Interannual Variability of the Global Meridional Overturning Circulation Dominated by Pacific Variability

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ABSTRACT

The most prominent feature of the time-mean global meridional overturning circulation (MOC) is the Atlantic MOC (AMOC). However, interannual variability of the global MOC is shown here to be dominated by Pacific MOC (PMOC) variability over the full depth of the ocean at most latitudes. This dominance of interannual PMOC variability is robust across modern climate models and an observational state estimate. PMOC interannual variability has large-scale organization, its most prominent feature being a cross-equatorial cell spanning the tropics. Idealized experiments show that this variability is almost entirely wind driven. Interannual anomalies of zonal mean zonal wind stress produce zonally integrated Ekman transport anomalies that are larger in the Pacific Ocean than in the Atlantic Ocean, simply because the Pacific is wider than the Atlantic at most latitudes. This contrast in Ekman transport variability implies greater variability in the near-surface branch of the PMOC when compared with the near-surface branch of the AMOC. These near-surface variations in turn drive compensating flow anomalies below the Ekman layer. Because the baroclinic adjustment time is longer than a year at most latitudes, these compensating flow anomalies have baroclinic structure spanning the full depth of the ocean. Additional analysis reveals that interannual PMOC variations are the dominant contribution to interannual variations of the global meridional heat transport. There is also evidence of interaction between interannual PMOC variability and El Niño–Southern Oscillation.

1. Introduction

Understanding Earth's climate requires understanding how motions in the atmosphere and ocean redistribute the energy provided by the sun. The ocean generates approximately one-quarter of the equator-to-pole energy transport, and the ocean contribution is even greater in the tropics (e.g., Held 2001; Trenberth and Caron 2001; Czaja and Marshall 2006). This energy

transport is accomplished through a combination of the horizontal gyre circulations and the meridional overturning circulation (MOC).

The annual mean climatology of the global MOC is shown in Fig. 1a, computed from the Estimating the Circulation and Climate of the Ocean (ECCO) state estimate (Forget et al. 2015). (Additional details about ECCO and the MOC computation are provided in section 2.) Time-mean global MOC consists of a few prominent, well-known features: shallow overturning cells near the equator in the Indian–Pacific basin (Fig. 1b), the Atlantic MOC (AMOC) occupying the upper half of the ocean (Fig. 1c), and abyssal overturning in the deep Indian–Pacific (Fig. 1b).

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FIG. 1. (left) Annual mean and (right) interannual standard deviation of the MOC streamfunction ψ for the (a),(d) global ocean, (b),(e) Indian-Pacific basin, and (c),(f) Atlantic Ocean, calculated from the ECCO state estimate. In (a)–(c), positive values indicate clockwise motion and negative values indicate counterclockwise motion. The shading intervals are 2.6 Sv for (a)–(c) and 0.4 Sv for (d)–(f). Grid cells below the ocean floor are shaded gray.

Much of the past discussion of the MOC has focused on these time-mean features and their low-frequency variability. For example, the AMOC is believed to play a role in multidecadal climate variations in the North Atlantic Ocean (e.g., [Delworth et al. 1993](#); [Knight et al. 2005](#); [Tandon and Kushner 2015](#)), although this has been the subject of recent debate (e.g., [Clement et al. 2015](#)). The shallow overturning in the tropical Pacific Ocean influences carbon dioxide storage and marine ecosystems (e.g., [McPhaden and Zhang 2002](#); [Zhang and McPhaden 2006](#)). The abyssal circulation in the Pacific is thought to be influenced by bottom topography and

ice cover in the Southern Ocean (e.g., [Ferrari et al. 2014, 2016](#)).

In this study, we show that the MOC exhibits substantial interannual variability at all depths of the ocean. This variability has spatial structure that does not resemble the time-mean global MOC: interannual MOC variability is dominated by variability in the Indian-Pacific MOC (PMOC) at most latitudes over the full depth of the ocean, including depths at which the time-mean PMOC is essentially zero. Below, we document this interannual PMOC variability in ECCO and modern climate models ([section 2](#)), examine its spatial and temporal

organization (section 3), provide a physical basis for expecting this relatively strong PMOC variability (sections 4 and 5), and highlight consequences for variability in meridional heat transport (MHT) (section 6).

2. Interannual PMOC variability in ECCO and modern climate models

For most of our analysis, we use ECCO, version 4 release 2, interpolated to a 0.5° horizontal grid with 50 vertical levels, covering the period 1992–2011 (Forget et al. 2015, 2016). This dataset is generated by an ocean model forced by atmospheric fields derived from ERA-Interim reanalysis (Dee et al. 2011) with additional constraints to sea surface temperature (SST) observations from the National Oceanic and Atmospheric Administration (Reynolds et al. 2002), satellite altimetry (Scharroo et al. 2004), the global network of Argo floats (Argo 2000), and other in situ ocean measurements.

In this paper, we use “MOC” to refer to the MOC mass transport streamfunction expressed in volume units ψ , defined as

$$\psi(y, z, t) = -\frac{1}{\rho_0} \int_{A_{xz}(y, z)} \rho(x, y, z', t) v(x, y, z', t) dA, \quad (1)$$

where ρ is the density of water, ρ_0 is a constant reference density of water, v is the meridional velocity in the ocean, x is longitude, y is latitude, z is depth (with $z = 0$ at a reference height for the ocean surface and positive values below that level), z' is the dummy depth, t is time, A_{xz} is the cross-sectional area of the relevant basin in the x – z plane below depth z , $dA = dx dz'$, and integration is performed in the positive x and z directions. Positive values of ψ indicate clockwise motion and negative values indicate counterclockwise motion when viewed from the east, which we assume to be the case hereinafter.

For models that use the Boussinesq approximation, the contribution of density variations to the mass continuity equation become negligible. In this case, we can assume $\rho \approx \rho_0$ in (1), which then reduces to the volume streamfunction,

$$\psi(y, z, t) = - \int_{A_{xz}(y, z)} v(x, y, z', t) dA. \quad (2)$$

The model used in ECCO is Boussinesq, and so we use (2) to compute ψ from ECCO output. For these data, v is obtained from the sum of the monthly mean resolved velocity (variable “NVELMASS”) and the monthly mean parameterized bolus velocity (variable “NVELSTAR”).

None of our conclusions are affected if bolus velocity is excluded from the calculations. Contours of ψ (e.g., Fig. 1, left column) are tangent to the zonally integrated flow, and time variations in the MOC are possible only if there are also time variations in v . This connection between MOC and v will be crucial to our dynamical interpretation of MOC variations.

In the Atlantic Ocean, MOC is synonymous with the single overturning cell that (in the time mean) occupies the upper half of the basin. In the Indian–Pacific basin, however, there are multiple cells, and thus confusion might arise. In this study, we use MOC to refer specifically to the MOC streamfunction ψ , regardless of any large-scale organization. When we are interested in a particular large-scale MOC feature, that will be made clear in context.

The ECCO-derived standard deviation of the annual mean global MOC is shown in Fig. 1d. Hereinafter, we refer to this quantity simply as the “interannual standard deviation” (ISTD) of the MOC, and none of our conclusions are affected if we instead compute the standard deviation of the high-pass-filtered annual mean MOC with a cutoff frequency of $(5 \text{ yr})^{-1}$ (not shown). Figure 1d shows substantial (1.5 – 3 Sv ; $1 \text{ Sv} \equiv 10^6 \text{ m}^3 \text{ s}^{-1}$) ISTD spanning the full depth of the ocean. The MOC variability in the deep ocean (by which we mean below the thermocline at ~ 500 -m depth) is of particular interest. Most of this deep MOC variability reflects variability in the Indian–Pacific basin (cf. Figs. 1e,f). We reach the same conclusions if we exclude the Indian Ocean from the analysis (not shown). We have combined the Indian and Pacific Oceans together to facilitate comparison with climate models that typically combine these basins when computing the MOC streamfunction. In the tropics, the PMOC ISTD exceeds the AMOC ISTD by approximately a factor of 3.

The dominance of interannual PMOC variability is also apparent in modern fully coupled climate models. This claim is substantiated in Fig. 2, which shows ISTD of PMOC and AMOC from preindustrial control simulations of eight climate models participating in phase 5 of the Coupled Model Intercomparison Project (CMIP5) (Taylor et al. 2012). Our analysis includes all of the models that archived at least 499 years of the mass streamfunction for the preindustrial control scenario. Analyzing such long runs helps to ensure that key processes found in the ECCO data are not artifacts of its relatively short 20-yr record. For the CMIP5 models, ψ was computed from the monthly mean mass streamfunction [variable “msftmyz,” which is specified to include bolus advection (Griffies et al. 2009)] divided by $\rho_0 = 1035 \text{ kg m}^{-3}$ to convert to volume transport units.

FIG. 2. Interannual standard deviation of the (left),(right center) PMOC and (left center),(right) AMOC for eight models participating in CMIP5. The shading interval is 0.4 Sv, and grid cells below each model's ocean floor are shaded gray.

This study includes additional analysis and idealized simulations using the Canadian Earth System Model, version 2 (CanESM2) (Arora et al. 2011), which has atmosphere, ocean, land, sea ice, and carbon cycle components. The atmospheric component of CanESM2 is a spectral model run with T63 triangular truncation and 35 vertical levels. The ocean component has 40 vertical levels with horizontal resolution of 1.41° longitude by 0.94° latitude. For CanESM2, the mass streamfunction in latitude–potential density coordinates (variable “msftmrhoz”) was also archived for the CMIP5 pre-industrial control scenario. In these coordinates, the dominance of interannual PMOC variability was clearly evident over the full depth of the ocean (not shown), indicating that our findings are not sensitive to the vertical coordinate used to compute the MOC streamfunction.

Thus, the dominance of interannual PMOC variability over interannual AMOC variability is robust across modern climate models and a modern observational product. Past discussions of MOC variability have mostly focused on the decadal and multidecadal variability of

the AMOC (e.g., Delworth et al. 1993; Knight et al. 2005; Tandon and Kushner 2015) and variability of the shallow overturning circulation in the Pacific (e.g., McPhaden and Zhang 2002; Zhang and McPhaden 2006; McPhaden and Zhang 2018). Jayne and Marotzke (2001) documented seasonal variability of the MOC over the full depth of the ocean, but this does not necessarily imply significant interannual variability of deep overturning. To our knowledge, the interannual variability of deep overturning, and its predominantly Pacific origin, has not been documented or explained.

3. Spatial and temporal structure of PMOC variability

To give a sense of the dominant time scales, Fig. 3 shows spectra of PMOC and AMOC computed from a 499-yr control run of CanESM2 using the Thomson (1982) multitaper method. (Specifically, we used MATLAB function “pmtm” with time-bandwidth product of 2. ECCO is ill suited for such a calculation because of its relatively short record.) The strongest PMOC variability (Fig. 3a)

FIG. 3. Spectra of annual mean (a) PMOC and (b) AMOC computed at 2000-m depth and at each latitude from a 499-yr control simulation of CanESM2. The shading scale is logarithmic. See the text for additional details.

is at time scales shorter than 10 years at most latitudes. Equatorward of 40° latitude, the strongest AMOC variability (Fig. 3b) is in the 4–10-yr band, and at latitudes north of 40°N , the strongest AMOC variability is on time scales greater than 10 years. PMOC variability is clearly weaker than AMOC variability on time scales greater than 10 years and stronger than AMOC variability on time scales shorter than 4 years. We reach the same conclusions if we compute spectra at other depths below 500 m (not shown). This time scale dependence is also clear when examining depth versus latitude plots of PMOC and AMOC after applying filtering of various time scales (not shown).

If wind stress is the dominant driver of interannual PMOC variability (a matter we address in detail below), then we would expect to see a spectral peak around 4 years near the equator, corresponding to the time scale of El Niño–Southern Oscillation (ENSO). However, the equatorial wind stress in CanESM2 has an unrealistically flat spectrum for time scales of 4 years and less (not shown), and accordingly the PMOC spectrum (Fig. 3a) also lacks a 4-yr spectral peak at the equator. Although some models show a clearer spectral peak at 4 years (e.g., CCSM4, not shown), they agree with CanESM2 in that PMOC variability dominates at interannual time scales and AMOC variability dominates at multidecadal time scales.

This interannual PMOC variability is not just noise: it has large-scale spatial structure. Figure 4 shows ECCO-derived annual mean PMOC anomalies for eight successive years. These plots reveal anomalous overturning cells spanning 20° – 40° in latitude over the full depth of the ocean. In most years (1995, 1996, 1998, 1999, 2001, and 2002) there is anomalous cross-equatorial overturning in the deep ocean, but in 1997 and 2000 the anomalous deep overturning has a dipole structure that is more equatorially antisymmetric. In 1995 and 1999 there are deep overturning anomalies poleward of 20° that are opposite in sign to the cross-equatorial

anomalies, but in 1998 and 2002 the cross-equatorial anomalies coincide with larger scale anomalies that extend into the midlatitudes. In some years (1997, 1999, and 2002), there are dipole anomalies in the upper ocean (by which we mean above ~ 500 m, i.e., within and above the thermocline) with a sign change at the equator, suggesting changes in the strength of the subtropical cells (cf. Farneti et al. 2014). But in other years (1995, 1996, 1997, 2000, and 2001), there is anomalous cross-equatorial overturning in the upper ocean. Interestingly, there is no clear association between the structure of anomalous overturning in the upper ocean and anomalous overturning in the deep ocean: cross-equatorial anomalies in the deep ocean do not consistently correspond with cross-equatorial anomalies in the upper ocean.

Empirical orthogonal function (EOF) analysis of annual mean PMOC anomalies reveals a prominent cross-equatorial cell spanning 18°S – 20°N below 500 m, accounting for 51% of the variability (Fig. 5a). Such a clear cross-equatorial cell is also apparent in the annual mean anomalies (region within dashed lines in Fig. 4, especially during 1995, 1996, 1999, and 2002). Associated with the anomalous cross-equatorial PMOC are interannual sign changes in ψ at 1000–3000 m in the tropics (not shown), a region where the mean PMOC is close to zero (cf. Fig. 1b). The second EOF (Fig. 5b, accounting for 27% of the variability) is a dipole anomaly with structure similar to the anomalous overturning in 2000 (Fig. 4). We obtained similar EOFs and fractions explained from the longer CMIP5 control runs (not shown).

More work is needed to understand the mechanisms that generate the anomalous cross-equatorial overturning associated with PMOC EOF1 (Fig. 5a). One possibility is that anomalous northward transport across the equator is linked to Ekman transport changes generated by anomalous eastward wind stress south of the equator and anomalous westward wind stress north of

FIG. 4. ECCO-derived annual mean MOC anomalies in the Indian-Pacific basin for eight successive years beginning in (top left) 1995 and ending in (bottom right) 2002. Dashed lines outline the domain from 18°S to 20°N below 500 m, over which the EOFs in Fig. 5 (below) are computed. The shading interval is 0.6 Sv, and grid cells below the ocean floor are shaded gray.

the equator, as is the case for the MOC seasonal cycle (e.g., Jayne and Marotzke 2001; Green and Marshall 2017) and the cross-equatorial cell in the Indian Ocean (Miyama et al. 2003). However, regression of zonal wind stress (ZWS) onto the PMOC principal component 1 (not shown) does not reveal an equatorially antisymmetric dipole anomaly of ZWS. Rather, the associated ZWS anomaly is eastward at most latitudes. We will show below that this deep PMOC variability is still

ultimately wind driven, but our analysis suggests that the anomalous cross-equatorial transport cannot be understood as a simple Ekman response to local wind forcing. Rather, there are additional nonlocal effects of wind forcing that require further investigation.

In summary, interannual PMOC variability has clear large-scale structure, dominated by an anomalous cross-equatorial cell that reverses direction approximately every year. This result makes clear that interannual

FIG. 5. The (a) first and (b) second EOF of annual mean PMOC computed from ECCO over the domain 18°S–20°N below 500 m (marked by dashed lines in Fig. 4). Depths above 500 m have been excluded to focus on variations in deep overturning rather than shallow overturning. The shading interval is 0.3 Sv, and the percentage of variance explained is indicated in parentheses above each panel.

PMOC variability is distinct from a thermohaline circulation. The age of water in the deep North Pacific Ocean is known to be approximately 1000 years or older (England 1995; Gebbie and Huybers 2012). This water is much older than water in the deep North Atlantic, where buoyancy-driven downwelling is a regular occurrence. If significant buoyancy-driven downwelling were occurring in the Pacific Ocean on interannual time scales, then water in the deep Pacific would be much younger than it actually is. Thus, we can safely infer that interannual MOC variability in the deep Pacific is not buoyancy-driven. In the next section, we show that this interannual PMOC variability is wind driven instead.

4. The role of wind stress variability

To assess the role of wind stress variability in PMOC variability, we have performed idealized “partial coupling” experiments using CanESM2. These runs were performed over 1979–2014 under the same historical forcings used in the “historical” scenario of CMIP5 (Taylor et al. 2012). In these experiments, the wind stress transmitted to the ocean in particular latitude bands is replaced by the model’s 1979–2014 climatological seasonal cycle (hence suppressing interannual variability of wind stress), while the wind stress is freely evolving elsewhere. Additional details about this partial coupling approach can be found in Saenko et al. (2016). Figure 6a shows the PMOC ISTD for an experiment in which the interannual variability of wind stress is suppressed poleward of 15° latitude. The ISTD between 15°S and 15°N is essentially identical to that of the fully coupled CanESM2 control run (Fig. 2). Poleward of 15°

latitude, PMOC ISTD in Fig. 6a is greatly diminished compared to the fully coupled CanESM2.

Complementary to this experiment, we have also performed an experiment in which the interannual variability of wind stress is suppressed between 15°S and 15°N and is freely evolving elsewhere. In this case, the PMOC ISTD is mostly suppressed between 15°S and 15°N (Fig. 6b), although the amount of ISTD that survives is greater than the ISTD that survives in the extratropics when interannual wind stress variability is suppressed there (Fig. 6a). Poleward of 15° latitude, PMOC ISTD in Fig. 6b reproduces that of the fully coupled CanESM2.

We have also performed an experiment in which interannual variability of wind stress is suppressed everywhere. In this case, the PMOC ISTD is approximately 0.3 Sv or less everywhere (not shown), suggesting that oceanic internal variability is not the dominant driver of interannual PMOC variability. Taken together, these idealized experiments suggest that interannual wind stress variability is crucial for generating interannual PMOC variability. Furthermore, this wind stress influence is mostly confined to the latitudes where the wind stress is varying interannually. While these experiments clearly show the role of wind stress variability as a proximal driver of interannual PMOC variability, these experiments do not address whether that wind stress variability is generated through atmosphere–ocean coupling or atmospheric internal variability.

In any discussion of interannual variability, it is natural to think of ENSO, which is the dominant driver of interannual variations in global patterns of temperature

FIG. 6. The interannual standard deviation of PMOC in idealized simulations of CanESM2. Shown are simulations in which (a) between 15°S and 15°N surface wind stress is freely evolving and poleward of 15° latitude the wind stress transmitted to the ocean is a fixed seasonal cycle and (b) wind stress poleward of 15° latitude is freely evolving and between 15°S and 15°N the wind stress transmitted to the ocean is a fixed seasonal cycle. The shading interval is 0.4 Sv, and grid cells below the model's ocean floor are shaded gray.

and precipitation (Sarachik and Cane 2010). Indeed, we have found evidence of a connection between ENSO and interannual PMOC variability. Figure 7, shows the ECCO-derived lag correlation between the annual mean Niño-3.4 index (detrended SST anomalies averaged over 5°S–5°N, 120°–170°W) and an annual mean index of the PMOC. Here, the PMOC index is defined as the sum of the first two principal components associated with the EOFs shown in Fig. 5.

Figure 7 shows that Niño-3.4 is positively correlated at zero lag with the PMOC index. This means that positive anomalies of equatorial east Pacific SST are generally associated with anomalously clockwise circulation of the cross-equatorial PMOC cell, with anomalous northward transport above ~1000 m at the equator. This cross-equatorial transport contrasts with the anomalous equatorial convergence expected with equatorial SST warming (e.g., Gill 1980). Such anomalous convergence does indeed occur in the atmosphere, but not in the ocean. As mentioned above, the mechanisms responsible

FIG. 7. Pearson correlation coefficients between detrended annual mean Niño-3.4 anomalies and the PMOC index, computed from ECCO for a range of lag values. The PMOC index is the sum of the principal component time series associated with the first two EOFs shown in Fig. 5. The correlations that are statistically significant at the 95% level (based on a two-tailed *t* test) are indicated by the horizontal dashed lines. The effective temporal degrees of freedom were computed as in Bretherton et al. (1999).

for this anomalous cross-equatorial transport in the Pacific require further investigation.

Interestingly, there is also evidence of feedback between ENSO and PMOC variations on interannual time scales. Figure 7 shows a statistically significant negative correlation when Niño-3.4 leads the PMOC index by 2 years. These results motivate future work to understand the mechanisms responsible for this covariability and the implications for ENSO variability. For this study, the key point of Fig. 7 is that the PMOC is not simply an alternative index of ENSO: while there is a statistically significant simultaneous correlation (0.43), a majority of interannual PMOC variability cannot be explained by ENSO SST anomalies.

5. Dynamical interpretation

The importance of wind stress variability for interannual PMOC variability allows us to apply additional dynamical principles toward understanding PMOC variability. First, we focus on the ZWS, denoted τ_x , and its cross-basin integral, $\langle \tau_x \rangle$. Imagine applying a zonally uniform anomaly of τ_x spanning the Atlantic and Indian–Pacific basins. Then, at most latitudes, the fact that the Indian–Pacific basin is wider than the Atlantic Ocean means that the anomalous $\langle \tau_x \rangle$ over the Indian–Pacific basin is larger than the anomalous $\langle \tau_x \rangle$ over the Atlantic. Over a time series of such anomalies, the ISTD of $\langle \tau_x \rangle$, which we denote $\sigma\langle \tau_x \rangle$, would be larger over the Indian–Pacific basin than over the Atlantic.

This seemingly simplistic thought experiment explains surprisingly well the Pacific–Atlantic contrast in variability of basin-integrated ZWS. Figure 8a shows that, except for latitudes north of 40°N, the ISTD of the zonally averaged τ_x over the Indian–Pacific basin (red) is similar to that over the Atlantic Ocean (black). Thus, in the tropics, $\sigma\langle\tau_x\rangle$ over the Indian–Pacific is larger than $\sigma\langle\tau_x\rangle$ over the Atlantic (Fig. 8b). South of 10°N, Indian–Pacific $\sigma\langle\tau_x\rangle$ exceeds Atlantic $\sigma\langle\tau_x\rangle$ by a factor of 3–4, reflecting the difference in basin widths over these latitudes. We have examined Hovmöller plots of τ_x anomalies over the Pacific and Atlantic (not shown), and the spatial structure of these anomalies span enough of each basin to justify our simplified cross-basin perspective.

We expect such ZWS variations to result in Ekman transport variations, that is, variations in the meridional flow in the top ~100 m of the ocean. This expectation follows from the well-known expression for Ekman transport,

$$\int_0^h v_E(x, y, z, t) dz = -\frac{\tau_x(x, y, t)}{f\rho_0}, \quad (3)$$

where v_E is the Ekman-driven meridional velocity, f is the Coriolis parameter, and h is the depth of the Ekman layer. The volume transport in the Ekman layer V_E is then

$$V_E = \int_{A_E} v_E dA = -\frac{\langle\tau_x\rangle}{f\rho_0}, \quad (4)$$

where A_E is the cross-sectional area of the Ekman layer in the x – z plane. Then the ISTD of V_E is

$$\sigma(V_E) = \frac{\sigma\langle\tau_x\rangle}{f\rho_0}. \quad (5)$$

We can think of $\sigma(V_E)$ as an approximation for the ISTD of ψ at the bottom of the Ekman layer, $\sigma(\psi_E)$, allowing for the fact that ψ may be slightly different when integrating from the ocean surface to h , rather than integrating from h to the ocean floor [compare the middle of (4) with (2)].

Figure 8c (thick lines) shows $\sigma(V_E)$ computed using (5) for $\rho_0 = 1029 \text{ kg m}^{-3}$. (The results were not sensitive to the precise choice of ρ_0 within a realistic range.) For comparison, the thin lines in Fig. 8c show $\sigma(\psi_E)$ computed at 100-m depth (the approximate bottom of the Ekman layer). These values correspond well to $\sigma(V_E)$, suggesting that most of $\sigma(\psi_E)$ is indeed Ekman-driven. The agreement breaks down within 2° of the equator, where f vanishes. At these latitudes, the near-surface volume transport variations are driven directly by sea

FIG. 8. (a) The interannual standard deviation of the cross-basin average of ZWS over the Indian–Pacific basin (red) and the Atlantic Ocean (black), computed from ECCO. (b) As in (a), but for the cross-basin integral of ZWS, $\sigma\langle\tau_x\rangle$. (c) Thick lines show the ISTD of the volume transport implied by the ZWS variations in (b), calculated using (5). For comparison, the thin lines show the ISTD of the MOC streamfunction at 100-m depth.

surface height (SSH) variations generated off of the equator (not shown).

Thus, because of Ekman transport, the fact that $\sigma\langle\tau_x\rangle$ is higher in the Pacific than in the Atlantic implies that $\sigma(\psi_E)$ is also higher in the Pacific than in the Atlantic. Such Ekman transport anomalies would drive compensating flow anomalies below the Ekman layer (e.g., Pedlosky 1968; Jayne and Marotzke 2001), but why do these MOC anomalies extend to the deep ocean? On long enough time scales, the wind-driven flows are mostly confined to depths within and above the thermocline, an equilibrium state in which the deep ocean is essentially motionless. However, because we are considering

interannual variability (rather than a long term average), the deep ocean is not in equilibrium, and we have to consider the ocean adjustment process in more detail.

The ocean adjustment to ZWS forcing involves propagation of waves across the basin at all depths (e.g., Cane and Sarachik 1977; Anderson and Killworth 1977). [For our purposes, it does not matter if the forcing is a step change in wind forcing or a periodic forcing like that in Cane and Sarachik (1981).] Near the equator, there are eastward-propagating Kelvin waves that are reflected from or reflect into westward-propagating Rossby waves at the western or eastern boundary, respectively (Cane and Sarachik 1977). Away from the equator, westward-propagating Rossby waves dominate. The wind-driven barotropic mode propagates across the Pacific within approximately two weeks, and while this mode is apparent in the seasonal cycle of meridional ocean transport (e.g., Jayne and Marotzke 2001), we expect this mode to get almost completely filtered out when taking an annual average. Baroclinic Rossby waves, however, propagate more slowly. For latitudes poleward of 5° , the phase speed of these waves falls below 1 m s^{-1} , implying a cross-Pacific transit time of longer than a year. Thus, for interannual time scales, we should not expect an equilibrated ocean response to τ_x anomalies. Rather, ψ anomalies over the full ocean depth are to be expected on interannual time scales, and these anomalies should be the basin-integrated manifestation of baroclinic Kelvin and Rossby wave disturbances of the meridional flow (v).

We summarize our proposed mechanism as follows: Interannual variations in ZWS drive interannually varying Ekman transport, and these Ekman transport anomalies drive compensating flow below the Ekman layer. Based on the theoretical baroclinic adjustment time, we expect the meridional velocity anomalies comprising this compensating flow to have a baroclinic structure occupying the full depth of the ocean.

Additional analysis validates our theoretical expectations. Figure 9 shows the ECCO-derived correlation between anomalies of annual mean MOC at a given latitude-depth point and annual mean $\langle \tau_x \rangle$ at that latitude. These plots show strong correlation reaching the deep ocean at most latitudes. As expected, the correlation is mostly negative in the Northern Hemisphere (where anomalous westward wind stress generates anomalous northward Ekman transport) and positive in the Southern Hemisphere (where anomalous westward wind stress generates anomalous southward Ekman transport). The correlations are less vertically uniform equatorward of 20° , likely due to the influence of nonlocal wind forcing, as discussed earlier. Accordingly, our idealized partial coupling experiments showed substantial

FIG. 9. At each latitude and depth, the shading shows the Pearson correlation between anomalies of the annual mean MOC and the annual mean cross-basin integral of ZWS at that latitude in (a) the Indian-Pacific basin and (b) the Atlantic Ocean, computed from ECCO. The shading interval is 0.1, and grid cells below the ocean floor are shaded gray.

PMOC ISTD in the tropics, even when interannual wind stress variability was suppressed in the tropics (Fig. 6b).

An additional point is worth emphasizing: in regions where the correlation in Fig. 9 is close to vertically uniform, this does not imply that the deep ocean response of v to wind stress is barotropic. Rather, the MOC anomalies in these regions can resemble that in the Northern Hemisphere in Fig. 4, year 1998. This ψ anomaly has a single sign over the full ocean depth, but a maximum near 1500 m, which implies one sign change in v in the vertical (i.e., baroclinic structure).

We have also examined the correlation of MOC and zonally integrated wind stress curl at both lag 0 and lag 1 year (not shown), and the correlations are much weaker and less spatially coherent compared to the correlations with $\langle \tau_x \rangle$. This combined with our calculations in Fig. 8c suggest that the response of PMOC to interannual wind stress variations is primarily Ekman. Any large-scale geostrophic response is likely a secondary effect of the anomalous Ekman transport and the associated SSH anomalies (e.g., Pedlosky 1968; Jayne and Marotzke 2001), rather than a direct effect of the wind forcing.

FIG. 10. Depth-vs-longitude structure of annual mean v anomalies at (left) 7° and (right) 25°N for (top) 1996, (top middle) 1997, (bottom middle) 1998, and (bottom) 1999, computed from ECCO. The Pacific Ocean is in the left portion of each panel, and the Atlantic Ocean is in the right portion. The shading interval is 18 m day^{-1} , and grid cells under land or below the ocean floor are shaded gray.

Furthermore, the meridional flow variations associated with these MOC anomalies have vertical structures indicative of baroclinic waves. These structures are evident in Fig. 10, which shows longitude-depth profiles of v anomalies for 1996–99. The anomalies show one sign change in the vertical over much of the ocean, suggestive of the first baroclinic mode. There are also regions with two or more sign changes (e.g., just west of the mid-ocean ridges in the Pacific at 7°N ,

Fig. 10, left column), indicating higher-order baroclinic modes.

Modal decomposition provides further evidence of the role of baroclinic waves. The sum of the projections of the v anomalies onto the first three baroclinic modes is shown in Fig. 11. The vertical structure functions were obtained using the “InternalModes” MATLAB function in the GLOceanKit package (Early et al. 2020), assuming fixed stratification and a

FIG. 11. As in Fig. 10, but for the sum of the projections of ECCO annual mean v anomalies onto the first three baroclinic modes. See the text for details about the modal decomposition.

free surface at the upper boundary. We set the stratification equal to the climatological zonally averaged value of the buoyancy frequency below 500 m at the latitude of interest. (The buoyancy frequency was calculated from monthly mean ECCO variables “RHOAnoma” and “DRHODR.” We obtained very similar modal decompositions when using more realistic depth-varying stratification profiles.) We excluded the top 100 m of the ECCO data (approximately corresponding to the Ekman layer) when projecting the modes onto those data.

The sum of the projections onto the first three baroclinic modes (Fig. 11) explains much of the structure of the v anomalies in Fig. 10. The spatial correlation between the v anomalies and the sum of the projections onto the first three baroclinic modes is 0.65 at 7°N and 0.80 at 25°N. This suggests that the first three baroclinic modes explain approximately 42% of the v anomalies at 7°N and 64% at 25°N. If we project onto the barotropic mode in addition to the first three baroclinic modes, then the fraction explained increases to 43% at 7°N and 79% at 25°N, confirming

that the barotropic mode makes a relatively minor contribution to v anomalies. Furthermore, Hovmöller plots (not shown) reveal clear westward propagation of v anomalies at 25°N in the deep ocean, further confirming the key role of baroclinic Rossby waves outside of the tropics. Such westward propagation is also clear when examining Hovmöller plots of the individual baroclinic modes (not shown), providing validation of our modal decomposition approach.

Thus, we have developed a physical understanding of why we expect MOC variability over the full depth of the ocean on interannual time scales, and why this variability is stronger in the Pacific than in the Atlantic: at most latitudes, the Pacific is wider than the Atlantic, and thus interannual variability in the cross-basin integral of ZWS (which is proportional to ψ at the bottom of the Ekman layer) is larger in the Pacific than in the Atlantic. These Ekman transport variations in turn drive compensating flow variations in the deep ocean. The deep MOC variations are associated primarily with baroclinic waves that occupy the full depth of the ocean and typically take longer than a year to propagate across the Pacific basin.

6. Implications for meridional heat transport

Interannual PMOC variations are not just a dynamical curiosity: they are also highly consequential for interannual variations of MHT. To demonstrate this, we have computed MHT as

$$\text{MHT} = c_p \rho_0 \int_{A_{xz}} v \theta dA, \quad (6)$$

where θ is the ocean potential temperature (derived from monthly mean ECCO variable “THETA”) and c_p is the ocean heat capacity. Here, we set $c_p = 4281 \text{ J kg K}^{-1}$ and $\rho_0 = 1035 \text{ kg m}^{-3}$. Figure 12 shows that the ISTD of MHT is much greater in the Indian–Pacific basin (red curve) than in the Atlantic Ocean (black curve), as was the case for the MOC (Fig. 1). In much of the tropics, (especially from 20°S to 10°N) the ISTD of Indian–Pacific MHT (PMHT) is greater than the ISTD of Atlantic MHT (AMHT) by more than a factor of 3. At most latitudes, the ISTD of global MHT (green curve) is mostly accounted for by variations in PMHT.

To what extent is this global MHT variability associated with overturning in the Pacific Ocean? Fig. 13 shows the correlation of annual mean global MHT anomalies with the annual mean PMHT (red) and the annual mean AMHT (black). The PMHT correlation coefficients exceed 0.8 and are larger than the AMHT correlations at all latitudes south of 40°N, indicating that

FIG. 12. Interannual standard deviation of MHT in the Atlantic Ocean (black), the Indian–Pacific basin (red), and the global ocean (green), computed from ECCO.

PMHT plays a bigger role than AMHT in interannual variations of global MHT.

Following Bryan (1982), we further decompose the PMHT variations into contributions from overturning, PMHT_o (Fig. 13, blue circles), and gyre transport, PMHT_g (green crosses). That is,

$$\text{PMHT}_o = \rho_0 c_p \int_{A_{xz}} \langle v \rangle \langle \theta \rangle dA \quad \text{and} \quad (7)$$

$$\text{PMHT}_g = \rho_0 c_p \int_{A_{xz}} \langle v^* \theta^* \rangle dA, \quad (8)$$

where angle brackets indicate a zonal average and asterisks indicate deviations from the zonal average. Figure 13 shows that, south of 40°N, the strong correlation of PMHT with global MHT is almost entirely attributable to overturning in the Pacific.

To connect PMHT variations more explicitly with dynamical variations, Fig. 14 shows time series of detrended PMHT anomalies (red) as well as the PMHT when the θ time series is replaced by the 1992–2011 climatology of θ (black). The black curve very closely reproduces the red curve, showing that interannual MHT variations are driven almost entirely by variations in meridional flow rather than temperature variations. Quantitatively, the MHT anomalies are comparable to mean values of MHT in the Indian–Pacific basin (e.g., Trenberth and Caron 2001), indicating that interannual variability of Pacific MHT can generate large departures from climatological MHT. By construction, the anomalies in Fig. 14 cancel perfectly in the long-term average. However, because of the meridional temperature gradient, it is possible that interannual MOC anomalies generate long-term cumulative effects on the ocean heat

FIG. 13. Pearson correlation at each latitude of the annual mean global MHT with Indian-Pacific MHT (red), Atlantic MHT (black), Pacific MHT due to overturning (blue), and Pacific MHT due to the gyre circulation (green) computed from ECCO. See the text for additional details about these calculations.

distribution. Quantifying such long-term effects requires additional work beyond the simple Eulerian diagnostics used here.

We again use (7) and (8) to isolate the overturning and gyre contributions to the PMHT anomalies in Fig. 14. This calculation shows that almost all of the MHT variations can be explained by variations in overturning (blue circles). We reach the same conclusions when examining MHT variations at the equator and in the Southern Hemisphere (not shown).

A simple calculation shows that this interannual PMHT variability is likely associated with deep rather than shallow overturning. One can approximate

$$\text{MHT}' \approx \rho_0 c_p \psi' \Delta T, \quad (9)$$

where ΔT is the vertical temperature contrast in the ocean and primes indicate departures from the time mean. From Fig. 14a, one can estimate $\text{MHT}' \approx 0.2$ PW; from Fig. 1e, $\psi' \approx 2$ Sv. Then (9) produces $\Delta T \approx 23$ K, which is comparable to the vertical temperature contrast over the full depth of the ocean and much greater than the vertical temperature contrast in the upper ocean. This result suggests that the ψ anomalies generating interannual PMHT anomalies span the full ocean depth. We can alternatively pick a higher value of ψ' , representative of the tropical ocean above 1000 m, but MHT' is also higher in this region (Fig. 14b), and the conclusion remains the same.

Altogether, these results show the strong role of deep PMOC variations in driving interannual MHT variations in the Pacific and globally. Thus, interannual PMOC variability is a potentially important influence on interannual climate variability.

FIG. 14. Detrended annual mean anomalies of PMHT (red) at (a) 25° and (b) 7°N, computed from ECCO. Also shown are the PMHT anomalies after replacing potential temperature with the 1992–2011 potential temperature climatology (black), the overturning contribution to the PMHT anomalies (blue), and the gyre contribution to the PMHT anomalies (green). See the text for additional details about these calculations. The vertical scales in the two panels are different.

7. Summary and conclusions

Through in-depth analysis of an ocean state estimate and output of fully and partially coupled climate model simulations, we have shown the following:

- 1) Interannual MOC variability is larger in the Pacific than in the Atlantic at most latitudes and over the full depth of the ocean. This finding is robust across modern climate models and ECCO.
- 2) The dominance of interannual PMOC variability is expected since the Pacific Ocean is wider than the Atlantic Ocean at most latitudes, leading to larger Pacific variation in the cross-basin integral of ZWS.
- 3) Strong interannual MOC variability is expected in the deep Pacific Ocean. This is because the baroclinic adjustment time of the deep Pacific Ocean to wind forcing (i.e., the cross-basin transit time for baroclinic waves) is longer than a year at most latitudes.

- 4) Interannual PMOC variability has large-scale spatial structure, its most prominent feature being a cross-equatorial cell spanning the tropics.
- 5) Interannual PMOC variability is the dominant driver of interannual variations in global MHT at most latitudes.

Important questions remain that call for further study. We found that interannual PMOC variability is mostly wind driven, but it remains unclear why the variability has the precise large-scale spatial structure that it has. We also found evidence of interaction between interannual PMOC variability and ENSO, along with a possible feedback between ENSO and PMOC (Fig. 7). This is a topic in need of further investigation, as such interaction may be important for ENSO phase changes, ENSO diversity and variations in the strength of ENSO teleconnections.

The robustness of our results across models and ECCO, along with the theoretical support for our findings, suggests that the interannual PMOC variability shown in this study is realistic. In situ observations of the deep Pacific Ocean are too sparse to directly verify the existence of strong interannual PMOC variations. The results of this study add to other evidence of the Pacific's importance for the global ocean circulation (e.g., Newsom and Thompson 2018) and serve as motivation to greatly improve observational monitoring of the deep Pacific Ocean.

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