the conceptual understanding that has resulted from careful, pared-down thinking about the circulation. Readers also interested in a more general review of the AMOC are referred to Buckley and Marshall (2016).

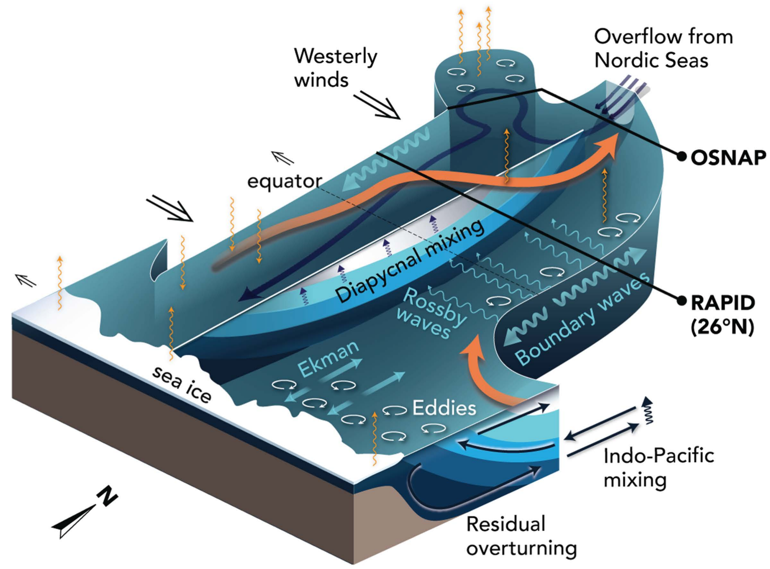

Figure 1. Schematic illustrating the key processes which determine the strength, structure, and variability of the Atlantic Meridional Overturning Circulation. The net northward flow of warm water in the upper kilometer is indicated in red, with the compensating southward flow of colder North Atlantic Deep Water shown by the blue arrow, which deepens along the western boundary. Transformation between these light and dense flows occurs in the subpolar North Atlantic, where buoyancy is lost, and in the Southern Ocean, where buoyancy is gained. Adiabatic flow is possible along isopycnals, illustrated on the vertical section, which outcrop in both regions. In the Southern Ocean, strong westerly winds drive surface Ekman transport and upwelling which steepens isopycnals; the resultant baroclinic instability leads to eddy bolus fluxes which oppose the wind-driven circulation such that the residual overturning circulation, shown in black, is primarily directed along isopycnals in the ocean interior but across isopycnals in the surface mixed layer. A fraction of the upwelled water becomes denser, rather than lighter, in the Southern Ocean, forming Antarctic Bottom Water which subsequently upwells diapycnally to middepths in the Indian and Pacific Oceans. Diapycnal mixing in the Atlantic Ocean interior (dark blue) may also play a role in the water mass transformation. The short extent of the African continent compared to the South American continent promotes import of salty subtropical water from the Indian Ocean as thermocline and intermediate waters return to the Atlantic from the east (red). Mesoscale eddies (white) and wind stress also play a key role in the dynamics and buoyancy budget of the subpolar North Atlantic, with eddies connecting regions of net buoyancy loss to regions of net downwelling. Adjustment of the overturning to changes in wind stress and buoyancy fluxes occurs via boundary and Rossby waves, indicated in pale blue. The locations of the RAPID and Overturning in the Subpolar North Atlantic Programme (OSNAP) observational arrays are marked with black lines.

We consider “recent” advances to include those since the beginning of the AMOC's modern instrumental record, that is, since roughly the start of this century, although we discuss older literature where necessary to demonstrate or motivate a more recent advance. We define “theoretical contributions” broadly, to include analytical theories, conceptual models, and a subset of idealized numerical calculations. We restrict our attention to what is often termed the “upper cell” (more accurately termed the “middepth cell”), that is, the cell whose northward branch transports thermocline and intermediate waters and whose southward branch transports NADW. We discuss the circulation of the “abyssal cell,” associated with Antarctic Bottom Water (AABW) formed in the Southern Hemisphere, only where it is important in understanding the circulation of NADW. The AMOC sometimes refers to an overturning in density space via a diapycnal velocity and sometimes it refers to an overturning in depth space, where it represents the Eulerian vertical velocity (Figure 2). We discuss the AMOC in both density and depth space and strive to make connections between the two.

We begin by reviewing the contributions that theoretical models have made to our understanding of the time-averaged AMOC, focusing on the insight the models have provided, individually and collectively, rather than on their mathematical details. Section 2 describes the inclusion of Southern Hemisphere processes, in particular winds, in theoretical models of the AMOC. Section 3 considers the move away from zonally averaged single-basin models to incorporate more complex spatial structure in forcing and circulation. Section 4 revisits the relationship between meridional density gradients and the AMOC, while section 5 reviews recent theoretical progress concerning AMOC stability. Section 6 summarizes current thinking on why we

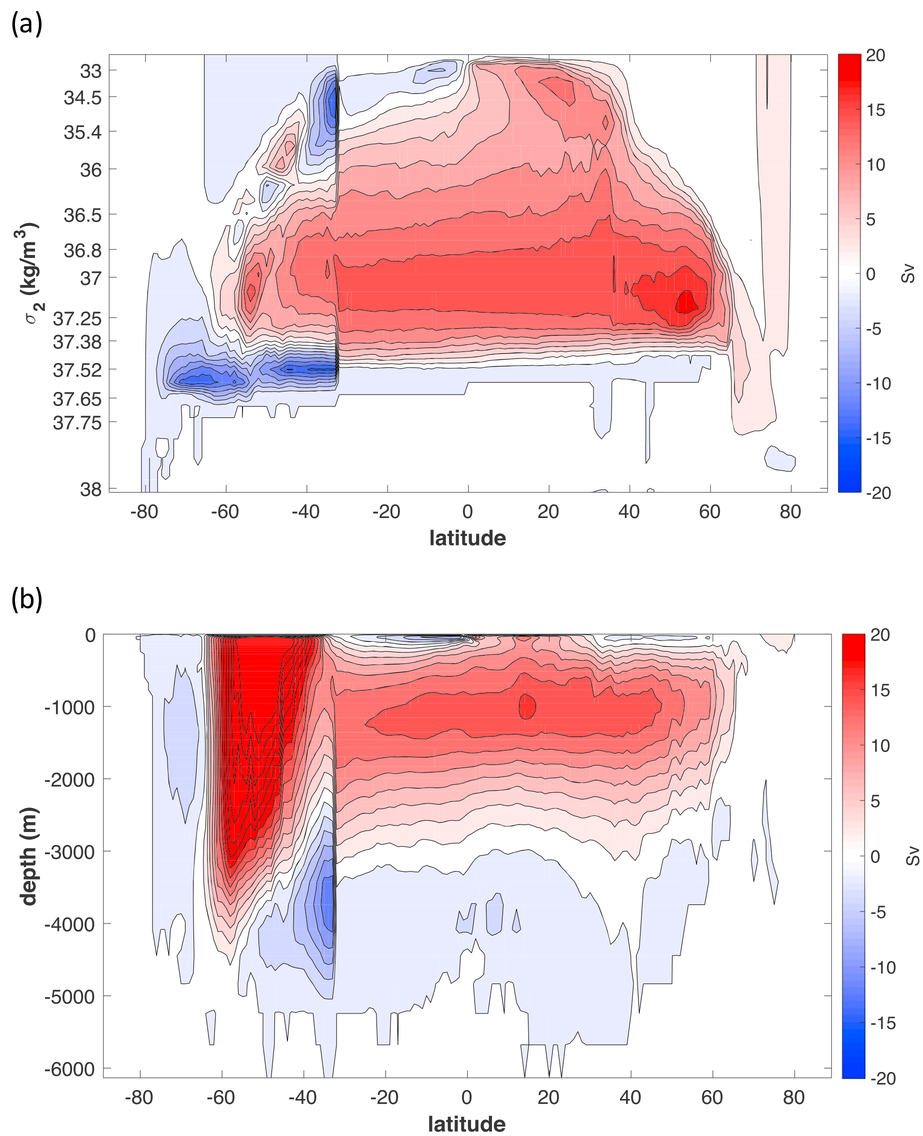

Figure 2. (a) The time-mean residual overturning circulation, zonally integrated over the Atlantic sector, as a function of latitude and σ_2, calculated using the Eulerian plus eddy bolus velocities from the Estimating the Circulation and Climate of the Ocean version 4 state estimate (release 3). South of 33°S, the integral is taken over all longitudes. Notice that at middepths the flow is largely along constant σ_2 surfaces. (b) The time-mean meridional overturning circulation, zonally integrated over the Atlantic sector, as a function of latitude and depth, calculated using the Eulerian velocities from the Estimating the Circulation and Climate of the Ocean version 4 state estimate (release 3). South of 33°S, the integral is taken over all longitudes. Notice the very strong clockwise Ekman cell south of 33°S, which is largely compensated by eddy volume fluxes in the residual framework, and the weaker Northern Hemisphere maximum relative to the residual framework in (a).

have an AMOC at all, rather than a Pacific meridional overturning circulation (PMOC), before section 7 reviews theoretical work on NADW formation and the downwelling limb of the AMOC.

We then move on to review models of AMOC variability and consider what we have learned from theory about the detection and meridional propagation of AMOC anomalies. Section 8 focuses on the value that theoretical studies have added to the monitoring and understanding of AMOC variability at 26°N. Section 9 considers recent advances in our understanding of what drives AMOC variability more generally, and section 10 summarizes alternative frameworks for thinking about the AMOC which provide additional insight. The paper closes in section 11 with a summary of the major contributions that theory has made over the last decade and some speculation about what it may contribute in the next decade.

2. Southern Hemisphere Processes

A key ingredient in almost all theoretical models of the global-scale AMOC developed over the last two decades is the eastward wind stress over the Southern Ocean. Marshall and Speer (2012) provide a helpful review of the crucial role the Southern Ocean plays in the AMOC. Its unique importance arises from the absence of meridional continental barriers, requiring that the input of momentum by the wind stress at the surface is balanced by the bottom form stress across abyssal topographic obstacles (Munk & Palmén, 1951). The surface meridional Ekman layer transport is compensated by geostrophic return flow at depth, such that the wind-driven Ekman overturning cell extends down to the abyss, upwelling cold, dense water to the south, and downwelling warm, buoyant water to the north, creating strong lateral density gradients over the full fluid column. The resultant baroclinic instability generates an intense baroclinic eddy field and eddy-induced isopycnal slumping which opposes the Eulerian mean isopycnal steepening (Danabasoglu et al., 1994). In equilibrium, the net "residual circulation," including both Eulerian-mean and eddy-induced components (Gent et al., 1995), is primarily directed along isopycnals in the ocean interior but across isopycnals within the surface mixed layer, the latter at a rate set by surface buoyancy fluxes (Marshall, 1997; Speer et al., 2000; Marshall & Radko, 2003). This net residual circulation, together with diapycnal mixing over the global ocean, balances the formation of deep water in the North Atlantic (Figure 1).

The impact of Southern Ocean winds on the AMOC was highlighted by Toggweiler and Samuels (1995) in a series of integrations with an ocean general circulation model (GCM). The first theoretical model to include a pivotal role for Ekman transport in the Southern Ocean around Antarctica in setting the mean strength of the AMOC was Gnanadesikan (1999). His simple and elegant box model explores the relative roles of NADW formation at high latitudes in the North Atlantic, Southern Ocean Ekman transport, Southern Ocean eddies, and diapycnal mixing. Many subsequent models have followed up on these ideas and in particular the extent to which the AMOC is directed along isopycnals, such that no water mass transformation (and hence diapycnal mixing) is required within the ocean interior.

The potential for an adiabatic pole-to-pole overturning circulation (Figure 3a) is most clearly illustrated by Wolfe and Cessi (2011) using a zonally averaged, analytical model, in combination with both coarse resolution and eddy-resolving numerical simulations. They show that, in a single ocean basin connected to a reentrant channel, an adiabatic pole-to-pole overturning circulation is possible provided there is (a) a thermally indirect, wind-driven overturning circulation in the reentrant channel and (b) a set of isopycnals which outcrop in both the reentrant channel and the Northern Hemisphere. In the limit of weak interior diapycnal mixing, the net overturning circulation must be dominated by this adiabatic cell. An important, and as yet poorly understood, component of this pole-to-pole overturning cell is the transformation of potential vorticity that must take place in the northward flowing limb to convert the negative potential vorticity coming from the Southern Hemisphere into the positive potential vorticity found in the Northern Hemisphere (and vice versa for the deep southward flowing limb). In GCMs, this transformation occurs in narrow laminar viscous boundary layers where vorticity can be modified adiabatically (Edwards & Pedlosky, 1998), whereas turbulent processes are likely to play a role in the ocean.

Radko and Kamenkovich (2011) also develop an analytical model that allows for adiabatic solutions. Their model combines classical elements of large-scale circulation theory, including simple models of the thermocline, inertial western boundary current, and eddy-controlled Antarctic Circumpolar Current (ACC) in a 2.5 layer framework to predict the stratification and AMOC transport as a function of surface forcing. Inclusion of possible diabatic effects substantially influences the AMOC transport but not the stratification.

The isopycnal connection between the Southern Ocean and the high-latitude North Atlantic links the surface buoyancy distribution at these end points of the AMOC. This implies that, as well as the Southern Hemisphere westerlies, the Northern Hemisphere winds are also important (see sections 8 and 9 and Cessi, 2018). In addition, in the purely adiabatic case there can be no net source or sink of buoyancy within the volume enclosed by the isopycnals connecting the AMOC end points. Since buoyancy is lost at the isopycnal outcrops in the North Atlantic (corresponding to NADW formation; see section 7), either there is diapycnal mixing along the path of the AMOC or buoyancy must be gained in the upwelling region of the Southern Ocean, or both. The contribution of diapycnal mixing is increased if the path of the lower branch of the overturning is extended, and this can be achieved by connecting the return flow of the AMOC to the abyssal AABW cells in the Indian and Pacific in what has become known as a "figure-of-eight" loop (see Figure 1, section 3, and Talley, 2013).

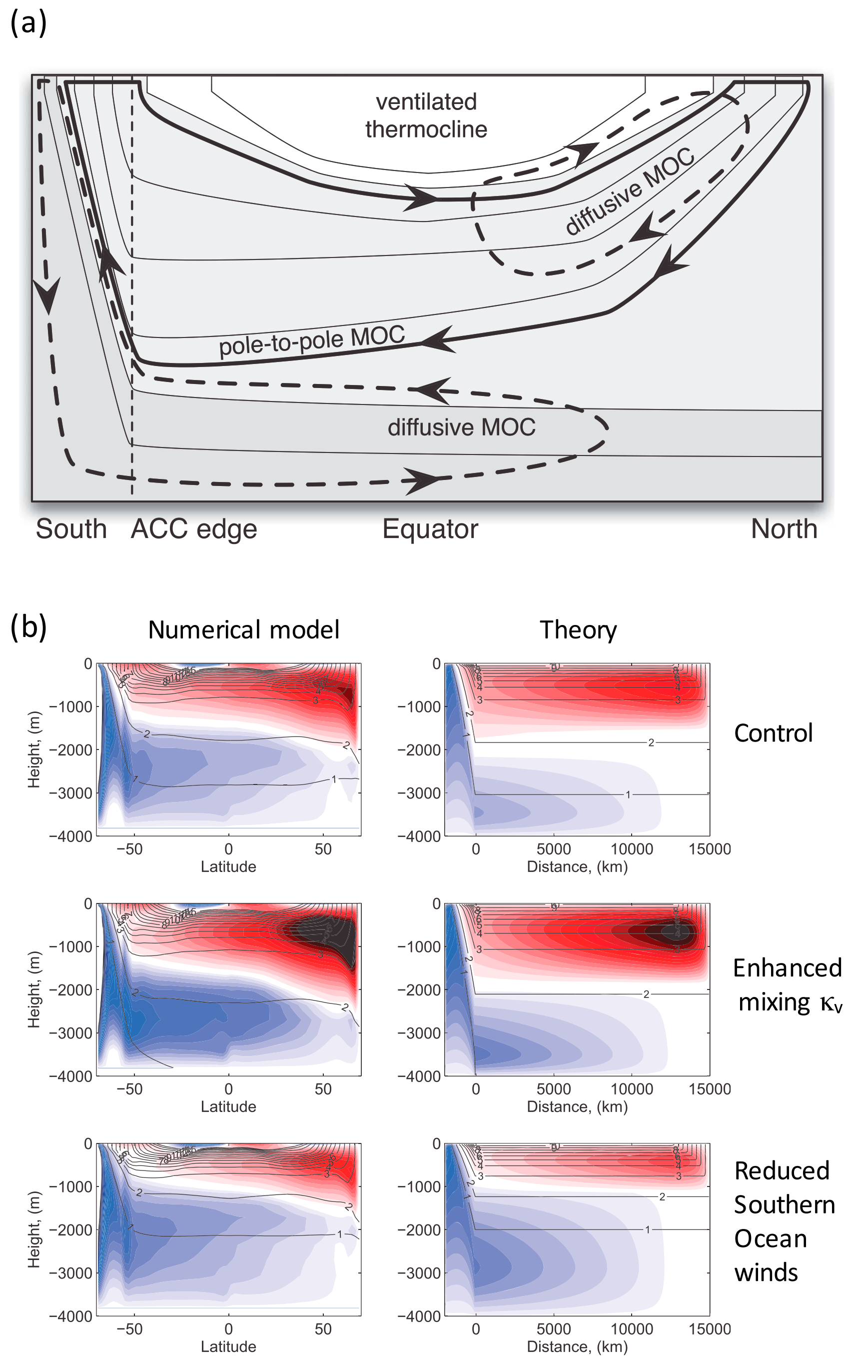

Figure 3. (a) A sketch of the meridional overturning circulation (MOC) from Wolfe and Cessi (2011). A pole-to-pole cell (thick solid line) with sinking in high latitudes in the North Atlantic and upwelling in the Antarctic Circumpolar Current (ACC) region coexists with weaker diffusive cells characterized by high-latitude sinking in each hemisphere and upwelling mostly confined to the same hemisphere (thick dashed lines). The thin solid lines show isopycnals. The isopycnals in the ventilated thermocline region do not outcrop in the ACC region. The isopycnals in the heavily shaded region outcrop in the channel but not in the North Atlantic. The group of three intermediate isopycnals outcrop both in the ACC and the North Atlantic, and it is along these surfaces that an adiabatic pole-to-pole residual overturning circulation can exist, with water mass transformation confined to the mixed layer. (b) Zonally averaged sections of residual circulation (Sv; blue/red) and temperature (°C, black lines) from (left) the general circulation model simulations and (right) theory in Nikurashin and Vallis (2012). The three rows show (top) the control experiment with $\kappa_v = 2 \times 10^{-5}$ m^2/s and $\tau = 0.2$ N/m^2, (middle) an enhanced mixing experiment with $\kappa_v = 5 \times 10^{-5}$ m^2/s, and (bottom) a reduced Southern Ocean wind experiment with $\tau = 0.1$ N/m^2.

Shakespeare and Hogg (2012) develop a three-layer extension of Gnanadesikan (1999) to include the abyssal cell showing that, in the adiabatic limit, the upper cell's overturning circulation scales linearly with both Southern Ocean wind stress and Northern Hemisphere buoyancy loss. They note, however, that buoyancy loss around Antarctica is also important because of the role it plays in setting the stratification; the upper and abyssal cells are coupled due to the requirement for thermodynamic equilibrium.

Although analytically tractable only in the adiabatic limit, the model of Nikurashin and Vallis (2012) is notable as the first theory for the zonally averaged stratification and overturning circulation, as continuous functions of depth and latitude, that includes both adiabatic and diabatic components. The model matches solutions in three regions—the circumpolar channel, northern isopycnal outcrop region, and the basin in between—and agrees well with the results of three-dimensional numerical simulations in a single-basin domain with a reentrant channel to the south (Figure 3b). The upper cell overturning circulation scales linearly with the Southern Ocean westerly winds in the weak diapycnal and eddy mixing limit, while for strong diapycnal mixing (or weak winds) the overturning scales as $\kappa^{2/3}$, where κ is the coefficient of diapycnal mixing, in line with many previous studies (Welander, 1971).

Before moving on from zonally averaged models, we note an important caveat in our theoretical understanding of the AMOC, which is the reduced sensitivity of both the slope of the density surfaces and the residual overturning in the Southern Ocean to the surface wind stress in models with explicit, rather than parameterized eddies, known, respectively, as “eddy saturation” and “eddy compensation” (e.g., Tansley & Marshall, 2001; Hallberg & Gnanadesikan, 2001, 2006; Farneti et al., 2010; Viebahn & Eden, 2010; Abernathey et al., 2011; Morrison & Hogg, 2013; Munday et al., 2013). Both eddy saturation and eddy compensation may impact the relationship between the AMOC and Southern Hemisphere winds: the slope of density surfaces across the Southern Ocean is important in determining the equilibrium stratification of the Atlantic (following Gnanadesikan, 1999), while the residual overturning in the Southern Ocean must equal the strength of the AMOC in the limit of an adiabatic pole-to-pole circulation (Wolfe & Cessi, 2011). As a consequence, the linear relationship between the AMOC and the strength of Southern Hemisphere winds exhibited by models such as Shakespeare and Hogg (2012) and Nikurashin and Vallis (2012) may overestimate the sensitivity of the AMOC, especially under strong winds (see, e.g., Radko & Kamenkovich, 2011; McCreary et al., 2016). Preliminary investigations with a new eddy closure, in which the eddy diffusivity varies in proportion to the eddy energy field, suggest that it may be possible to capture the effects of eddy saturation and eddy compensation without explicitly resolving the eddies (Mak et al., 2018), although the robustness of these results and detailed implications for the sensitivity of the AMOC to Southern Hemisphere wind stress requires further investigation.

3. From Two Dimensions to Three Dimensions to Multiple Basins

Samelson (2009) and Bell (2015) take a different approach to including the impact of surface wind stress, allowing the stratification to vary zonally as well as meridionally in response to the wind forcing, which then in turn allows for zonal asymmetry in the effectiveness of surface buoyancy fluxes. Bell (2015) balances the water mass transformation due to surface heat loss in the Northern Hemisphere with that due to the wind-driven shoaling of isopycnals in the south west Atlantic, just north of Drake Passage, where heat is gained by the ocean (Figures 1 and 4). The strength of the overturning circulation in this model depends on the range of wind stresses over the Southern Ocean, and the southernmost latitude at which heat loss occurs in the North Atlantic, consistent with both a linear dependence on Southern Hemisphere winds and with the importance of a shared range of isopycnal outcrops between hemispheres.

Although Bell (2015) and the reduced-gravity model of Samelson (2009) retain a simple layered structure in the vertical and are restricted to a single basin, the move away from a zonally averaged picture permits the wind stress to play a more complex role in the dynamics of the overturning circulation, an emerging theme which is also reflected in some theoretical models of AMOC variability (see sections 8 and 9), and appears to be increasingly borne out by observations and GCM simulations (e.g., Polo et al., 2014; Williams et al., 2014; Pillar et al., 2016). To extend these kinds of theoretical solutions to the full global ocean will require a three-dimensional matching condition between the circulations in the Southern Ocean and in the Atlantic: Marshall et al. (2016) and McCreary et al. (2016) have recently extended ideas of Gill (1968) to explore how a dynamically consistent solution for the Southern Ocean and Atlantic basin affects the stratification, overturning, and ACC.

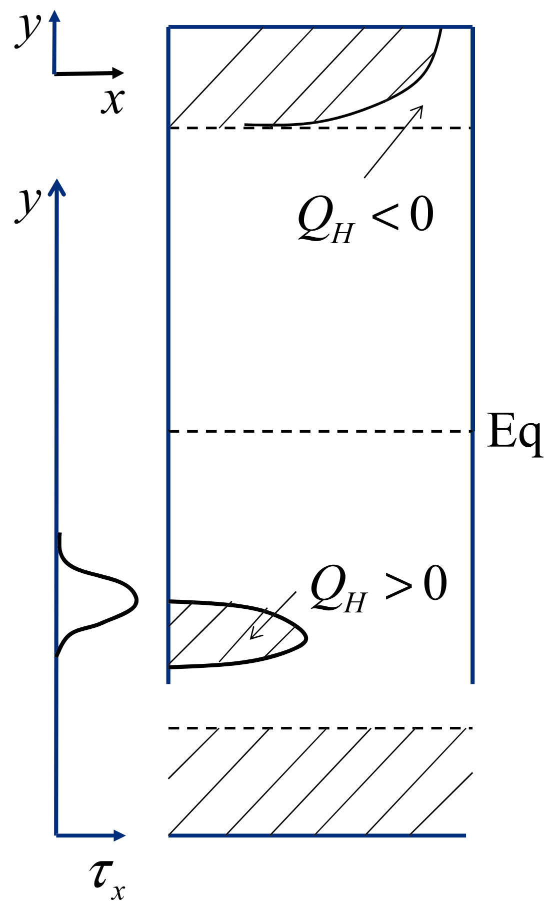

Figure 4. Schematic illustrating the regions in which there is a nonzero air-sea heat flux Q_H in the two-layer isopycnal model of Bell (2015). The ocean occupies a rectangular basin spanning the equator with a periodic channel at its southern boundary. The lower layer shoals or outcrops in the diagonally shaded regions. The zonal wind stress $\tau_x(y)$ is shown on the left of the figure. Heat loss in the north of the basin ($Q_H < 0$) is balanced by heat gain in the southwest of the basin ($Q_H > 0$) where wind-driven shoaling of isopycnals brings dense water to the surface.

The move away from zonally averaged models is symptomatic of a trend over the last decade from two-dimensional to three-dimensional AMOC theories, and it is becoming increasingly clear that models need to consider more than one basin. Evidence from hydrographic observations suggests that much of the NADW that upwells in the Southern Ocean, rather than being directly converted to thermocline and intermediate waters by buoyancy input at the surface, is exported to depth as AABW, from where it must subsequently upwell diffusively, mostly in the Indo-Pacific, to rejoin the upper cell (Figure 1; Schmitz, 1995; Talley, 2013).

Thompson et al. (2016) present a multibasin residual-mean model of the global overturning circulation that involves zonal mass transport between basins via the ACC and the Indonesian Throughflow (Figure 5). The model is formulated as a two-basin, four-layer, box model. A portion of the NADW that flows into the Southern Ocean is returned directly upward and northward in the Atlantic sector to close the AMOC cell, while another portion produces a convergence of abyssal waters, which flows into the Indo-Pacific sector of the Southern Ocean. From the abyssal Indo-Pacific this water upwells diapycnally to eventually rejoin the upper cell. This model finds that closure of the overturning cell directly in the Atlantic versus closure through the Indo-Pacific abyss depends on the amplitude of the overturning relative to the deep diffusivity, with the former prevailing for large overturning (or weak diffusivity). The results suggest a minimal role for a closed overturning cell in the present-day Atlantic alone and have implications for the distribution and residence time of tracers, as well as the pathways and sensitivity of the overturning to changed forcings (e.g., see Ferrari et al. (2014) who consider how the circulation may have worked under glacial boundary conditions).

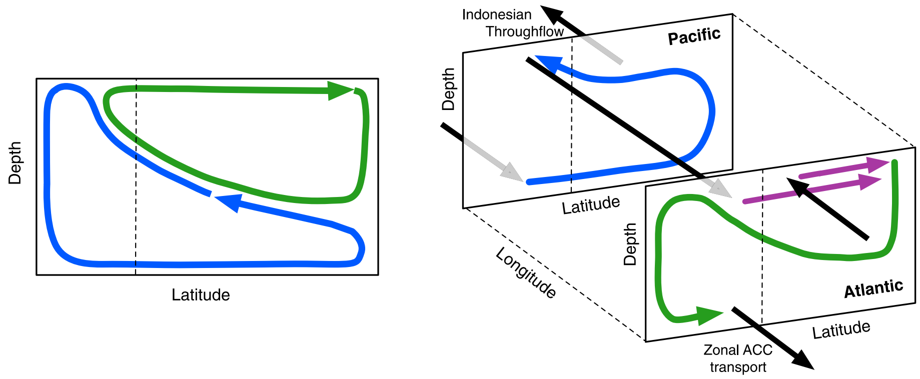

Figure 5. Two-dimensional and three-dimensional depictions of the meridional overturning circulation from Thompson et al. (2016). The schematic on the left shows the zonally averaged overturning in depth-latitude space. The green and blue curves are typically viewed as distinct overturning cells associated with North Atlantic Deep Water formation and Antarctic Bottom Water formation, respectively. The idealized three-dimensional schematic on the right illustrates the single figure-of-eight overturning circulation of Talley (2013). Here, the overturning cycles through both the Atlantic and Pacific Basins, either through the Antarctic Circumpolar Current (ACC) or the Indonesian Throughflow, before closing. Rather than two distinct cells, the overturning more closely approximates a single figure-of-eight loop.

Other studies have addressed the same question using a reduced-gravity framework. Building on the work of Johnson and Marshall (2004) and Allison (2009), Jones and Cessi (2016) present a two-basin version of the Gnanadesikan (1999) model which shows that a difference in layer depth between the basins leads to a geostrophically balanced exchange flow from the Pacific-like (nonsinking) basin to the Atlantic in the surface layer, compensated at depth. Ferrari et al. (2017) use a similar two-layer, two-basin model to illustrate that, in the present-day climate, the overturning circulation is best described as the combination of three circulations: an adiabatic overturning circulation in the Atlantic Ocean, associated with transformation of thermocline and intermediate waters to deep waters in the north, a diabatic overturning circulation in the Indo-Pacific Ocean, associated with transformation of abyssal to deep waters by mixing, and an interbasin circulation that exchanges waters geostrophically between the two basins through the Southern Ocean.

These models reflect our increasing appreciation that, even if an adiabatic overturning is possible within the Atlantic basin, water mass transformation in the Indo-Pacific plays an important role in the buoyancy budget of the global overturning circulation (e.g., Newsom & Thompson, 2018) and impacts indirectly on the overall strength of the AMOC.

4. Does the AMOC Scale With Meridional Density Gradients?

It is clear from the papers reviewed in section 2 that the overturning circulation depends strongly on Southern Hemisphere wind stress, as well as diapycnal mixing. However, surface buoyancy loss in the North Atlantic is critical for the formation of NADW, and earlier conceptual models of a thermohaline circulation assumed that the overturning should depend on the large-scale meridional density gradient. Many authors (beginning with Wright & Stocker, 1991; Wright et al., 1995, and more recently Wolfe & Cessi, 2010; Cimatoribus et al., 2013) have sought to reconcile this with the fact that meridional transports must be in thermal wind balance with zonal, rather than meridional, density differences.

In reduced-gravity models, fast boundary waves act to remove pressure gradients along the equator and along the eastern boundary (e.g., Cessi & Louazel, 2001; Johnson & Marshall, 2002a; Marshall & Johnson, 2013, 2017; Bell, 2015;). When buoyancy fluxes force the layer interface to outcrop along the eastern boundary, the resulting pressure gradient drives a convergence of flow toward this boundary which is balanced by deep water formation. Rossby waves propagate the eastern boundary pressure westward across the basin such that the deep water formation region is zonally connected to the western boundary. The strength of the overturning circulation in such models is proportional to the depth-integrated meridional pressure difference, in turn proportional to $\Delta\rho H^2$ where $\Delta\rho$ is the difference between the layer densities and H the eastern boundary layer thickness. Assuming that upwelling occurs outside the North Atlantic and that the large-scale circula-

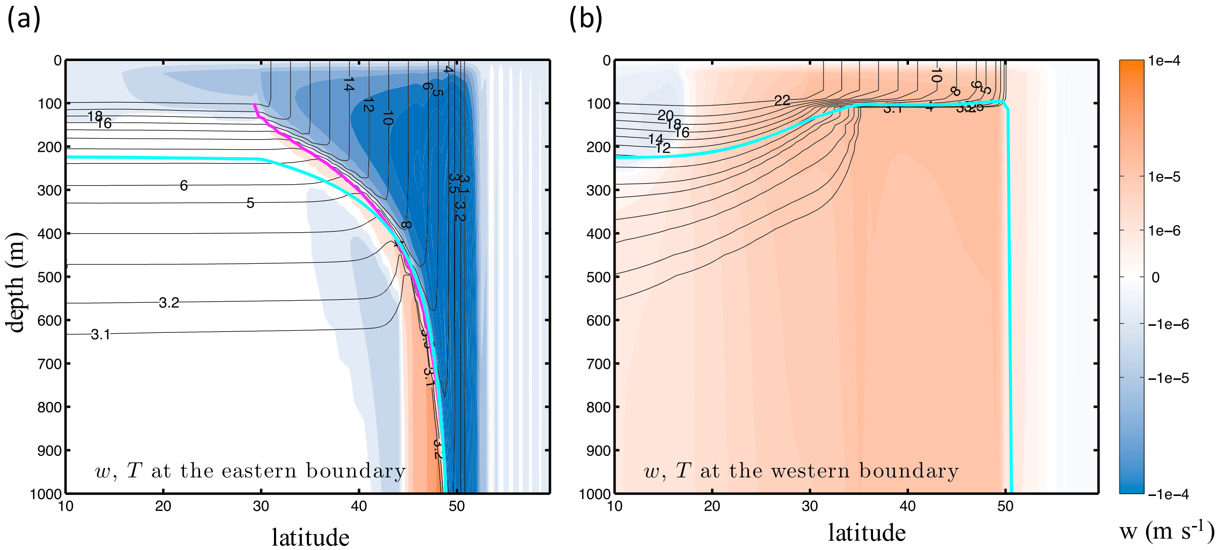

Figure 6. Temperature (contours) and vertical velocities (shading) as a function of depth and latitude along (a) eastern, and (b) western boundaries in an idealized general circulation model simulation of the circulation in a rectangular, single hemisphere domain driven by a large-scale surface temperature gradient and winds (Schloesser et al., 2014). The magenta curve indicates the eastern boundary mixed layer thickness derived by Sumata and Kubokawa (2001), and cyan curves indicate analytical solutions for the eastern and western boundary layer depths in a two-layer model.

tion is geostrophic, it follows that zonal and meridional pressure gradients are closely related. This scaling also highlights that the vertical stratification is critical in determining how a meridional density difference $\Delta\rho$ is converted into a pressure gradient.

While the circulation in GCMs forced by large-scale buoyancy fluxes is more complex, it shares many of the key features described above. In particular, numerical model experiments agree with simple layer models and scaling arguments in finding that the AMOC strength is proportional to the depth-integrated meridional pressure difference $\Delta\rho H^2$ (e.g., Robinson & Stommel, 1959; Bryan, 1987; Marotzke, 1997; Gnanadesikan, 1999; de Boer et al., 2010). The depth scale H is generally interpreted as the pycnocline depth, but several recent studies have made progress in developing more useful scalings. For example, de Boer et al. (2010) choose the density difference at the surface (or integrated over 0-1,400m) between the equator and high latitudes and show across a broad suite of GCM experiments that the AMOC scaling is only appropriate if H is interpreted as the depth of the maximum overturning streamfunction rather than the depth of the pycnocline.

Schloesser et al. (2012, 2014) and Butler et al. (2016) base their scalings directly on the meridional pressure gradient rather than the density gradient, showing that meridional overturning should therefore scale with the meridional density difference twice integrated in the vertical. Because isopycnal slopes are relatively weak along the equator, taking the density difference along the eastern or western boundary in a closed basin yields similar results. Butler et al. (2016) retain the depth dependence in $\Delta\rho$ to obtain a scaling for the meridional overturning streamfunction as a function of depth for each basin, which agrees well with both deterministically and stochastically buoyancy forced GCM experiments on multidecadal and longer timescales. Both de Boer et al. (2010) and Butler et al. (2016) use gradients along the western boundary in their scalings.

The importance of the western boundary in meridional density scalings is also emphasized by Sijp et al. (2012) and is consistent with the results of Marshall and Pillar (2011) who diagnose the rotational component of the forces in the meridional momentum equation, that is, the nondivergent part of each force, which projects directly onto the acceleration of the flow. They show that on the western boundary the rotational component of the buoyancy force cannot be compensated by the rotational component of the Coriolis force, which vanishes at the wall, and hence, an overturning cell is accelerated along the western boundary.

In response to a large-scale meridional density gradient, water sinks and the thermocline deepens along the eastern boundary (see Sumata & Kubokawa, 2001, and Figure 6). The unstable eastern boundary isopycnal structure is quickly homogenized by mixing processes as it is propagated westward by Rossby waves

(Schloesser et al., 2012), and the resulting flow convergence and water mass transformation are similar to that in simple isopycnal models which do not include these eastern boundary processes. By documenting how instability processes erode the eastern boundary density structure and lead to northward convergence in an eastern boundary current in an eddy-resolving GCM, Cessi and Wolfe (2013) reconcile the opposing requirements of a meridional gradient in buoyancy at the surface and no flow normal to the boundary, and highlight the importance of ageostrophic eastern boundary processes (Figure 1).

As a consequence of these boundary processes, different circulations develop in response to surface density gradients along lateral boundaries compared with in the ocean interior (see also section 7). Pedlosky and Spall (2005) explore how buoyancy-forced dissipation of Rossby waves as they travel westward across the basin creates convergence of poleward surface flow in the North Atlantic, away from the lateral boundaries. Although overturning circulation driven by density gradients in the interior of an ocean basin also scales with the meridional pressure gradient, it is substantially weaker than that associated with density gradients along lateral boundaries (e.g., Schloesser, 2015). Hence, it is the density gradient along the boundaries that dominates in setting the strength of the large-scale AMOC (Spall & Pickart, 2001).

Exploiting the relative flatness of the isopycnals along the eastern boundary, equatorward of regions of deep water formation, Marshall and Johnson (2017) derive a simple relationship based on thermal wind balance for the relative strengths of the AMOC and the ACC (excluding contributions to the latter from bottom flow). This relationship involves the ratios of three depth scales—the depth of the maximum AMOC streamfunction, the depth of the ACC, and the *e*-folding depth of the stratification in the Atlantic—and explains the 8 ± 2 ratio between the mean ACC and AMOC volume transports.

While surface buoyancy fluxes alone are sufficient for generating a meridional overturning circulation in idealized model experiments (e.g., Gjermundsen & Lacasce, 2017), large-scale wind forcing allows for the lateral separation of the northward and southward flows. For example, in the subpolar region of the Northern Hemisphere, the northward, surface branch of the AMOC occurs in the interior of the subpolar gyre, while the southward return flow occurs along the western boundary. Because water is cooled as it flows northward in the gyre, wind forcing provides an additional mechanism for convergence of the surface meridional flow in the North Atlantic (Schloesser et al., 2014). This, together with the importance of meridional Ekman transport for the AMOC (see section 8), impacts the applicability of scalings based on the meridional density gradient in realistic numerical simulations and the ocean.

5. AMOC Stability

The debate about whether meridional density gradients set, or merely reflect, the strength of the AMOC has important implications for our understanding of its stability, which depends on thermohaline feedbacks. Northward advection of salt by the overturning circulation results in the amplification of overturning anomalies and the possibility of abrupt change between bistable regimes (the salt advection feedback identified by Stommel (1961)). Many studies over the past decade have attempted to reconcile the existence of multiple equilibria with a mechanically rather than buoyancy forced AMOC. Several authors have tackled this problem using box models, beginning with Johnson et al. (2007) who incorporated salt advection into the Gnanadesikan (1999) model using a Stommel (1961) formulation. Fürst and Levermann (2012), Cimatoribus et al. (2014), and others have followed up, and it is clear that multiple equilibria and abrupt transitions between them are possible, at least in a box model framework, in both wind- and mixing-driven regimes.

More recently, Wolfe and Cessi (2014, 2015) have explored the role of salt advection in setting the range of isopycnal outcrops that are common to both hemispheres, showing that even in an adiabatic pole-to-pole overturning circulation, salt advection can lead to hysteresis and oscillatory behavior in an idealized numerical simulation of a single basin with a reentrant channel.

An interesting but controversial suggestion has been that the northward freshwater transport across 30°S by the meridional overturning circulation, commonly referred to as M_{ov}, acts as an indicator of whether the AMOC is in a bistable regime, that is, a regime in which both a strong, thermally driven overturning circulation and a weakened or reversed circulation are stable. Based on a simple model, de Vries and Weber (2005) argue that multiple equilibria can only exist when the overturning circulation exports freshwater from the Atlantic basin, that is, when $M_{ov} < 0$ and the upper, incoming branch of the overturning is saltier than the NADW that leaves the Atlantic. If $M_{ov} > 0$, the feedback is damping and the AMOC is monostable.

This idea has been proposed, fleshed out, explored, and tested in a range of theoretical, idealized, and more realistic models (see discussion in Drijfhout et al. (2013)). Refer to Dijkstra (2007), Huisman et al. (2010), and Cimatoribus et al. (2012, 2014) for evidence of the role of M_{ov} in stability; Weber et al. (2007), Drijfhout et al. (2011), Hawkins et al. (2011), Weaver et al. (2012), Mecking et al. (2016), and Gent (2018) for M_{ov} diagnostics in climate models; and Liu and Liu (2013), Liu et al. (2013), and Sijp (2012) for other, related, stability indicators. Rather worryingly, M_{ov} is positive in many climate models, yet negative when diagnosed from observations (Liu & Liu, 2014; Mecking et al., 2017), suggesting that many climate models have a fresh South Atlantic thermocline bias and may not correctly represent the stability of the AMOC. The horizontal gyre component of freshwater transport into the Atlantic is also important: some models suggest that it is of the opposite sign to, and larger in magnitude than, the overturning component (Gent, 2018). Evidence in support of the mechanism in GCMs is inconclusive, and the importance of M_{ov} remains under debate.

The stability of the AMOC is the subject of another review paper in this special collection, so we refrain from further discussion here and refer the reader to Weijer et al. (2019).

6. Why Does Deep Water Form in the North Atlantic But Not the North Pacific?

Related to the ongoing debate about the stability of the AMOC is the question of why we have an AMOC, rather than a PMOC, in the first place. It is well established that the North Pacific is about 2 psu fresher than the North Atlantic, and this small difference prevents deep water formation in the Pacific (Warren, 1983). Therefore, to understand why we have an AMOC instead of a PMOC, the higher salinity of the Atlantic over the Pacific must be explained.

A recent review summarizes the state of knowledge on the Atlantic-Pacific salinity difference (Ferreira et al., 2018), and its conclusions are summarized here. Atmospheric and oceanic processes both contribute to the asymmetry in the salinity distribution, in approximately equal proportions. The main reasons that atmospheric processes favor more precipitation in the Pacific than in the Atlantic sector are that (i) evaporation occurs mainly in coastal regions which represent a larger fraction of the area in a narrow basin (Schmitt et al., 1989), (ii) there is a fetch of about 3,000 km before coastal evaporation precipitates, penalizing precipitation in the narrow Atlantic sector in favor of the wide Pacific sector (Ferreira et al., 2010), and (iii) the monsoonal circulation localizes moisture convergence in the Pacific (Emile-Geay et al., 2003).

The main oceanic processes favoring Atlantic salinification are (i) the advection of salt-rich subtropical waters northward by the AMOC itself, which maintains the preference for Atlantic sinking—this is the salt advection feedback (Stommel, 1961), and (ii) the relatively short meridional extent, and hence subtropical termination, of the African continent east of the Atlantic, which promotes import of salty subtropical water from the Indian Ocean (Reid, 1961; Gordon, 1986; Nilsson et al., 2013; Cessi & Jones, 2017). Because of the massive wind-driven subtropical gyre of the Southern Hemisphere (the "supergyre" of Ridgway & Dunn, 2003; Speich et al., 2007), thermocline and intermediate water flows westward around the tip of South Africa, that is, from the Indian to the Atlantic, rather than returning to the Atlantic eastward from the Pacific around South America. This path around the supergyre results in an interbasin transport of salt in the Southern Ocean, from the Indo-Pacific toward the Atlantic (see Figures 7 and 1 and Cessi & Jones, 2017; Nilsson et al., 2013). The dense outflow from the highly evaporative Mediterranean Sea has previously been considered important for the preference of Atlantic over Pacific sinking (Reid, 1979). However, recent studies (Blanke et al., 2006; Jia et al., 2007) do not conclusively support the notion that the outflow from this semienclosed area contributes to the salt budget of the upper limb of the AMOC beyond the salinification that would be effected by a freshwater loss of about 0.05 Sv over a comparable area of the Atlantic.

It remains unclear which, if any, of the three atmospheric and two oceanic mechanisms listed above is most important and whether any are necessary rather than simply contributing. The multiplicity of causes may reflect an intrinsic complexity of the system or simply our lack of understanding to date.

In addition to having higher salinity, the North Atlantic is more connected to the Arctic than the North Pacific is, and this facilitates the formation of deep water as discussed in the following section.

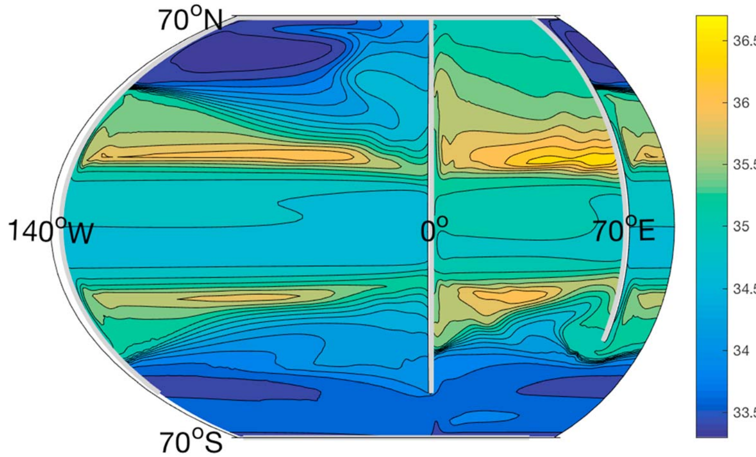

Figure 7. The time-mean surface salinity in a numerical model of the ocean circulation in two idealized basins separated by narrow continents (in gray) and connected by a circumpolar channel to the south. The model is forced at the surface by prescribed zonally uniform wind stress, freshwater flux, and temperature relaxation (Cessi & Jones, 2017). The dynamics spontaneously generate a zonally asymmetric circulation, with a meridional interhemispheric overturning in the narrow basin and no overturning in the wide basin. Despite zonally uniform surface freshwater flux, the surface salinity near 70°N is almost 2 psu larger in the narrow basin, where sinking takes place, than in the wide basin where there is no sinking. The subtropical interbasin connection at 35°S is crucial for the salinification of the narrow basin and the localization of the overturning, because it allows transport of salty subtropical waters from the wide to the narrow basin.

7. Deep Water Formation and the Downwelling Limb of the AMOC

Buoyancy loss at high latitudes in the North Atlantic and Nordic Seas results in both a densification of near-surface waters and a net downwelling from the upper ocean to middepths. This transformation connects the northward flowing upper limb of the AMOC to the southward flowing lower limb of the AMOC. While the buoyancy loss takes place over large areas of the ocean, the dominant water masses are formed in relatively isolated regions of deep convection found in the Labrador Sea, Mediterranean Sea, Greenland Sea (Marshall & Schott, 1999), and more intermittently, the Irminger Sea (Pickart et al., 2003; de Jong & de Steur, 2016). The properties of the waters formed in these regions determine the isopycnals that connect the downwelling limb of the AMOC to the outcropping regions in the Southern Ocean.

It is important to distinguish between water mass transformation and Eulerian downwelling. Both are important, and although connected, they are distinct in magnitude, location, and physics. Send and Marshall (1995) found that the net downwelling in a nonhydrostatic model of buoyancy-forced convection was nearly zero, suggesting that the dominant regions of water mass transformation are not regions of net sinking. This was also inferred from the low-resolution basin-scale models of the AMOC by Marotzke and Scott (1999) and Spall and Pickart (2001), where it was shown that regions of deep convection are found in the basin interior while regions of net sinking are found near the boundaries. The localization of downwelling near side walls was anticipated by the linearized theories of Barcilon and Pedlosky (1967), Spall (2003), and Pedlosky (2003) and is consistent with the theoretical work by MacCready (1994) on the slow decay and descent of deep boundary currents as they travel around a basin. For stratified flows, the downwelling width scales with the baroclinic deformation radius times the square root of the horizontal Prandtl number. The buoyancy loss, due to direct cooling by the atmosphere or lateral eddy fluxes into the interior, is balanced by downward advection of the mean stratification (the downwelling limb of the AMOC) and alongstream advection in narrow boundary currents. The downwelling produces stretching of planetary vorticity, which is balanced by lateral diffusion of vorticity into the boundary. The concentration of downwelling near the boundaries is supported by observations (Pickart & Spall, 2007), in laboratory experiments (Cenedese, 2012), and in eddy-permitting GCMs (Waldman et al., 2018).

Since the large-scale upper ocean flows are in geostrophic balance to leading order, the net inflow to/outflow from a marginal sea at any given depth, which is an indication of up/downwelling within the marginal sea, is proportional to the pressure change across the opening. The pressure field is, to a good approximation, hydrostatic, and therefore, the net downwelling is indirectly determined by the density field. To understand

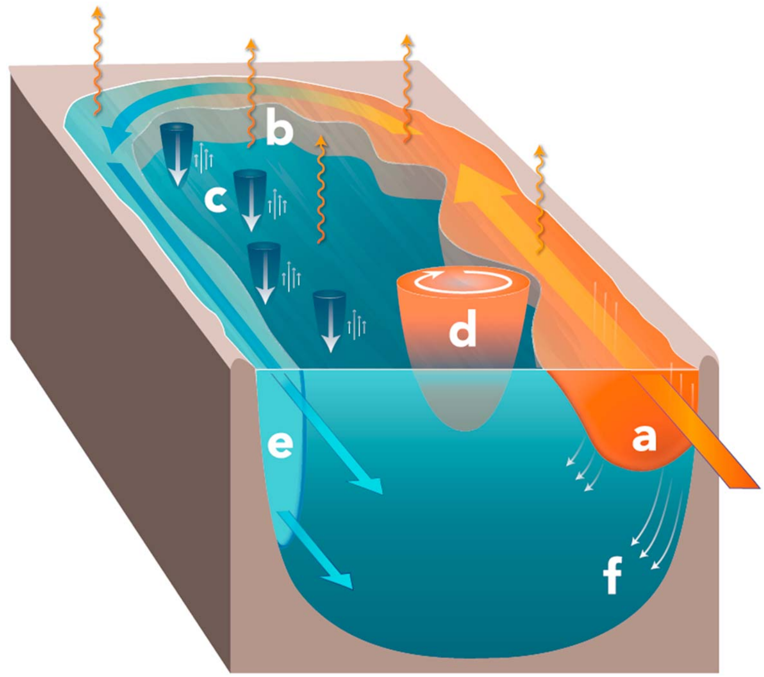

Figure 8. Schematic illustrating the key circulation features operating in North Atlantic marginal seas where dense water formation occurs: (a) warm inflowing boundary current, (b) buoyancy loss to the atmosphere, (c) convective plumes, (d) buoyant eddies shed from the boundary current, (e) cooler outflowing boundary current, and (f) downwelling along the topography.

what determines the net downwelling, one must first understand the buoyancy budget of the marginal sea. Regions of deep convection in the North Atlantic share some common characteristics: They have weak mean flows in the interior, doming isopycnals and are surrounded by a cyclonic boundary current (Marshall & Schott, 1999). Heat loss in the basin interior is balanced by lateral eddy fluxes from the boundary current. This results in an increase in density along the boundary and a barotropization of the boundary current as it flows cyclonically around the basin. In order to remain in geostrophic balance with this reduced vertical shear, downwelling is required near the boundary, from the upper layers to the depths of deep convection. This is the downwelling limb of the AMOC that is forced by surface buoyancy fluxes. The key processes are illustrated in Figure 8. Buoyancy loss by deep convection in the interior (diapycnal AMOC) and vertical transport near the boundary (Eulerian AMOC) are connected through lateral eddy fluxes. The temperature and salinity of the convective water mass and the waters that are exported from the basin depend fundamentally on the balance between mean advection, eddy fluxes, and air-sea exchange (Spall, 2004, 2012; Straneo, 2006). Mesoscale eddies play two important roles in the downwelling limb of the AMOC. First, waters are able to downwell, while isopycnals are rising (following the mean flow cyclonically around the basin), even in the limit of weak diapycnal mixing, through the action of mesoscale eddies, as revealed by a Transformed Eulerian Mean formulation (Cessi et al., 2010; Spall, 2010). Eddies play a similarly fundamental role in balancing the upwelling of the lower branch of the AMOC in the Southern Ocean in the presence of downward Ekman pumping (Danabasoglu et al., 1994; Marshall, 1997; Marshall & Radko, 2003). Second, eddies provide the heat and salt to balance air-sea exchange in the basin interior. In regions where the boundary current is very stable, eddies are not able to flux much heat into the interior and the atmosphere strongly cools the basin. This decreases the heat flux into the atmosphere because the mean air-sea temperature difference is reduced. On the other hand, if the boundary current is able to shed warm, salty eddies into the interior, the regions of deep convection are warmer and the resulting air-sea heat flux is increased. So eddies are essential for proper representation of both the Eulerian and the diapycnal components of the AMOC, and air-sea exchange, in regions of deep convection.

Finally, we note that an exception to the gradual descent of dense water within deep boundary currents occurs when dense water exits a marginal sea over a sill and encounters the lighter ambient waters of the Atlantic, resulting in a dense overflow and significant entrainment. The dynamics were first explored in a numerical streamtube (one-dimensional) model by Price and Baringer (1994). While there has been significant recent progress on improving the representation of overflows and entrainment in ocean and climate models (e.g., Legg et al., 2009) and their impact on the AMOC and its constituent water masses may be significant (e.g., Danabasoglu et al., 2010), this improved understanding has yet to be incorporated into basin-scale theoretical models of the AMOC.

8. Monitoring and Understanding Observed AMOC Variability at 26°N

Since 2004 the RAPID-MOCHA array has been measuring the AMOC at 26°N. Profiles of density on eastern and western boundaries of the Atlantic (as well as on either side of the mid-Atlantic ridge) are used to determine the geostrophic ocean interior flow via thermal wind balance. When combined with cable measurements of the flow through Florida Straits, direct current meter measurements on the western boundary, an estimate of the surface Ekman layer transport based on atmospheric reanalysis winds, and a mass conservation constraint, the array provides daily estimates of the strength and vertical structure of the meridional overturning circulation (Cunningham et al., 2007; McCarthy et al., 2015). This unprecedented decade-long record of the AMOC in the subtropical North Atlantic has revealed large amplitude variability on all timescales accessible to date, as well as what appears to be a longer-term trend (Smeed et al., 2014; Srokosz & Bryden, 2015).

Theoretical studies have played a key role in both challenging and defending the ability of an array such as RAPID-MOCHA to capture long-term trends. One important example concerns the potential role of eddies in low-frequency AMOC variability at 26°N. Wunsch (2008) projected sea surface height variability, which reaches a root-mean-square amplitude of 16 cm near the western side of the subtropical Atlantic, onto vertical modes of horizontal velocity to show that eddies alone could result in a 16 Sv root-mean-square transport variability measured at the array. Wunsch (2008) therefore predicted that any signal of longer-term change in the AMOC at the RAPID-MOCHA array should be swamped by eddies. However, Kanzow et al. (2009) used linear wave theory, together with simple reduced-gravity simulations, mooring, and altimetry data to show that, in fact, the eddy energy drops off sharply right at the western boundary. Matching incoming long Rossby waves with reflected short Rossby waves at the western boundary ensures that density anomalies on the boundary are significantly smaller than those associated with the incoming Rossby wave. Kanzow et al. (2009) and Zhai et al. (2010) show that the same is true for nonlinear eddies in reduced-gravity models. Therefore, provided thermal wind balance for the meridional flow is applied all the way across the basin from boundary to boundary, eddies do not dominate the AMOC variability signal. (Clément et al. (2014) use mooring data to show that the variability of the transbasin geostrophic transport attributed to eddies and Rossby waves is estimated to be 2.6 Sv). The role of eddies in generating low-frequency AMOC variability is still under debate, with some eddy-resolving models (e.g., Hirschi et al., 2013; Thomas & Zhai, 2013) suggesting that the eddy contribution to AMOC variability is significant compared to the seasonal cycle at some latitudes.

Theory continues to play a role in refining and improving our strategies for observing the AMOC. For example, Hughes et al. (2013) demonstrate how the geostrophic portion of the meridional overturning circulation can be monitored using only boundary measurements, while others (e.g., Duchez et al., 2014; Frajka-Williams, 2015) lean on established theory such as Sverdrup balance to develop metrics for the overturning circulation that may allow us to reconstruct its variability over a longer time period. Hughes et al. (2018) show that mesoscale energy is suppressed in bottom pressure on the continental slope, such that variations in pressure along boundaries are coherent over large distances, with likely implications for the coherence of the AMOC.

Several theoretical and simple idealized process models have helped interpret the variability in the observed AMOC time series at the RAPID-MOCHA array. Anything which alters the density or pressure on the eastern or western boundary of the North Atlantic, by definition, influences the AMOC, and there has been a growing realization over the last decade that winds may play a large role beyond that due to the local Ekman transport. The significant seasonal cycle in AMOC transport at 26°N appears to be governed by eastern boundary pressure anomalies. Zhao and Johns (2014a) force a simple linear Rossby wave model with the climatological seasonal cycle in wind stress to show that variations in the thermocline depth forced by wind stress curl anomalies on the eastern side of the RAPID-MOCHA array have a dominant effect on the sea-

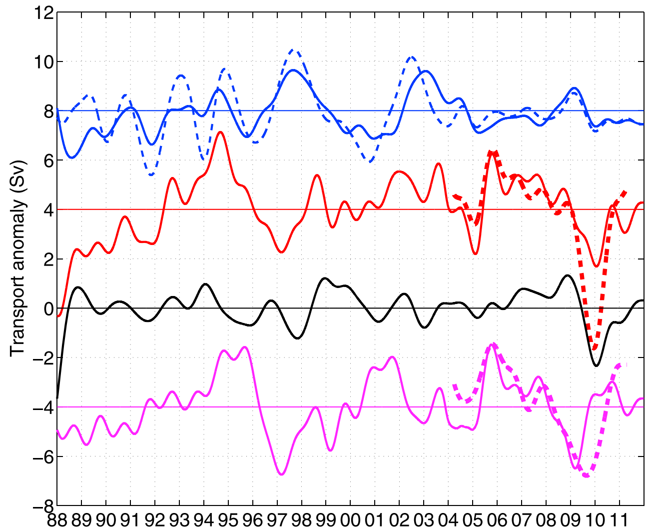

Figure 9. Low-pass filtered time series of variability in the Atlantic Meridional Overturning Circulation (red), Florida Current transport (blue), Ekman transport (black), and upper mid-ocean geostrophic transport (magenta) at 26°N, from Zhao and Johns (2014b). The dashed lines show in situ observations made by the RAPID-MOCHA array between April 2004 and April 2011. The solid lines show the results of a linear two-layer model forced only by observed wind stress. For better visualization, the mean level of the Florida Current, Atlantic Meridional Overturning Circulation, and upper mid-ocean time series has been shifted by 4 Sv each.

sonal cycle of the interior geostrophic transport, consistent with the observations reported in Chidichimo et al. (2010).

However, Yang (2015) insert walls in a two-layer Atlantic model to show that while eastern boundary pressure anomalies are a key contributor to the seasonal cycle in meridional geostrophic transport, which in turn is the largest contributor to the AMOC seasonal cycle at 26°N, the pressure at the eastern boundary is *not* forced mainly by the local wind stress curl at the African coast. Rather, it is the result of basin-wide adjustments to local and remote wind forcing. In their model, Yang (2015) see a meridional redistribution of the water in each layer between the subtropical and subpolar gyres in response to seasonal variability in Ekman pumping, and it is this which dominates the AMOC seasonal cycle.

It is becoming clear that changes in the geostrophic contribution to the AMOC often involve a complex oceanic adjustment to the wind forcing. Zhao and Johns (2014b) show that most of the observed interannual variability in overturning and its components at 26°N can be reproduced in a linear, two-layer model forced only by winds (Figure 9).

9. Generation and Propagation of Wider AMOC Variability

Progress has also been made in the last decade on understanding the role of surface buoyancy fluxes on longer timescales. For example, Zanna et al. (2011) solve a generalized eigenvalue problem in an idealized single-basin ocean GCM to identify the optimal three-dimensional structure of temperature and salinity perturbations that amplify AMOC anomalies. Their study makes it clear that linearized ocean dynamics can give rise to enhanced AMOC variability if overflows, eddies, or deep convection can excite *deep* high-latitude density anomalies in the northern part of the basin, and these impact the AMOC 7.5 years later via thermal Rossby modes (te Raa & Dijkstra, 2002) and boundary waves (Marshall & Johnson, 2013).

Reduced-gravity, 1.5 layer models that allow for zonal, but not vertical, structure in the overturning circulation and stratification are also highlighting the roles of boundary propagation and interior Rossby waves in the adjustment of the AMOC to a change in forcing. Samelson (2011) derives time-dependent analytical solutions for the warm branch of the AMOC in a single-basin plus reentrant channel which includes Southern Ocean Ekman and eddy volume fluxes, as well as diabatic processes to the north. The adjustment timescale is multidecadal to centennial and proceeds via the boundary and interior adjustment mechanisms identified in Johnson and Marshall (2002a, 2002b), with the eastern boundary upper layer thickness proving to be the key variable. Nieves and Spall (2018) present similar reduced-gravity theory for variability in the deep branch of the overturning circulation in a single basin, explicitly assuming that the eastern boundary layer thickness is equivalent to the domain-averaged layer thickness (valid in the low frequency limit), consistent with flat isopycnals on the eastern boundary and Rossby wave propagation into the interior.

Reduced-gravity models such as these provide a simple analytical framework for exploring the different characteristics of AMOC variability generated by wind forcing and surface buoyancy fluxes, and the combined influence of the two. Zhai et al. (2014) build on Johnson and Marshall (2002a) to propose an analytical theory for the interaction of wind stress and buoyancy fluxes associated with the North Atlantic Oscillation in generating AMOC variability. They construct a volume budget for the upper branch of the AMOC assuming fast boundary wave propagation and westward Rossby wave propagation in the ocean interior (these adjustment mechanisms are illustrated in Figure 1). The effect of stochastic wind forcing over the subpolar basin is integrated along Rossby wave characteristics and consequently results in low-frequency AMOC variability in the rest of the basin. In contrast, North Atlantic Oscillation-related surface buoyancy fluxes excite stochastic variability directly on the western boundary, and this is subsequently communicated into the ocean interior.

Surface buoyancy fluxes and wind stress are intrinsically coupled, and as Williams et al. (2014) point out, variations in the wind over the North Atlantic naturally lead to opposing signs of thermal anomalies, Ekman heat convergence, and (via thermal wind) MOC-minus-Ekman volume transport in the subtropical and subpolar parts of the basin.

Boundary waves are clearly important in determining AMOC variability (also see the adjoint sensitivity studies in Czeschel et al. (2010) and Pillar et al. (2016)). However, the meridional propagation of density and AMOC anomalies in climate models occurs on a wide range of timescales. Marshall and Johnson (2013) provide a possible explanation for some of the differences, solving analytically for the propagation of waves along eastern and western boundaries in a reduced-gravity model. They show that, even in this very simple model, on timescales longer than 1–2 months, boundary waves are not Kelvin waves but rather short and long Rossby waves that satisfy a lateral boundary condition. These Rossby waves propagate cyclonically along coastlines with an along-boundary propagation speed of cL_d/δ, where c is the gravity wave speed, L_d is the Rossby deformation radius, and δ is the frictional boundary current width. Although the waves are likely more complicated with sloping bottom topography rather than vertical sidewalls (Wise et al., 2018; Hughes et al., 2019), the results of Marshall and Johnson (2013) may explain some of the spread in the meridional propagation speed of AMOC anomalies between different models.

10. Alternative AMOC Frameworks

The AMOC transport can be diagnosed using depth and latitude (at fixed depth) as coordinates, or density and latitude (at fixed neutral or potential density). The mean overturning streamfunction calculated from the Estimating the Circulation and Climate of the Ocean (ECCO) state estimate is shown in Figure 2. In density coordinates, the time average and zonal integral characterizing the climatological AMOC must be performed at constant density, requiring synoptic velocity and density measurements. The density coordinate framework highlights the transformation between waters of different density classes and is thus more representative of the transport of tracers (Andrews & McIntyre, 1978). The difference between the two frameworks is especially important in high latitudes where the isopycnals slope most dramatically. In the Antarctic circumpolar region the main difference between the transport in depth versus density coordinates is due to the large contribution by mesoscale eddies and standing waves to the transport of density (e.g., Doos & Webb, 1994). In the high latitudes of the North Atlantic the difference arises from the (horizontal) transport effected by the subpolar gyre. The first continuous observations of the overturning in density space at high northern latitudes, made over 2014–2016 by the Overturning in the Subpolar North Atlantic Programme (Lozier et al., 2017), suggest that most of the water mass transformation and its variability occur to the east of Greenland (Lozier et al., 2019).

The water mass transformation framework (Walin, 1982) formulates mass budgets for ocean volumes bounded by scalar surfaces (e.g., isopycnals) and relates ocean circulation to processes which result in a change in the position of these surfaces (see Groeskamp et al., 2019, for a comprehensive review). In an AMOC context, it has been applied, for example, to understand the link between surface forcing, diapycnal mixing, and advection of water masses in the North Atlantic (e.g., Speer & Tziperman, 1992), interannual variations of the AMOC (Marsh, 2000), and the formation of dense Nordic Seas waters (Isachsen et al., 2007),

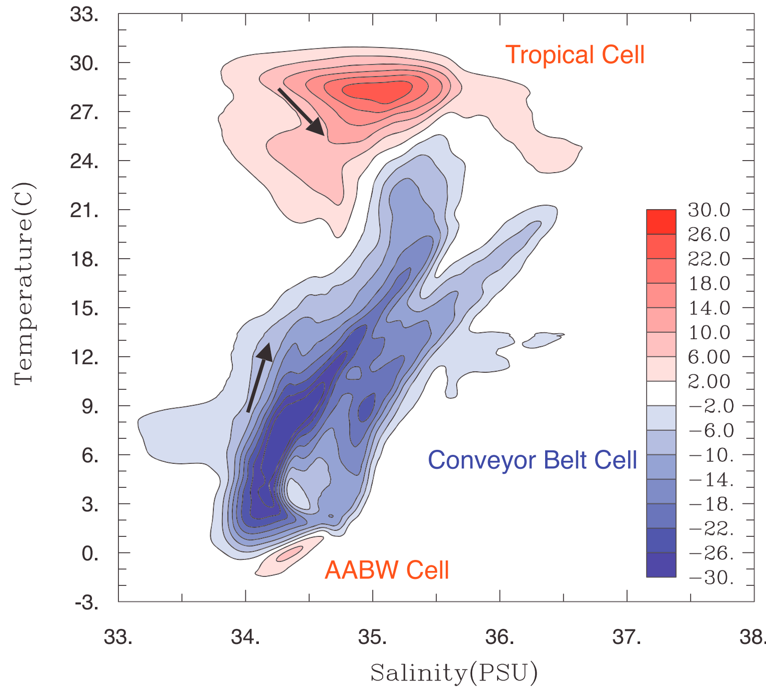

Figure 10. The thermohaline streamfunction from Doos et al. (2012). The deep overturning circulation, indicated in red in Figure 2, is shown here by the blue clockwise circulation, which corresponds to water mass conversion from warm water to more saline, then to cold, then to less saline, and back to warm water. The tropical and Antarctic Bottom Water (AABW) cells flow in the opposite direction. Units are in Sverdrups (Sv) with a volume transport of 4 Sv between contours.

as well as to characterize the AMOC in models (e.g., Marsh et al., 2005). Building on Speer (1993), who decomposed the circulation in buoyancy coordinates into its separate contributions from salinity and potential temperature, Doos et al. (2012), Zika et al. (2012), and Groeskamp et al. (2014) define a streamfunction in thermohaline coordinates (θ and S; see Figure 10) which provides insight into the AMOC's role in global heat and freshwater transports and allows for the study of processes directly affecting the thermohaline circulation (e.g., Hieronymus et al., 2014). Considering the water mass transformation only in θ coordinates allowed Evans et al. (2017) to reconcile observed wind-driven changes in the AMOC at 26°N with observed changes in water masses. Recent efforts to combine the overturning circulation in thermohaline coordinates with a box-inverse method (Mackay et al., 2018) are adding value to current observational campaigns.

11. Concluding Discussion

Several key themes and questions emerge that have influenced those constructing theoretical models of the AMOC over the last decade. Many of these are illustrated in Figure 1.

- It is now clear that the wind, in both hemispheres, plays a prominent role in setting the mean AMOC strength and determining its variability. This includes the interaction between the wind stress and the surface buoyancy distribution, because wind-driven upwelling, and the vertical flux of buoyancy associated with eddies and gyres, brings water to the surface where its density can be transformed by buoyancy fluxes.
- Simplified models are moving from two-dimensional zonally averaged representations to geometries that capture the fundamentally three-dimensional aspects of the circulation. This includes a distinction between the western boundary and the basin interior and a focus on the circulation between multiple basins.
- The degree to which water mass transformation in the ocean interior is important, or whether the circulation in the Atlantic sector is essentially adiabatic, remains an open question.
- Multiple lines of evidence suggest that eddies at high latitudes in both hemispheres are essential to the dynamics of the AMOC.

Some theoretical models have moved beyond a treatment of ocean dynamics to explore the AMOC's role in the coupled climate system. For example, Marshall and Zanna (2014) explore the processes, including the AMOC, that determine ocean heat uptake in a multilayer version of the Gnanadesikan (1999) model, using it to interpret the results of a climate change experiment in a coupled climate model. Zhai et al. (2011) explore the links between AMOC variability and ocean heat content in the thermocline, showing that low frequency variability in the AMOC is required to impact the ocean's heat content at middle and high latitudes. Meanwhile, other authors are beginning to develop theoretical models that include a representation of the biogeochemical processes that interact with the ocean's overturning circulation to determine the impact of AMOC variability on climate (e.g., Goodwin, 2012).

A major recent theoretical advance, whose implications for our understanding of the AMOC are not yet clear, is our changing paradigm concerning the link between ocean turbulent mixing and water mass transformation. It is classically assumed (Munk, 1966; Munk & Wunsch, 1998; Wunsch & Ferrari, 2004) that a vertical advective-diffusive balance is appropriate in the interior, with dense waters transformed into lighter ones as a result of diffusive fluxes. However, turbulent mixing due to breaking internal waves is stronger over rough bottom topography and decays away from the bottom, which means that a parcel of water in the stratified interior mixes more with the water below it than that above, such that mixing in fact converts light waters into denser ones (de Lavergne et al., 2016; Ferrari et al., 2016). Abyssal waters can therefore only rise toward the surface in narrow turbulent boundary layers that develop along continental margins and abyssal ridges. The abyssal meridional overturning circulation is the small residual of diapycnal sinking in the stratified interior and diapycnal upwelling in boundary layers.

The increase in mixing with depth in the stratified interior is at least partially offset by the decrease in mixing area with depth (de Lavergne et al., 2017). In any case, this new conceptual picture is most likely to be relevant below 2,000 m where small-scale mixing driven by internal wave breaking provides the dominant source of energy to lift water masses, and it is as yet unclear how important it is going to be for the AMOC. There are other important implications, however, such as for the horizontal pathways of abyssal ocean flow, the residence time and distribution of tracers, and the relevance of idealized numerical simulations with vertical sidewalls, since it is now clear that sloping boundaries are crucial in allowing ocean basins to remain stratified.

Theoretical and conceptual models continue to play a vital and revealing role in articulating our current understanding of the overturning circulation and helping us to frame first-order questions that remain open about AMOC dynamics. We close by outlining some of the remaining challenges which we anticipate theory will make a significant contribution to solving over the next decade.

- A quantitative understanding of the processes which determine the mean strength of the residual overturning circulation is still lacking. In the Southern Ocean the Eulerian mean and eddy-driven overturning cells appear to be largely powered by local winds, but the interplay between these and the surface buoyancy fluxes and wind stress in the North Atlantic is still being explored, as is the diapycnal mixing along the whole path of the overturning.
- We know from a host of model studies that AMOC variability depends on the background state. Theoretical models have the potential to help us understand this relationship and make sense of the bewildering variety of AMOC variability that is exhibited by state-of-the-art ocean and climate models.
- The continued development of simple models of the coupled ocean-atmosphere system, building on studies such as Lucarini and Stone (2005), will better describe the thermal air-sea interaction, with important repercussions for the representation of AMOC stability.
- Insight from theoretical studies will help us to prioritize the most critical, often subgrid scale, processes requiring improved representation in numerical ocean circulation models and to develop the appropriate physical parameterizations. For example, it is now clear that submesoscale processes such as mixed layer instabilities and eddies can play an important role in restratifying the water column and balancing surface heat fluxes, with dramatic impacts on mixed layer depths (e.g., Boccaletti et al., 2007). The development of a mixed layer eddy parameterization therefore has the potential to significantly improve the representation of the AMOC in models.

By their very nature, theoretical advances are difficult to foresee! However, several of the developments over the last decade have been catalyzed by recent observations such as those made by the RAPID-MOCHA array at 26°N (e.g., our understanding of the mechanisms via which wind forcing dominates the AMOC seasonal cycle, and of the role of eddies and boundary propagation). The unprecedented amount of data currently being collected by existing and new arrays throughout the Atlantic (e.g., RAPID-MOCHA (McCarthy et al., 2015), Overturning in the Subpolar North Atlantic Programme (OSNAP, Lozier et al., 2017), SAMOC at 34°S (Meinen et al., 2018), MOVE at 16°N (Send et al., 2011), etc. See Cunningham & coauthors, 2019, for a summary) together with long-term global ocean observing systems such as Argo (http://www.argo.ucsd.edu) and satellite altimetry (https://www.aviso.altimetry.fr/en/) mean that there has never been such an exciting time to be thinking about the Atlantic Ocean. This, along with the substantial contributions made by theoretical models to date, provides powerful motivation for equipping the next generation of physical oceanographers with the skills required to use them as part of their toolbox. We can expect that such models will play a major role in understanding the observations collected over the coming decade, as well as in formulating hypotheses for the community to test using these observations alongside numerical ocean and climate models.

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