

Time-mean Tropical Circulation in an Idealized Moist GCM: Dependence on Zonally Asymmetric Forcings

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11 ABSTRACT: This study investigates how the time-mean tropical circulation is affected by zon-
12 ally varying surface boundary conditions in an idealized moist aquaplanet model. We compare a
13 zonally symmetric control with two perturbation cases driven by prescribed zonal wavenumber-1
14 heating and cooling patterns of varying intensities, creating tropical warm and cold pools. The
15 forcings substantially slow the Hadley circulation (HC) and induce a comparatively weak Walker
16 circulation (WC). Through modifications of existing theories, the HC slowdown is attributed to
17 stronger zonal-mean relative vorticity, which outweighs the competing effect of increased eddy
18 momentum flux divergence.

19 Using the local meridional circulation (LMC) framework to examine zonal changes in the spatially
20 explicit divergent circulations, we find two key results. First, the HC slowdown reflects the can-
21 cellation of opposing regional responses: LMC strengthening over the warm pool and weakening
22 over the cold pool. Second, despite the zonally asymmetric forcings, large-scale vertical motion is
23 overwhelmingly dictated by the LMC, as opposed to the WC. To better understand the LMC, we
24 develop a diagnostic framework based on the zonal momentum budget. This illustrates how the
25 LMC is shaped by local relative vorticity, eddy momentum flux divergence, and the zonal gradient
26 of the Bernoulli function. The zonal changes in the LMC arise fundamentally from the Bernoulli
27 function, which is offset partly by the flux divergence, while the relative vorticity plays an outsized
28 role over the cold pool. Our results suggest that the LMC may hold the key to disentangling the
29 roles of the HC and WC in the tropics.

1. Introduction

The time-mean tropical circulation (hereafter simply referred to as the tropical circulation) plays a central role in the global climate system by redistributing energy, momentum, and moisture. It is traditionally conceptualized through two distinct overturning patterns: the zonal-mean (ZM) meridional or Hadley circulation (HC), which drives the exchange between the deep tropics and subtropics, and the deep-tropical zonal or Walker circulation (WC), which facilitates the exchange within the deep tropics. Both components of the circulation, orthogonal in direction, work together to shape the distribution of regional precipitation in the tropics, and to interact with the extratropics through teleconnections, with profound implications for the global weather and climate.

The factors controlling the HC strength have been studied rigorously, often conceptualized through the simplified upper-tropospheric zonal momentum equation (Held 2000; Schneider 2006):

$$f(1 - R_0)[\bar{v}] = [\bar{S}], \quad (1)$$

where f is the Coriolis parameter, R_0 the local Rossby number ($R_0 = -[\bar{\zeta}]/f$; see Schneider 2006), ζ the relative vorticity, v the meridional wind, and S the total eddy momentum flux divergence, where $\bar{\cdot}$ represents a time mean and $[\cdot]$ a zonal mean. Following this equation, one may consider the local Rossby number as controlling the sensitivity of the mean circulation to imposed changes in the eddy forcing (Held 2000); in response to the eddy forcing, both the local Rossby number as well as the meridional wind may respond to satisfy Eq. (1).

Theoretical work conventionally categorizes the HC into two limiting regimes based on this balance. In the first limit ($R_0 \approx 1$), eddies cannot influence the mean circulation, absolute angular momentum (AAM) is conserved, and energetic constraints, such as the equal-area constraint, dictate the HC strength and extent (Schneider 1977; Held and Hou 1980; Lindzen and Hou 1988; Plumb and Hou 1992; Fang and Tung 1996; Hill et al. 2022; Lionello et al. 2024). In the second limit ($R_0 \approx 0$), the mean meridional circulation is directly proportional to eddy momentum fluxes, functioning similarly to the midlatitude Ferrel cell (Held 2000). As shown by Walker and Schneider (2006), the scaling of the HC strength with eddy activity is upheld in many climate configurations, only faltering when $R_0 \approx 0.7$, indicating a regime where both the eddies and AAM are impactful.

56 While this construction provides profound insight into the HC's response to axisymmetric forc-
57 ings (e.g., Merlis et al. 2013; Singh et al. 2017), the balance in Eq. (1) becomes opaque when
58 considering an imposed forcing with no ZM component. This limitation becomes problematic
59 when interpreting the interactions between the HC and WC, since the WC is fundamentally a
60 zonally dependent circulation with no ZM component.

61 The WC is usually thought to arise in direct response to the zonal asymmetries in diabatic
62 heating or sea surface temperatures (SST) (Bjerknes 1969); more recent work on the WC continues
63 to support this basic picture (e.g, Gill 1980; Bretherton and Sobel 2002; Iipponen and Donner
64 2021). While both observational and modeling studies showed that zonally asymmetric (ZA)
65 forcings tend to weaken the HC (Oort and Yienger 1996; Yun et al. 2021; MacDonald and Ming
66 2022), no mechanistic explanation has been provided based on existing theories of the HC strength.
67 Beyond this, if the imposition of a purely ZA forcing of the tropical SST (that is, a forcing with
68 no ZM component) modifies the HC, one would expect the ZM integral that constitutes the HC
69 strength to comprise cancellations between different longitudinal sectors. To that end, it would be
70 useful to theorize the meridional circulation beyond the ZM.

71 The longitudinally explicit meridional circulation derives from Keyser et al. (1989) and has been
72 adapted to quantify the meridional and zonal circulations at a specific location or over a region
73 (Schwendike et al. 2014; Hu et al. 2018; Lionello et al. 2024). It is particularly effective in depicting
74 the local meridional circulation (LMC)—which aggregates precisely to the HC—revealing that the
75 El Niño-Southern Oscillation (ENSO) can give rise to substantial variability in the LMC (Klein
76 et al. 1999; Wang 2019; Galanti et al. 2022; Li et al. 2023). Unlike the HC, however, the LMC has
77 no established theories, raising questions about what factors may govern it and how they relate to
78 the HC.

79 The goal of this study is threefold. First, we aim to determine the mechanisms behind the
80 HC weakening in response to a ZA forcing. Second, we use the LMC framework to understand
81 how the overall weakening manifests in the zonal direction. Third, we investigate the dynamics
82 that control the LMC. Specifically, we employ an idealized general circulation model (GCM)
83 and perturb its mixed-layer ocean energy budget to force WCs of varying strengths. We modify
84 existing theories to explain the simulated HC weakening, while introducing a new theory for the

85 local upper-tropospheric meridional wind to quantify the controlling factors of the LMC. We then
86 apply the new theory to study how ZA forcings alter the tropical circulation over different regions.

87 The paper is organized as follows. Section 2 outlines the model configuration and experimental
88 design. In Section 3, we establish the weakening effect that the ZA forcings exert on the HC,
89 evaluate the zonal structure of the meridional circulation, and compare the opposite regional
90 responses. Section 4 contains a discussion on the key dynamical mechanisms governing the
91 circulation changes. The main findings are summarized in Section 5.

92 **2. Methodology**

93 *a. Model Description*

94 We use the Model of an Idealized Moist Atmosphere (MiMA) in an aquaplanet configuration
95 (Vallis et al. 2018). Originating from an idealized moist GCM developed on the Flexible Modeling
96 System (FMS) at NOAA’s Geophysical Fluid Dynamics Laboratory (GFDL) (Frierson 2007; Balaji
97 2011), MiMA runs on a spectral dynamical core at T42 resolution corresponding roughly to a
98 gaussian grid of 2.8° latitude by 2.8° longitude. The vertical domain comprises 50 pressure levels
99 distributed following an uneven sigma coordinate ranging from 3 to 995 hPa. The comprehensive
100 RRTMG radiation scheme (Mlawer et al. 1997) is adopted to simulate both shortwave and longwave
101 radiative transfer interacting fully with water vapor, carbon dioxide (CO_2), and ozone.

102 Subgrid convection is parameterized using a simplified Betts-Miller scheme (Betts 1986; Betts
103 and Miller 1993; Frierson 2007). Precipitation is triggered when the relative humidity exceeds
104 a threshold of 80%, and lasts until it is relaxed back to the threshold over a fixed timescale.
105 Precipitation is treated as falling immediately to the surface without re-evaporation. As such, the
106 model does not maintain clouds (i.e., atmospheric liquid water or ice content). For the boundary
107 layer, the model relies on the Monin-Obukhov similarity theory to determine drag coefficients for the
108 bulk aerodynamic formulas and turbulent diffusivities for sensible heat, momentum and moisture
109 transport caused by unresolved eddies. The specific implementation used here considers an
110 unstable branch, which has been shown to be important for equatorial wave simulation (O’Gorman
111 and Schneider 2008; MacDonald and Ming 2022). A comprehensive description of the MiMA
112 configuration can be found in Jucker and Gerber (2017).

113 The model's lower boundary is a mixed-layer ocean with a uniform heat capacity. Its thermody-
 114 namic budget is governed by:

$$\rho C_p D \frac{dT}{dt} = R - L_v E - H - \nabla \cdot \mathbf{F}, \quad (2)$$

115 where ρ is the density of liquid water, C_p the specific heat capacity of liquid water, D the ocean
 116 mixed-layer depth, and T the SST. The right-hand side represents the fluxes into and out of the
 117 mixed-layer: R the net surface radiative flux, L_v the latent heat of vaporization, E the evaporation
 118 rate, H the sensible heat flux, and $\nabla \cdot \mathbf{F}$ the divergence of a prescribed ocean heat flux.

119 *b. Experimental Setup*

120 To perturb the mixed-layer energy budget and induce a WC, we introduce a ZA pattern of
 121 ocean heat flux divergence. The prescribed forcing is constructed as a wavenumber-1 sinusoid in
 122 longitude multiplied by a Gaussian distribution in latitude:

$$\nabla \cdot \mathbf{F} = Q \sin(\lambda) e^{-\phi^2/(2\phi_0^2)}, \quad (3)$$

123 where λ is longitude and ϕ latitude. The latitudinal extent of the heating perturbation is set by ϕ_0
 124 fixed at 13° . Finally, Q represents the amplitude of the forcing. We evaluate three experiments by
 125 varying Q : 0, 25, and 50 W m^{-2} . The control case (Q_0) represents a zonally symmetric climate
 126 without a WC, while the perturbation cases (Q_{25} and Q_{50}) feature WCs of varying strengths.
 127 For all experiments presented in this study, the orbital eccentricity and obliquity are set to zero,
 128 maintaining a state of perpetual equinox, and the CO_2 concentration is fixed at a preindustrial value
 129 of 300 ppm. Because clouds are omitted, a globally uniform surface albedo of 0.22 is chosen to
 130 maintain an Earth-like climate. The ocean mixed-layer depth D is set at 20 m. Each experiment
 131 was spun up for 10 years, followed by an additional 50 years of simulation for analysis. In the
 132 subsequent sections, we analyze these experiments to quantify and understand the differences
 133 between the symmetric control case and the ZA perturbation cases. Note that our experimental
 134 setup is similar to MacDonald and Ming (2022), who investigated the dynamics of tropical waves.

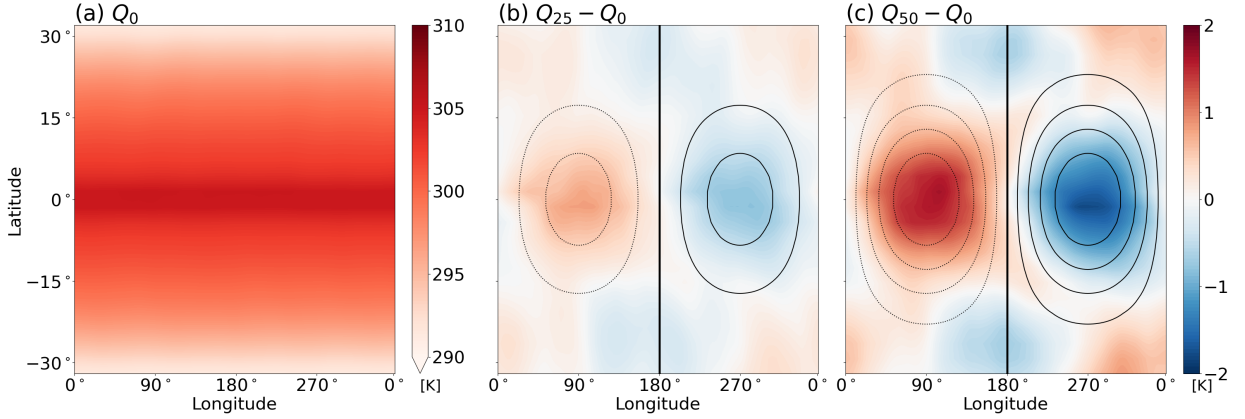


FIG. 1: (a) SST (K) in the Q_0 experiment. (b) Differences in SST ($Q_{25} - Q_0$), and (c) ($Q_{50} - Q_0$) (shading). Contour lines denote the ocean heat flux divergence in increments of 10 W m^{-2} , with thin solid (dotted) lines representing positive/divergence (negative/convergence), and thick solid lines representing zero.

3. Results

a. Basic Characteristics

The climatological SST in the control experiment (Q_0) is perfectly zonally symmetric, decreasing monotonically with latitude in both hemispheres (Fig. 1a). In a steady state, the prescribed ocean heat fluxes introduced in Q_{25} and Q_{50} must be compensated by the radiative and turbulent fluxes across the lower boundary, resulting in higher (lower) SST in a region of ocean heat flux convergence (divergence). This is borne out in the model simulations. The resulting SST anomalies relative to the control reveal a warm pool spanning $0\text{--}180^\circ$ longitude and a cold pool spanning $180\text{--}360^\circ$ (Figs. 1b and 1c). The SST difference between the centers of the warm and cold pools (i.e., 90° and 270° longitude on the equator) stands at 1.5 K in Q_{25} and 3.2 K in Q_{50} . The ZM SST is essentially identical across all three experiments (not shown), despite the ZA perturbations.

Consistent with the SST pattern, the near-equatorial tropical circulation in Q_0 exhibits strong, zonally symmetric ascents (i.e., negative pressure velocity ω) (Fig. 2a). The subsidence begins around 7° latitude. The introduction of the ZA forcings in Q_{25} and Q_{50} produces robust anomalous circulations, which manifest as ascents over the warm pool and descents over the cold pool featuring a distinct wavenumber-1 structure in the deep tropics (Figs. 2b and 2c).

Mirroring closely the patterns of pressure velocity, precipitation (Figs. 2d-f) can be approximated as a product of moisture stratification and ascending motion (O’Gorman and Schneider 2008). Over

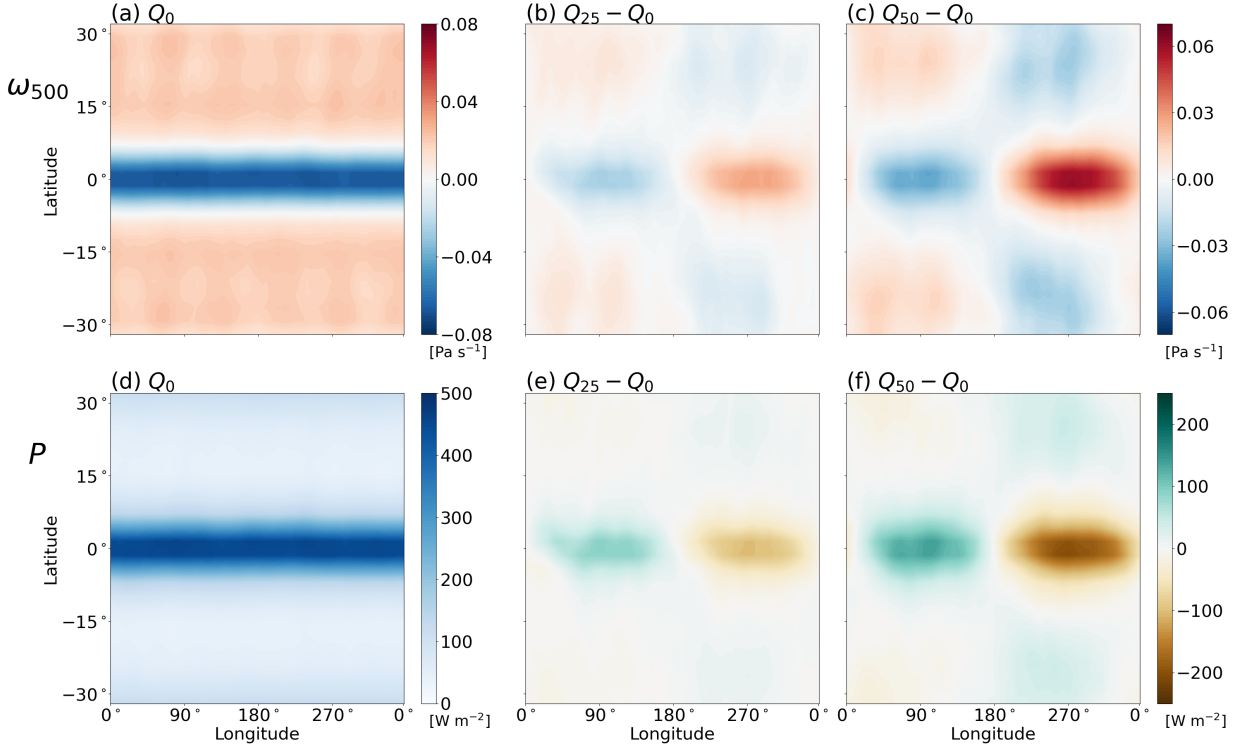


FIG. 2: (a) 500-hPa pressure velocity (ω_{500}) (Pa s^{-1}) in the Q_0 experiment. (b) Differences in ω_{500} ($Q_{25} - Q_0$), and (c) ($Q_{50} - Q_0$). (d) Precipitation (P) (W m^{-2}) in the Q_0 experiment. (e) Differences in P ($Q_{25} - Q_0$), and (f) ($Q_{50} - Q_0$).

the warm pool, the enhanced ascents are consistent with the localized increase in precipitation. The reverse occurs over the cold pool, where anomalous descents act to suppress precipitation.

b. Hadley Circulation and Walker Circulation

It is customary to quantify the HC with the classical meridional mass streamfunction ψ_{CM} (see Appendix A):

$$\psi_{CM}(\phi, p) = \frac{2\pi a \cos \phi}{g} \int_0^p [\bar{v}](\phi, p') dp' \quad (4)$$

where a is the radius of Earth, and g the gravitational acceleration. Because ψ_{CM} is closed, its contours are streamlines. The circulation strength of an individual cell is measured by the absolute value of its ψ_{CM} extrema (Peixóto and Oort 1983). In the control experiment (Q_0), ψ_{CM} exhibits two cells that are symmetric with respect to the equator (Fig. 3a). We quantify the total air mass ascending or descending across the entire HC per unit time by summing the absolute values of

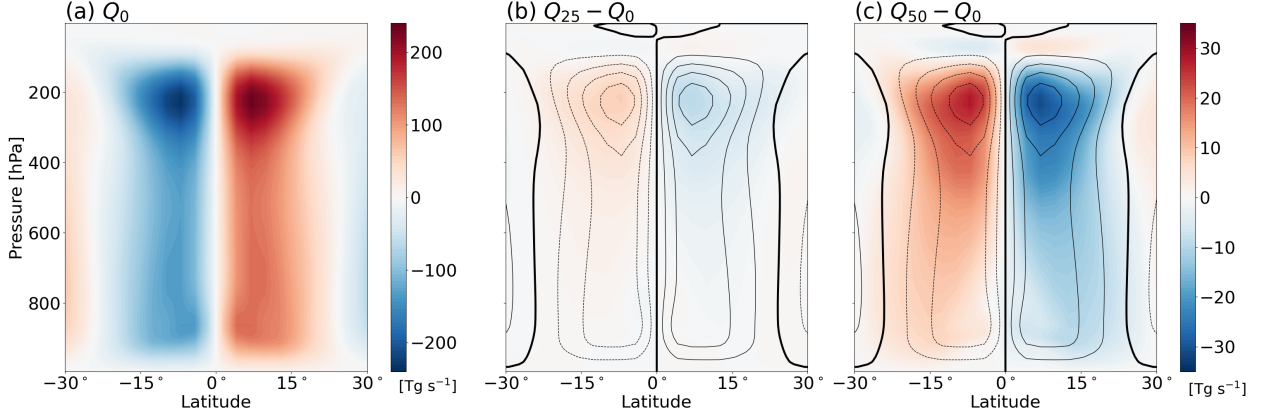


FIG. 3: (a) Classical meridional streamfunction ψ_{CM} (Tg s^{-1}) in the Q_0 experiment, with positive (negative) values representing clockwise (counter-clockwise) rotation. (b) Differences in ψ_{CM} ($Q_{25} - Q_0$), and (c) ($Q_{50} - Q_0$) (shading), with thin contour lines representing ψ_{CM} in Q_0 in increments of 50 Tg s^{-1} , and thick solid lines representing zero.

the extrema of both cells. Centered around 240 hPa, the circulation is more top-heavy than those found in comprehensive models (e.g., Clark et al. 2018). The strength of the circulation stands at 466 Tg s^{-1} , divided evenly between two cells. Under the ZA forcings, the basic two-cell structure remains intact, but the HC undergoes a pronounced slowdown (Figs. 3b and 3c). The circulation strength is 448 Tg s^{-1} in Q_{25} and 404 Tg s^{-1} in Q_{50} , which amount to reductions from Q_0 of 4% and 13%, respectively. This result is consistent with MacDonald and Ming (2022).

In contrast to the HC, it remains ambiguous how to best describe the WC in the literature. Deep-tropical zonal SST and SLP gradients have been widely used as proxies for the circulation strength (e.g., Vecchi et al. 2006; Zhao and Allen 2019; Kang et al. 2020; Heede and Fedorov 2023), but neither is a direct metric of the WC. Looking beyond surface properties, some studies evaluated the WC with the ZA pressure velocity (ω^*), where $\cdot^*$ denotes a deviation from the ZM (e.g., Vecchi and Soden 2007; Merlis and Schneider 2011). Others constructed zonal streamfunctions from vertical integrals of different zonal wind components such as the ZA zonal wind (u^*) (e.g., Hartmann et al. 1984; Hartmann 2015) and the divergent zonal wind (u_{div}) (e.g., Yu and Zwiers 2010; Bayr et al. 2014; Li et al. 2021). Here, we use the same classical zonal streamfunction ψ_{CZ} as defined in Li et al. (2021) (see Appendix A):

$$\psi_{CZ}(\lambda, p) = \frac{a\Delta\phi}{g} \int_0^p \langle \bar{u}_{\text{div}} \rangle_{\Delta\phi}(\lambda, p') dp', \quad (5)$$

179 where $\Delta\phi$ is the latitudinal range over which the WC is defined, u_{div} the divergent zonal wind
 180 based on the Helmholtz decomposition, and $\langle \cdot \rangle_{\Delta\phi}$ a meridional average over the specified range
 181 $\Delta\phi$. Figure 4 shows ψ_{CZ} in the deep tropics (defined between -5° and 5° latitude) as contours. In
 182 both perturbation experiments, the WC consists of two cells, displaying a characteristic westward
 183 tilt with height above 800 hPa, a feature attributed to the vertical structure of the Kelvin waves
 184 excited by diabatic heating (Hartmann et al. 1984; Schumacher et al. 2004; Iipponen and Donner
 185 2021). The clockwise cell to the east of the warm pool is stronger than the counter-clockwise one
 186 to the west. Their centers are located around 300 hPa, slightly lower than the HC. They are also
 187 very weak below 800 hPa. In contrast, the HC extends close to the surface. The WC strength is
 188 2.05 Tg s^{-1} (1.1 Tg s^{-1} in the clockwise cell, and 0.95 Tg s^{-1} in counter-clockwise cell) in Q_{25} ,
 189 and 3.9 Tg s^{-1} (2.1 Tg s^{-1} in the clockwise cell, and 1.8 Tg s^{-1} in the counter-clockwise cell) in
 190 Q_{50} , accounting for less than 1% of the mass transport accomplished by the HC.

191 To make connections with the existing literature, we compare ψ_{CZ} with ω^* in Fig. 4. If one
 192 were to treat the contours of ψ_{CZ} as approximate streamlines (despite the circulation not being
 193 strictly closed; see Section 3d), the vertical motion would be dictated by the horizontal gradient
 194 of the streamfunction. However, the spatial distributions of ψ_{CZ} and ω^* clearly deviate from this
 195 behavior. For example, in the Q_{50} experiment (Fig. 4b), the strongest ascents (descents) derived
 196 from ψ_{CZ} are located below 300 hPa and around 180° (350°) longitude, while the strongest ascents
 197 (descents) measured in ω^* occur at 240 hPa and 90° (270°) longitude. This spatial mismatch
 198 strongly indicates that one cannot use ω^* as a proxy for the WC.

199 It may be tempting to draw a direct link between the HC weakening and the WC strengthening
 200 in response to the ZA forcings. If the combined circulation strength has to stay constant for some
 201 reason [e.g., the atmospheric energetic constraint on global-mean precipitation (Held and Soden
 202 2006)], one may reason that the weakening of one circulation may lead to the strengthening of the
 203 other. It is, however, important to note that the HC weakening of 62 Tg s^{-1} from Q_0 to Q_{50} is 16
 204 times of the corresponding WC strengthening (3.9 Tg s^{-1}). This raises the possibility that the two
 205 circulations may be governed by different mechanisms.

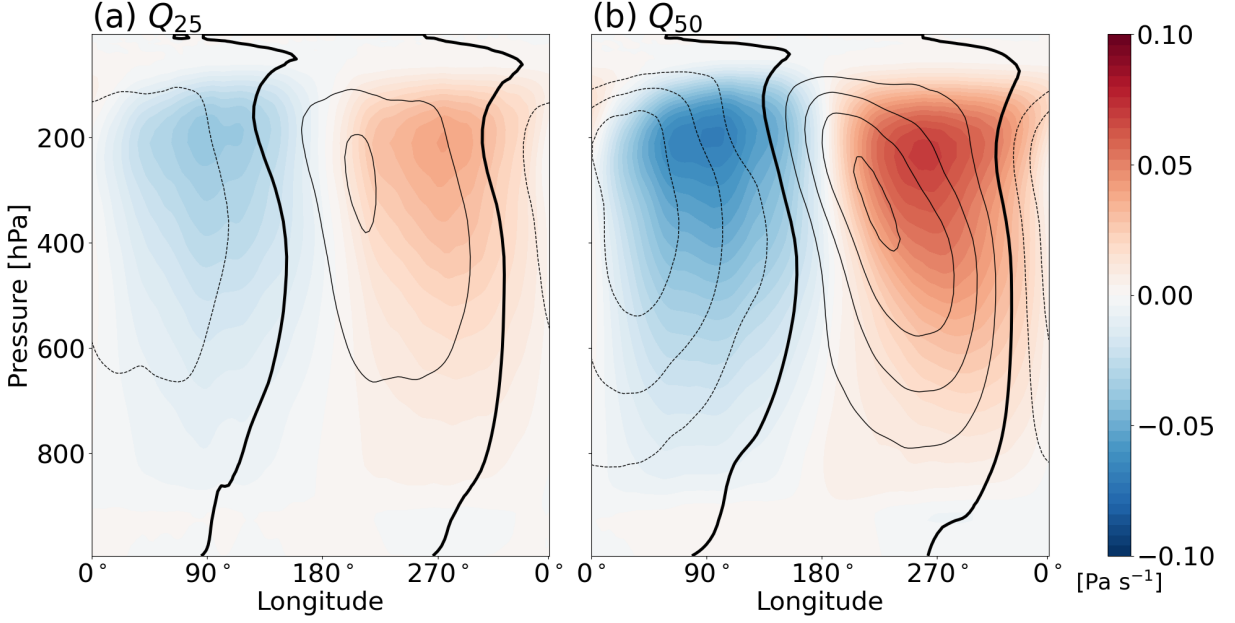


FIG. 4: Classical zonal streamfunction (ψ_{CZ}) (contours) and zonally asymmetric pressure velocity (ω^*) (Pa s^{-1} , shading) averaged between -5° and 5° latitude for the (a) Q_{25} and (b) Q_{50} experiments. Contours lines are in increments of 0.5 Tg s^{-1} , with thin solid (dotted) lines representing positive/clockwise (negative/counter-clockwise) rotation, and thick solid lines representing zero.

206 *c. Controlling Factors of Hadley Circulation*

207 To understand the underlying mechanisms of the HC weakening, we derive an expression for the
 208 ZM meridional wind from the zonal momentum equation (see Appendix B):

$$[\bar{v}] = \frac{[\bar{S}_T] + [\bar{S}_S]}{f + [\bar{\zeta}]}.$$
(6)

209 The ZM transient eddy momentum flux divergence can be written as:

$$[\bar{S}_T] = \frac{1}{a \cos^2 \phi} \frac{\partial ([\bar{u}'v'] \cos^2 \phi)}{\partial \phi} + \frac{\partial ([\bar{\omega}'u'])}{\partial p},$$
(7)

210 and the ZM stationary eddy momentum flux divergence as:

$$[\bar{S}_S] = \frac{1}{a \cos^2 \phi} \frac{\partial ([\bar{u}^*v^*] \cos^2 \phi)}{\partial \phi} + \frac{\partial ([\bar{\omega}^*u^*])}{\partial p},$$
(8)

where $\cdot'$ denotes a deviation from the time mean ($\bar{\cdot}$). It is clear from Eq. (6) that the ZM meridional wind is proportional to the total eddy momentum flux divergence, and inversely proportional to the ZM absolute vorticity ($f + [\bar{\zeta}]$). Both stationary and transient eddies contribute to the total eddy momentum flux divergence. As discussed in the introduction, classical theories often evaluate this balance in terms of the ZM local Rossby number $R_0 = -[\bar{\zeta}]/f$, yielding $f(1 - R_0)[\bar{v}] = [\bar{S}_T] + [\bar{S}_S]$ (Held 2000; Schneider 2006). Previous applications of this framework typically simplify the vorticity field either by assuming the eddy-driven limit of $R_0 \approx 0$ (Walker and Schneider 2006) or by assigning a constant bulk value of R_0 to the entire HC (e.g., Singh et al. 2017). While these approximations are highly effective for many climates, scalings based on the eddy-driven limit begin to falter in regimes where $R_0 \gtrsim 0.7$ (Walker and Schneider 2006). Because $R_0 \approx 0.8$ in our unperturbed Q_0 experiment, we must account for vorticity's role. Beyond this, we will show that capturing the forced spatial variations in both ZM and local relative vorticity is crucial for understanding the meridional circulation responses to the ZA forcings.

Integrating Eq. (6) vertically from the top-of-the-atmosphere to a specific pressure leads to an estimate for ψ_{CM} (denoted as ψ_{CM}^*):

$$\psi_{CM}^*(\phi, p) = \frac{2\pi a \cos \phi}{g} \int_0^p \frac{[\bar{S}_T] + [\bar{S}_S]}{f + [\bar{\zeta}]} dp'. \quad (9)$$

Note that, in the ZM, absolute vorticity is proportional to the meridional gradient of AAM. Equation 9 illustrates how both the total eddy momentum flux divergence and the ZM AAM distribution pose fundamental constraints on the HC intensity. When evaluated at the locations of the extrema of ψ_{CM} , Eq. (9) yields an estimate of 421 Tg s^{-1} for Q_0 , which is about 10% weaker than simulated. The estimated circulation slowdown is 17 Tg s^{-1} in Q_{25} , and 66 Tg s^{-1} in Q_{50} , which are in excellent agreement with the actual results (18 and 62 Tg s^{-1} , respectively), affirming the utility of Eq. 9 in interpreting the HC strength.

To understand how the ZA forcings weaken the HC, we first establish the components that set its baseline strength. Figure 5a shows the total eddy momentum flux divergence, and Fig. 6a shows the ZM zonal wind and AAM in Q_0 . The spatial structure of the total eddy momentum flux divergence reveals three distinct zones: eddies converge AAM directly over the equator (maintaining the upper-tropospheric superrotation), diverge AAM out of the off-equatorial tropics (decelerating the zonal

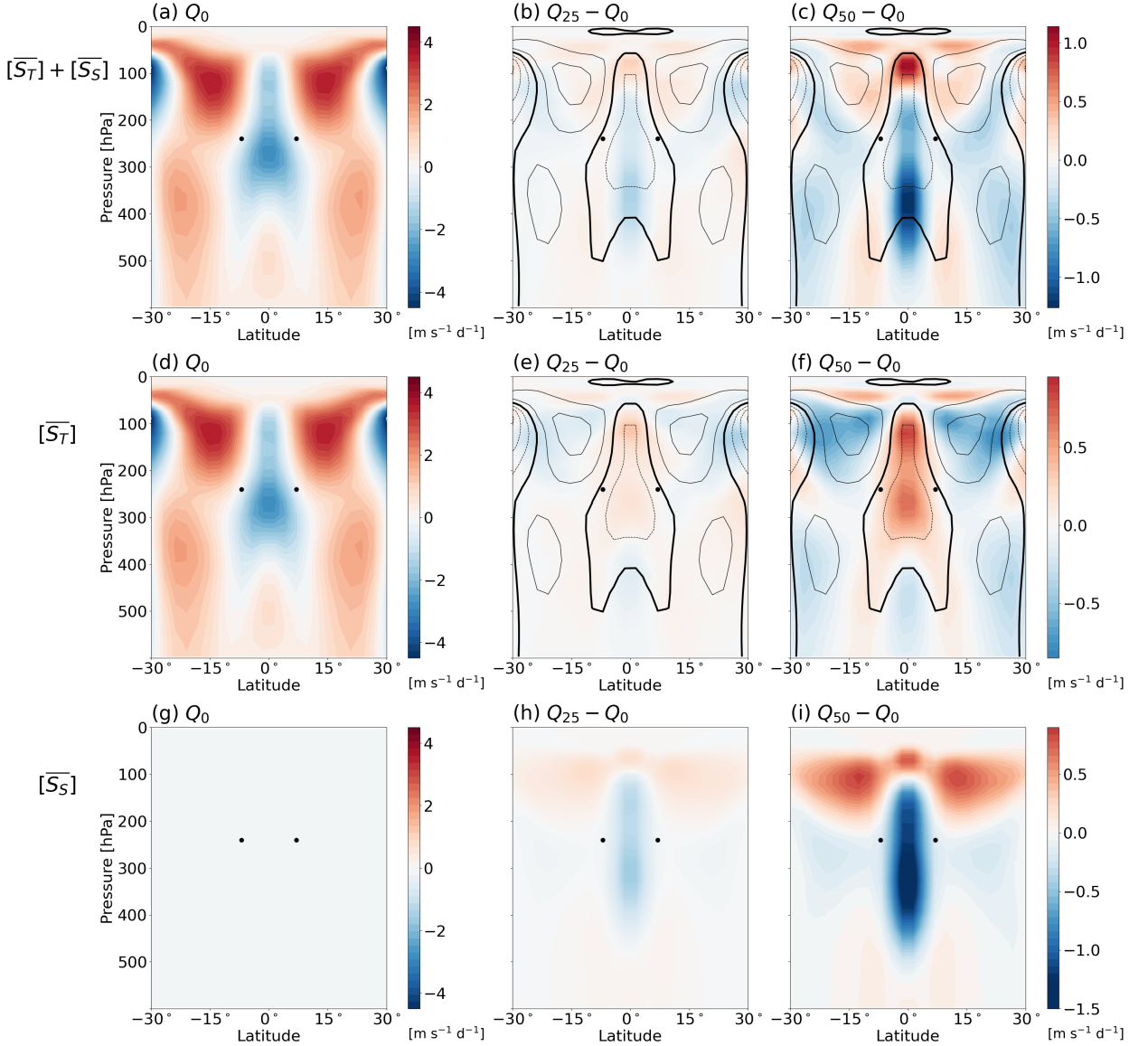


FIG. 5: (a) Zonal-mean total eddy momentum flux divergence ($\text{m s}^{-1} \text{ day}^{-1}$) in the Q_0 experiment. (b) Difference in the zonal-mean eddy momentum flux divergence ($Q_{25} - Q_0$), and (c) ($Q_{50} - Q_0$) (shading), with thin contour lines representing the values in Q_0 in increments of $1 \text{ m s}^{-1} \text{ day}^{-1}$, with thick solid lines representing zero. The black dots denote the location of the extrema of ψ_{CM} . (d)-(f) and (g)-(i) correspond to the transient and stationary eddy components, respectively.

wind), and converge AAM again in the extratropics (maintaining the eddy-driven jets). Equation (9) suggests that the ZM eddy momentum flux divergence must be zero where ψ_{CM} reaches its local extrema with respect to pressure. This is borne out approximately in Fig. 5a, indicating that strong flux divergence is critical for driving the HC. At the same time, the ZM absolute vorticity

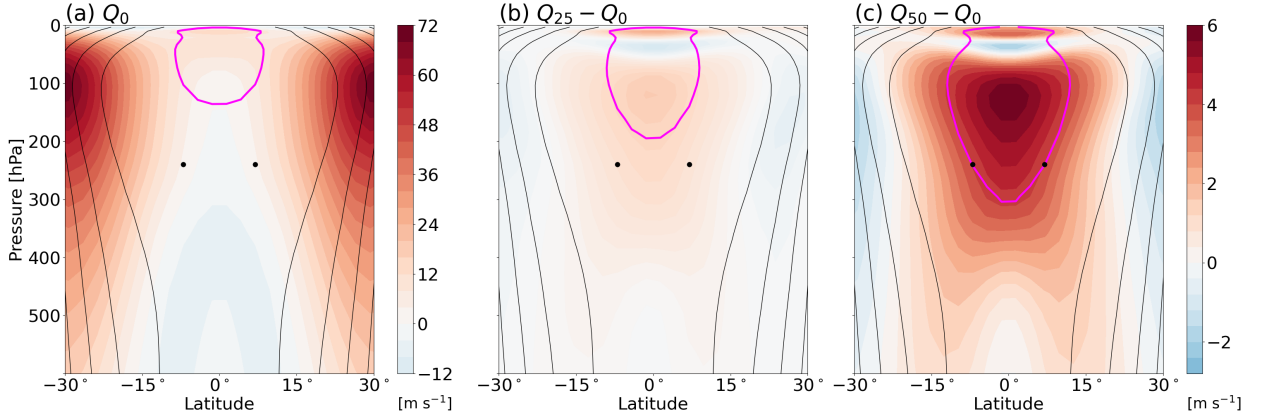


FIG. 6: (a) Zonal-mean zonal wind (shading, m s^{-1}) in the Q_0 experiment. (b) Differences in zonal-mean zonal wind ($Q_{25} - Q_0$), and (c) ($Q_{50} - Q_0$) (shading). Absolute angular momentum (AAM) contours are in increments of $0.05a\Omega^2$. The regions of superrotation (i.e., AAM greater than $a\Omega^2$) are enclosed by the thick magenta lines. The black dots denote the location of the extrema of ψ_{CM} .

is small since the AAM is relatively uniform in the tropical upper troposphere (Fig. 6a), with a tendency to reinforce the role of the flux divergence.

The introduction of the ZA forcings disrupts this baseline, affecting both the total eddy momentum flux divergence and relative vorticity. First, in the off-equatorial upper troposphere, the flux divergence becomes stronger (Figs. 5b and 5c), with a tendency to accelerate the HC. Interestingly, the transient and stationary eddies oppose each other; the former tend to decrease the flux divergence (Figs. 5e and 5f), while the latter tend to increase it (Figs. 5h and 5i). Second, in the deep equatorial upper troposphere, there is a substantial reduction in the eddy momentum flux divergence (i.e., enhanced flux convergence). This reduction is driven exclusively by the stationary eddies and gives rise to an appreciable expansion of the superrotational zone (MacDonald and Ming 2022; Zurita-Gotor and Held 2024, 2025) (Figs. 6b and 6c). Since the AAM remains relatively unchanged in the subtropics, this localized expansion sharpens the meridional gradient of AAM. The corresponding increase in the relative vorticity decelerates the HC.

To quantify the competing effects caused by changes in the eddy forcing and vorticity, we linearly decompose Eq. (6) relative to the symmetric Q_0 baseline. Although this linear sum overestimates the actual net HC slowdown (resulting in net changes of -20 and -99 Tg s^{-1} compared to the actual values of -17 and -66 Tg s^{-1}), it effectively isolates the contributions of the individual terms. The strengthening effect from the total eddy momentum flux divergence ($+32 \text{ Tg s}^{-1}$ in Q_{25} and $+56$

260 Tg s^{-1} in Q_{50}) is heavily outweighed by the deceleration driven by the increase in relative vorticity
 261 $(-52 \text{ Tg s}^{-1}$ in Q_{25} and -155 Tg s^{-1} in $Q_{50})$, clearly demonstrating that the HC weakening is dictated
 262 by the changes in vorticity.

263 *d. Local Meridional Circulation and Local Zonal Circulation*

264 As useful as the ZM framework has been for understanding the behaviors of zonally symmetric
 265 climates, its applicability to ZA climates, which project onto the meridional and zonal directions
 266 simultaneously, is rather limited. There is a need for moving beyond the ZM framework to
 267 examine the zonal structure of the meridional circulation explicitly. To rigorously partition a three-
 268 dimensional overturning circulation into its meridional and zonal components, we follow the local
 269 mass streamfunctions proposed by Schwendike et al. (2014), which satisfy the continuity equation
 270 and represent a mathematically consistent decomposition of the tropical circulation (see Appendix
 271 A).

272 The local meridional mass streamfunction (ψ_{LM}) is defined as:

$$\psi_{LM}(\lambda, \phi, p) = \frac{1}{g} \int_0^p \bar{v}_{\text{div}}(\lambda, \phi, p') dp', \quad (10)$$

273 where v_{div} is the divergent meridional wind. Figure 7 presents ψ_{LM} at 240 hPa, the pressure level
 274 of the ψ_{CM} extrema, to highlight the zonal structure of the LMC. The Q_0 experiment exhibits zonal
 275 symmetry, achieving a maximum value of $6.0 \text{ Mg m}^{-1} \text{ s}^{-1}$ at 7° latitude (Fig. 7a). The poleward
 276 flow terminates at 21° latitude, where the thermally indirect, eddy-driven Ferrel circulation ensues.
 277 Under the ZA forcings, this zonal symmetry breaks down as the meridional mass flux undergoes a
 278 marked intensification over the warm pool (Figs. 7b and 7c). In the Q_{25} experiment, ψ_{LM} peaks
 279 at 7° latitude and 90° longitude with a maximum value of $8.2 \text{ Mg m}^{-1} \text{ s}^{-1}$. The meridional mass
 280 flux further amplifies in Q_{50} , peaking at the same location with a value of $9.6 \text{ Mg m}^{-1} \text{ s}^{-1}$. The
 281 poleward extent of the upper-tropospheric meridional flow is also sensitive to the ZA forcings,
 282 reaching up to 30° latitude in Q_{25} and 34° in Q_{50} . In contrast, ψ_{LM} weakens significantly over the
 283 cold pool. The value of ψ_{LM} at 7° latitude and 270° longitude drops to $3.0 \text{ Mg m}^{-1} \text{ s}^{-1}$ in Q_{25} and
 284 just $0.4 \text{ Mg m}^{-1} \text{ s}^{-1}$ in Q_{50} . There is also a meridional narrowing of the circulation around the
 285 cold pool, with the poleward boundary shifting equatorward from 21° in Q_0 to 17° in Q_{25} , and 11°
 286 in Q_{50} .

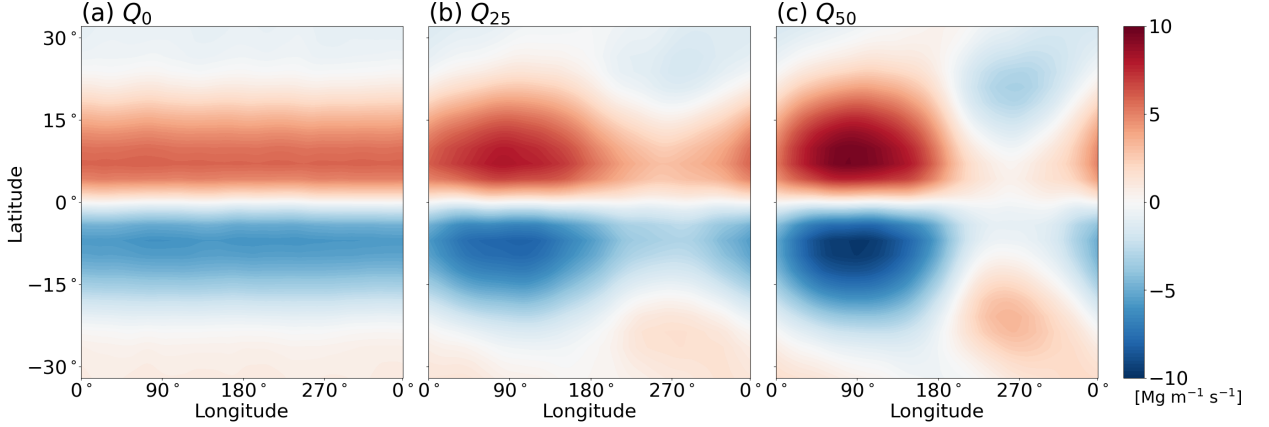


FIG. 7: Local meridional streamfunction ψ_{LM} ($\text{Mg m}^{-1} \text{s}^{-1}$) at 240 hPa in the (a) Q_0 , (b) Q_{25} , and (c) Q_{50} experiments. The sign convention is the same as in Fig. 3.

Similar to the local meridional mass streamfunction, we can also define the local zonal mass streamfunction ψ_{LZ} (see Appendix A):

$$\psi_{LZ}(\lambda, \phi, p) = \frac{1}{g} \int_0^p \bar{u}_{\text{div}}(\lambda, \phi, p') dp'. \quad (11)$$

ψ_{LZ} at 240 hPa reveals that while the zonal circulation peaks in the deep tropics, it extends across the entire tropics, with the extrema shifting gradually eastward with latitude (Fig. 8). While the magnitude scales roughly with the ZA forcing, the overall pattern remains the same.

While ψ_{LM} and ψ_{LZ} can capture changes in localized mass transport effectively, they do not translate directly to vertical motion (pressure velocity) in the same straightforward manner as the closed classical meridional streamfunction. To tie the partitioned horizontal mass transport back to local vertical motion, we can decompose the total pressure velocity into two components: $\omega = \omega_M + \omega_Z$, where ω_M represents the vertical motion required by mass continuity from the divergent meridional wind, and ω_Z represents that from the divergent zonal wind (see Appendix A):

$$\bar{\omega}_M = -\frac{g}{a \cos \phi} \frac{\partial(\psi_{LM} \cos \phi)}{\partial \phi}, \quad (12)$$

$$\bar{\omega}_Z = -\frac{g}{a \cos \phi} \frac{\partial \psi_{LZ}}{\partial \lambda}. \quad (13)$$

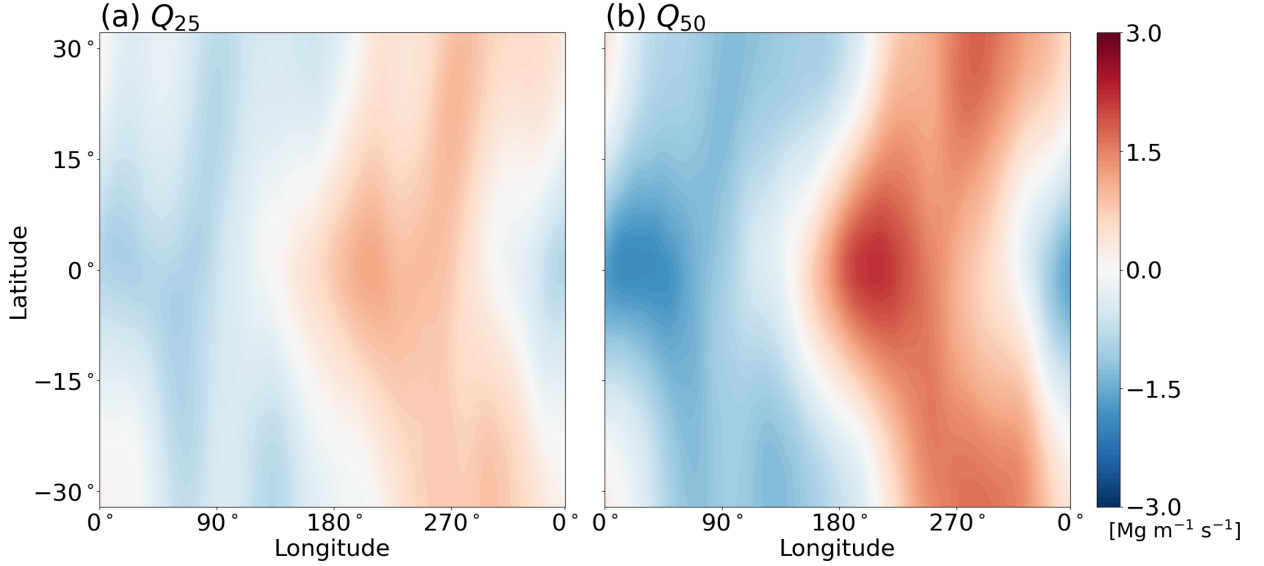


FIG. 8: Local zonal streamfunction ψ_{LZ} ($\text{Mg m}^{-1} \text{s}^{-1}$) at 240 hPa in the (a) Q_{25} , and (b) Q_{50} experiments. The sign convention is the same as the contours in Fig. 4.

Figure 9 shows that 500-hPa ω_M is nearly identical to the total ω field—a relationship that is maintained throughout the tropics at all pressure levels. The enhanced meridional gradient of ψ_{LM} over the warm pool corresponds to increased ascents in that region, while the reduced gradient over the cold pool is consistent with anomalous descents. In contrast, ω_Z is roughly two orders of magnitude smaller than ω_M everywhere, peaking at only $5.0 \times 10^{-3} \text{ Pa s}^{-1}$.

Even though both meridional and zonal winds respond to the ZA forcings, the LMC remains the primary driver of the pressure velocity response. This result is consistent with the discrepancy between ψ_{CZ} and ω^* noted earlier (Fig. 4): because the total vertical motion is overwhelmingly forced by the LMC, the WC alone cannot capture the full magnitude or structure of the zonally anomalous vertical motion.

e. Controlling Factors of Local Meridional Circulation

To isolate and quantify the underlying dynamical drivers of the LMC, we derive the following expression for the time-mean meridional wind from the zonal momentum equation (see Appendix B):

$$\bar{v}(\lambda, \phi, p) = \frac{\bar{S}_T + \bar{B}_\lambda}{f + \bar{\zeta}}. \quad (14)$$

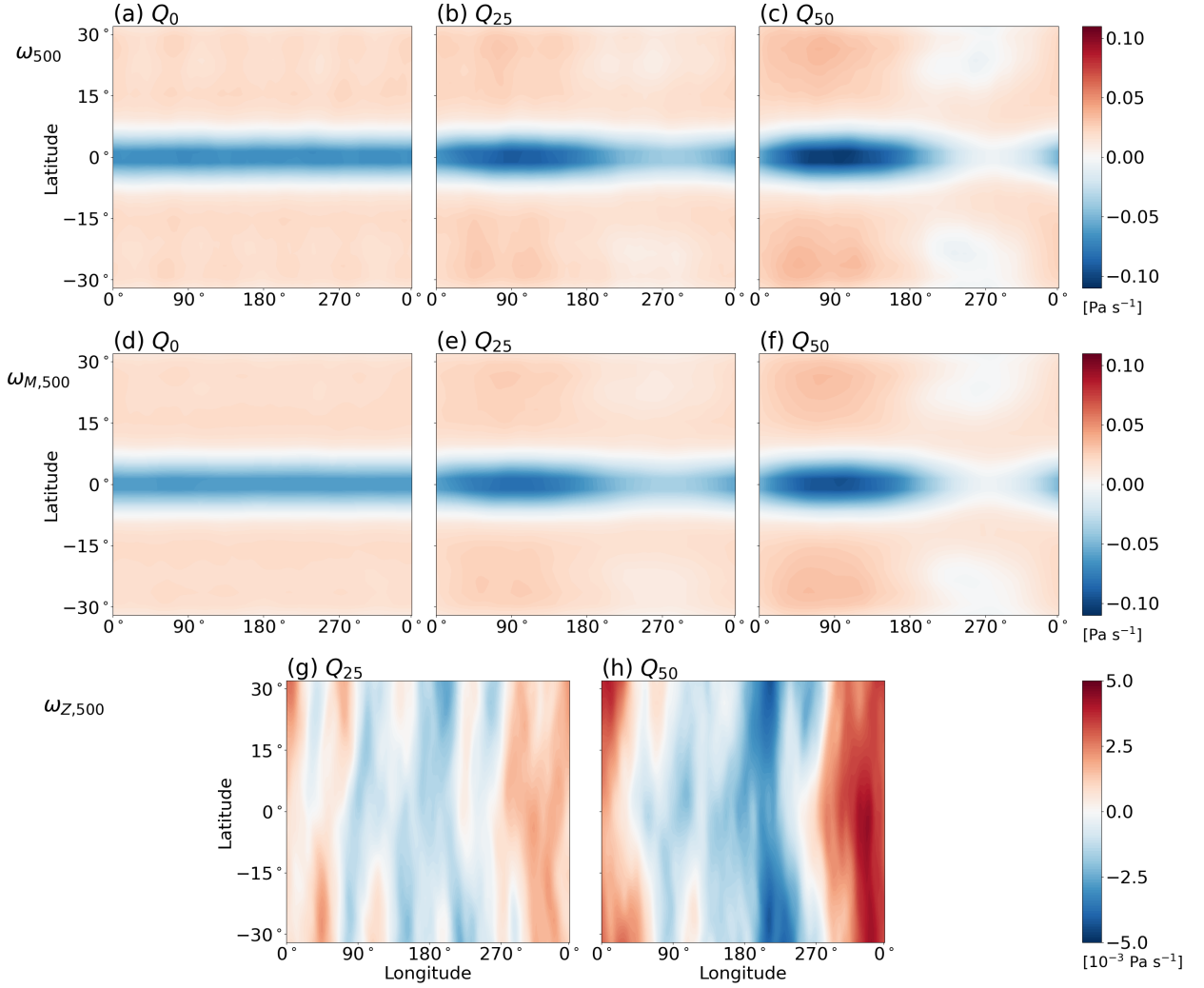


FIG. 9: Pressure velocity at 500 hPa (ω_{500}) in the (a) Q_0 , (b) Q_{25} , and (c) Q_{50} experiments. The meridional components ($\omega_{M,500}$) in (d) Q_0 , (e) Q_{25} , and (f) Q_{50} . The zonal components ($\omega_{Z,500}$) in (g) Q_{25} , and (h) Q_{50} . All are in units of Pa s^{-1} .

313 The zonal gradient of the time-mean Bernoulli function ($\bar{B}_\lambda$) is defined as:

$$\bar{B}_\lambda = \frac{1}{a \cos \phi} \frac{\partial}{\partial \lambda} \left(\bar{\Phi} + \frac{1}{2} \bar{u}^2 \right), \quad (15)$$

314 where Φ is the geopotential. The transient eddy momentum flux divergence ($\bar{S}_T$) can be computed
 315 as:

$$\bar{S}_T = \frac{1}{a \cos \phi} \frac{\partial (\overline{u'u'})}{\partial \lambda} + \frac{1}{a \cos^2 \phi} \frac{\partial (\overline{u'v'} \cos^2 \phi)}{\partial \phi} + \frac{\partial (\overline{\omega'u'})}{\partial p}. \quad (16)$$

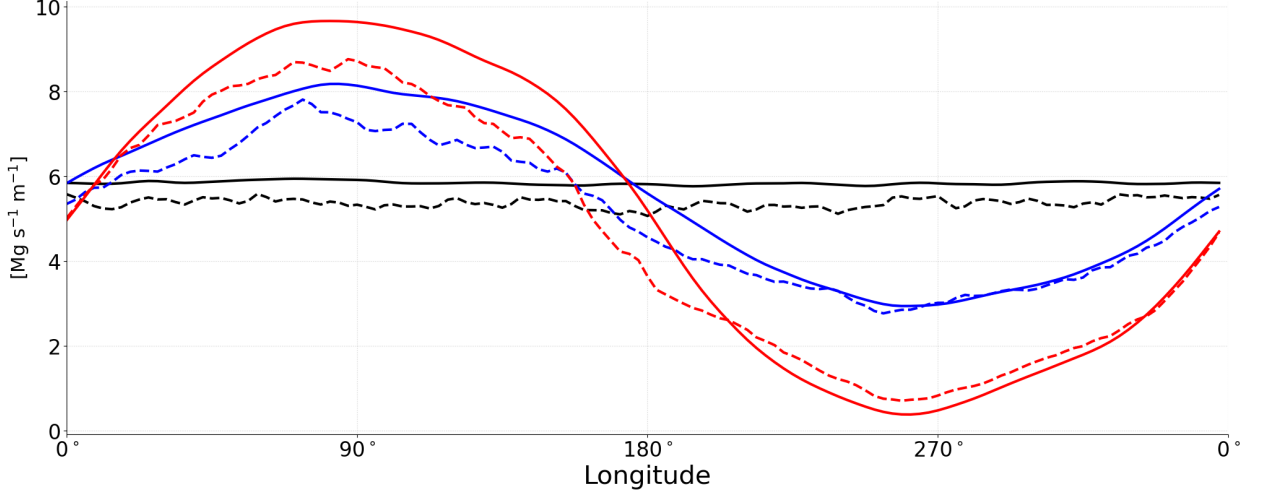


FIG. 10: Comparison of the estimated local meridional streamfunction ψ_{LM}^* ($\text{Mg m}^{-1} \text{s}^{-1}$, dashed lines) with the actual local meridional streamfunction ψ_{LM} ($\text{Mg m}^{-1} \text{s}^{-1}$, solid lines) at 7° latitude and 240 hPa in the Q_0 (black), Q_{25} (blue), and Q_{50} (red) experiments.

Equation (14) provides an estimate of the off-equatorial local meridional wind in the upper troposphere. We can integrate it over pressure from the top-of-the-atmosphere to a specific pressure level, and write an estimate of the local meridional streamfunction as:

$$\psi_{LM}^* = \frac{1}{g} \int_0^p \frac{\bar{S}_T + \bar{B}_\lambda}{f + \bar{\zeta}} dp' - \frac{1}{g} \int_0^p \bar{v}_{\text{rot}} dp'. \quad (17)$$

An important caveat is that the rotational meridional wind ($\bar{v}_{\text{rot}}$), for which we do not have a theory, must be well-behaved for ψ_{LM}^* to be useful.

A comparison of ψ_{LM}^* with the simulated ψ_{LM} at the latitude and pressure corresponding to the extrema of ψ_{CM} is shown in Fig. 10. Although the magnitude of ψ_{LM}^* is consistently smaller than the actual ψ_{LM} , it clearly captures the zonal structures of the LMC across all experiments. In particular, the warm pool experiences stronger circulation than the cold pool in Q_{25} and Q_{50} , and the contrast increases with the forcing strength to an extent that the minimum in Q_{50} is close to zero. Furthermore, ψ_{LM}^* places the maximum and minimum values at the right longitudes (around 90° and 270° , respectively).

With the theoretical estimate validated, we apply a linear perturbation analysis to isolate the specific dynamical drivers shaping these zonal variations. Linearizing the full expression for ψ_{LM}^*

(Eq. 17) yields:

$$\Delta\psi_{LM}^* = \underbrace{\frac{1}{g} \int_0^p \frac{\Delta\bar{S}_T}{f+\bar{\zeta}} dp'}_{\Delta\psi_{LM}^*(S_T)} + \underbrace{\frac{1}{g} \int_0^p \frac{\Delta\bar{B}_\lambda}{f+\bar{\zeta}} dp'}_{\Delta\psi_{LM}^*(B_\lambda)} - \underbrace{\frac{1}{g} \int_0^p \frac{(\bar{S}+\bar{B}_\lambda)\Delta\bar{\zeta}}{(f+\bar{\zeta})^2} dp'}_{\Delta\psi_{LM}^*(\zeta)} - \underbrace{\frac{1}{g} \int_0^p \Delta\bar{v}_{\text{rot}} dp'}_{\Delta\psi_{LM}^*(v_{\text{rot}})} \quad (18)$$

where Δ denotes the difference between the zonally symmetric (control) experiment (i.e., Q_0) and the zonally asymmetric experiments (i.e., Q_{25} and Q_{50}). This manipulation makes it possible to decompose the zonal LMC variations into the contributions from the transient eddy momentum flux divergence [$\Delta\psi_{LM}^*(S_T)$], the zonal gradient of the Bernoulli function [$\Delta\psi_{LM}^*(B_\lambda)$], the relative vorticity [$\Delta\psi_{LM}^*(\zeta)$], and the rotational meridional wind [$\Delta\psi_{LM}^*(v_{\text{rot}})$] (Fig. 11). All terms show imprints from the wave-1 forcings, but their magnitudes vary. Fortuitously, $\Delta\psi_{LM}^*(v_{\text{rot}})$ generally has the smallest amplitude, indicating that, within the experiments presented, the spatial structure of the LMC is fundamentally shaped by the interplay among the Bernoulli function, transient eddy momentum flux divergence, and the vorticity.

The gradient of the Bernoulli function, $\Delta\psi_{LM}^*(B_\lambda)$, acts as the primary driver of the zonal asymmetry. Because $\bar{B}_\lambda$ is strictly zero in Q_0 , $\Delta\psi_{LM}^*(B_\lambda)$ represents the actual zonal gradient of the Bernoulli function in Q_{25} and Q_{50} , which scales approximately linearly with the forcings. It tends to enhance the LMC over the warm pool, while weakening it over the cold pool. It is worth mentioning that since B_λ integrates to zero over a latitudinal circle, it cannot alter the HC strength. The changes in the transient eddies, $\Delta\psi_{LM}^*(S_T)$, generally opposes $\Delta\psi_{LM}^*(B_\lambda)$, but not enough to alter the overall zonal structure. When acting in isolation, transient eddies induce anomalous equatorward flow over the warm pool, and poleward flow over the cold pool. Unlike B_λ , the integrated effect of the transient eddies does not have to be zero, and is an important factor in determining the HC strength (see Section 3c). The effect of the relative vorticity on the LMC, $\Delta\psi_{LM}^*(\zeta)$, is negligible over the warm pool in both perturbation experiments. In stark contrast, it has a strong tendency to induce poleward flow over the cold pool (i.e., slowing down LMC). For Q_{50} , this heavily localized effect becomes so intense that it is comparable to the leading-order B_λ effect. The combined effects strain the linear approximation of Eq. (18), manifesting as the large deviation of the linear sum of all individual terms from the actual $\Delta\psi_{LM}^*$.

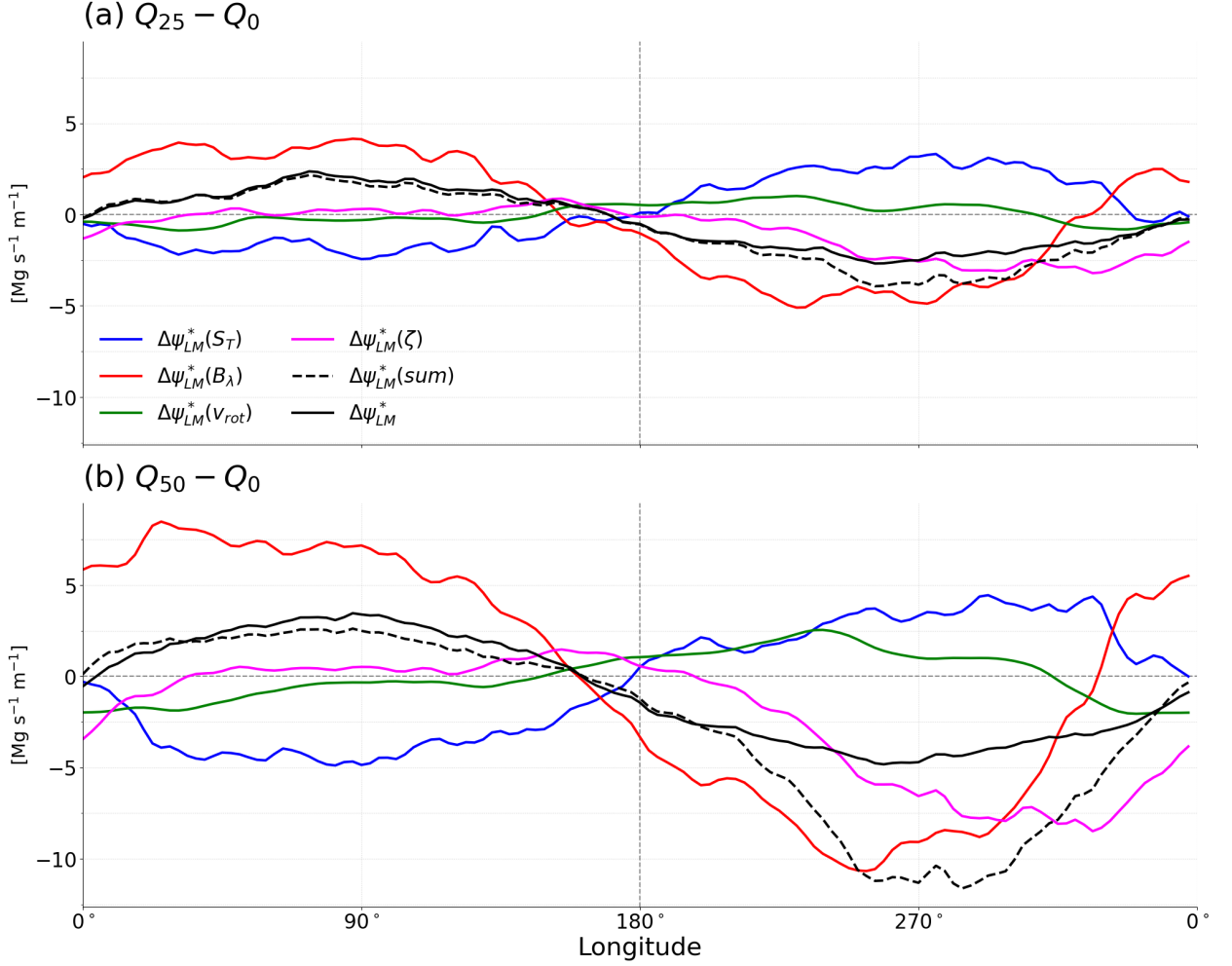


FIG. 11: Contributions from the transient eddy momentum flux divergence [$\Delta\psi_{LM}^*(S_T)$, blue lines], Bernoulli gradient [$\Delta\psi_{LM}^*(B_\lambda)$, red lines], relative vorticity [$\Delta\psi_{LM}^*(\zeta)$, magenta lines], and rotational meridional wind [$\Delta\psi_{LM}^*(v_{rot})$, green lines] to the differences in the estimated local meridional streamfunction $\Delta\psi_{LM}^*$ at 7°N integrated from top of the atmosphere to 240 hPa for (a) $Q_{25} - Q_0$ and (b) $Q_{50} - Q_0$. Also shown for comparison are the sum of these linear contributions, $\Delta\psi_{LM}^*(sum)$ (black dotted lines), and the exact difference $\Delta\psi_{LM}^*$ calculated directly from Eq. (17) (black solid lines). All are in units of $\text{Mg m}^{-1} \text{s}^{-1}$.

f. Regional Meridional Circulation

To better rationalize the roles of the warm and cold pools in the overall HC weakening, we make use of the regional meridional streamfunction ψ_{RM} , which is essentially the integral of ψ_{LM} over a longitudinal band (see Appendix A):

$$\psi_{RM}(\phi, p) = a \cos \phi \int_{\lambda_1}^{\lambda_2} \psi_{LM}(\lambda, \phi, p) d\lambda. \quad (19)$$

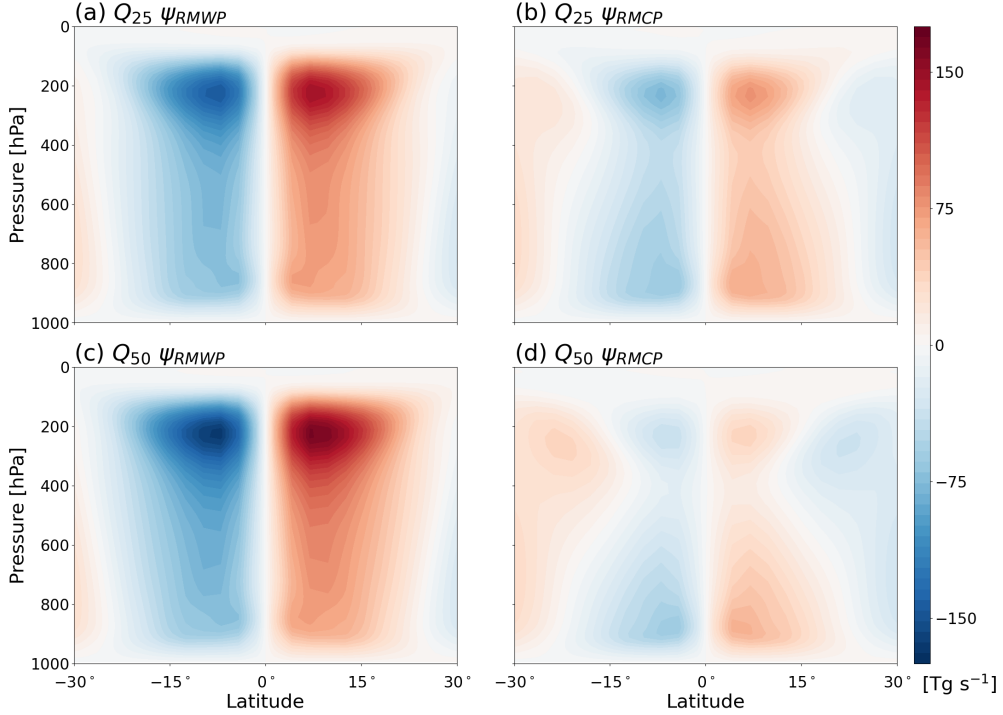


FIG. 12: Regional meridional streamfunction ψ_{RM} (Tg s^{-1}) over the warm pool (a and c) and over the cold pool (b and d) in the Q_{25} (a and b) and Q_{50} (c and d) experiments.

We define ψ_{RMWP} and ψ_{RMCP} by evaluating ψ_{RM} over the warm pool (WP; $\lambda_1 = 0$, $\lambda_2 = \pi$) and the cold pool (CP; $\lambda_1 = \pi$, $\lambda_2 = 2\pi$), respectively. By construction, they add up to the classical meridional mass streamfunction: $\psi_{CM} = \psi_{RMWP} + \psi_{RMCP}$.

Figure 12 shows ψ_{RMWP} and ψ_{RMCP} in the Q_{25} and Q_{50} experiments. The WP streamfunction, ψ_{RMWP} , peaks in strength at the same latitude and pressure as ψ_{CM} (7° and 240 hPa), with a maximum of 291 Tg s^{-1} in Q_{25} and 326 Tg s^{-1} in Q_{50} . In comparison, ψ_{RMCP} is comprised of two closed cells in each hemisphere, suggesting that both deep and shallow convection contribute to the meridional circulation over the CP. The maximum at 240 hPa associated with the deep cells is 157 Tg s^{-1} in Q_{25} and 78 Tg s^{-1} in Q_{50} . For reference, applying the same longitudinal partitioning to the zonally symmetric Q_0 experiment yields a maximum of 233 Tg s^{-1} in either sector. This indicates that the weakening of the meridional circulation over the CP is not fully compensated by the strengthening over the WP, driving the net weakening of the HC.

371 *g. Controlling Factors of Regional Meridional Circulation*

372 By integrating the linear perturbation decomposition (Eq. 18) over a longitudinal band, we
373 obtain an estimate of the regional streamfunction change relative to Q_0 :

$$\begin{aligned}\Delta\psi_{RM}^* &= a \cos \phi \int_{\lambda_1}^{\lambda_2} \Delta\psi_{LM}^*(\lambda, \phi, p) d\lambda \\ &= \Delta\psi_{RM}^*(S_T) + \Delta\psi_{LM}^*(B_\lambda) + \Delta\psi_{RM}^*(\zeta) + \Delta\psi_{RM}^*(v_{\text{rot}}).\end{aligned}\quad (20)$$

374 As noted before, the simulated regional changes are not strictly small perturbations, especially over
375 the CP, where the circulation in Q_{50} weakens by about 70%. Regardless, this linear decomposition
376 provides a first-order estimate of the relative importance of the controlling factors (Table 1).

TABLE 1: Contributions from the transient eddy momentum flux divergence $[\Delta\psi_{RM}^*(S_T)]$, Bernoulli gradient $[\Delta\psi_{RM}^*(B_\lambda)]$, relative vorticity $[\Delta\psi_{RM}^*(\zeta)]$, and rotational meridional wind $[\Delta\psi_{RM}^*(v_{\text{rot}})]$ to the differences in the estimated regional meridional streamfunction $\Delta\psi_{RM}^*$ relative to Q_0 . Also included for comparison are the sum of these linear contributions $[\Delta\psi_{RM}^*(sum)]$ and the exact difference $\Delta\psi_{RM}^*$ calculated by integrating Eq. (17) over the respective regions. All are in units of Tg s^{-1} .

Experiment	Region	$\Delta\psi_{RM}^*(S_T)$	$\Delta\psi_{RM}^*(B_\lambda)$	$\Delta\psi_{RM}^*(\zeta)$	$\Delta\psi_{RM}^*(v_{\text{rot}})$	$\Delta\psi_{RM}^*(sum)$	$\Delta\psi_{RM}$
Q_{25}	WP	-59.7	100.1	9.6	-10.5	40.1	46.6
Q_{25}	CP	75.6	-100.5	-74.4	11.0	-93.2	-66.3
Q_{50}	WP	-130.4	200.4	14.3	-26.8	57.1	73.6
Q_{50}	CP	109.3	-201.6	-190.5	27.9	-255.6	-130.4

377 The sum of the individual terms $[\Delta\psi_{RM}^*(sum)]$ consistently underestimates the strengthening in
378 the WP, but overestimates the weakening in the CP. As a result, the overall weakening of the HC
379 is significantly overestimated. However, $\Delta\psi_{RM}^*(sum)$ is able to reproduce the basic behavior of
380 the simulated $\Delta\psi_{RM}$ across both perturbations: a strengthening over the WP concurrent with a
381 weakening over the CP, with the latter outweighing the former.

382 The change in the Bernoulli term, $\Delta\psi_{RM}^*(B_\lambda)$, is the dominant driver of the regional behavior,
383 strengthening the WP circulation by 101.1 Tg s^{-1} and weakening the CP circulation by 101.5 Tg
384 s^{-1} in Q_{25} . These effects scale linearly with the forcing, and double in Q_{50} . As explained above,
385 this term must integrate to zero in the zonal direction, and cannot drive a net change in the HC. The
386 changes in the transient eddies, $\Delta\psi_{RM}^*(S_T)$, oppose the Bernoulli term in both regions, weakening

the WP and strengthening the CP. This damping scales nonlinearly across the ZA forcings. For Q_{25} , the eddy-driven strengthening of the circulation over the CP outweighs the weakening over the WP. For Q_{50} , this behavior flips: the WP weakening outpaces the CP strengthening.

A key insight from the decomposition is that $\Delta\psi_{RM}^*(\zeta)$ is the primary driver of the HC slowdown. While the change in relative vorticity has a slight strengthening effect over the WP (9.6 Tg s^{-1} in Q_{25} and 14.3 Tg s^{-1} in Q_{50}), its impact over the CP is massive, giving rise to a substantial reduction of 74.4 Tg s^{-1} in Q_{25} and 190.5 Tg s^{-1} in Q_{50} . Because the contribution from other terms is either negligibly small ($\Delta\psi_{RM}^*(v_{\text{rot}})$), integrates to zero ($\Delta\psi_{RM}^*(B_\lambda)$), or contains only a modest asymmetric effect ($\Delta\psi_{RM}^*(S_T)$), it is evident that the CP relative vorticity profile dictates the overall HC weakening.

4. Discussion

Although the zonal momentum balance framework offers strong explanatory power for the local and regional meridional circulation, it is still unclear how to link the controlling factors (i.e., S_T , B_λ , and ζ) to physical processes, and possible higher-level constraints. The model simulations show that the presence of a zonal wavenumber-1 ocean heat flux strengthens (weakens) the LMC over the WP (CP). So, a good starting point for thinking about this issue is how surface boundary conditions affect upper-tropospheric dynamics.

It is well known that ZA diabatic heating in the tropics excites stationary Rossby waves (see Held et al. (2002) for a comprehensive review). The pattern of increased precipitation over the WP and decreased precipitation over the CP (Figs. 2d-f) is effective in generating Rossby waves. Figure 13 shows the eddy streamfunction and ZA zonal wind (u^*) in the two perturbation cases at 240 hPa. The well-formed wave trains, consisting of an anticyclonic circulation over the WP and a cyclonic circulation over the CP, are almost identical to those forced by tropical heating depicted in Figs. 9a-c of Held et al. (2002). The ZA zonal wind follows the wave patterns; 240-hPa subtropical westerlies shift poleward over the WP, and equatorward over the CP, a pattern consistent with the latitudinal extent of the LMC (Fig. 7).

The stationary-wave-induced shifts of the background flow have implications for the wave breaking of mid-latitude baroclinic waves. These waves preferentially propagate equatorward, moving eastward with respect to the ground, and extract angular momentum through wave breaking. This

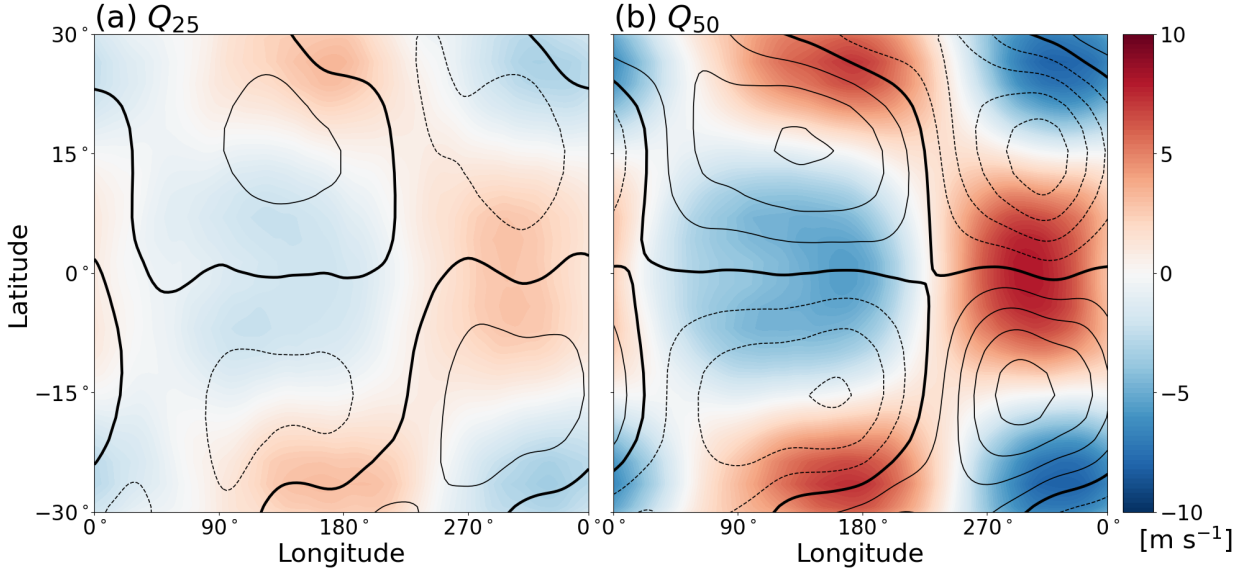


FIG. 13: 240-hPa eddy streamfunction (contours) and u^* (shading) for the (a) Q_{25} and (b) Q_{50} experiments. Contour lines are in increments of $2 \times 10^6 \text{ m}^2 \text{ s}^{-1}$, with thin solid (dotted) lines representing positive/clockwise (negative/counter-clockwise) rotation, and thick solid lines representing zero.

process occurs where the background zonal wind matches the phase speed of the waves at the critical latitude (Randel and Held 1991; Held 2000, 2019). Over the WP, the poleward shift of the westerlies and critical latitudes acts as a buffer that shields the deep tropics from eddy momentum fluxes originating from the extratropics, thus weakening the local circulation, whereas the opposite occurs over the CP (Fig. 14). While we approach the problem by examining how these eddies force the divergent circulation, this perspective aligns with previous work on reanalysis data, which demonstrated that tropical eddy momentum transport arises primarily from correlations between zonal anomalies in the LMC and upper-tropospheric Rossby gyres (Zurita-Gotor 2019).

While these transient eddy dynamics modulate the local circulation, the primary driver of the LMC is the Bernoulli function. Figure 15 presents the changes in the 240-hPa Bernoulli function (ΔB), which in the deep tropics is dominated by the changes in the geopotential ($\Delta \Phi$). Its spatial variations are mainly in the zonal direction, consistent with the large magnitude of $\Delta \psi_{LM}^*(B_\lambda)$ (Fig. 11). The maximum ΔB in the deep tropics is located around 150° longitude, 60° to the east of the maximum surface heating (90°). Similarly, the minimum ΔB is about 60° to the east of the maximum cooling (270°). The extrema in ΔB , as dictated by hydrostatic balance, align with the extrema of the temperature difference (Fig. 16). The vertical structure of the temperature response,

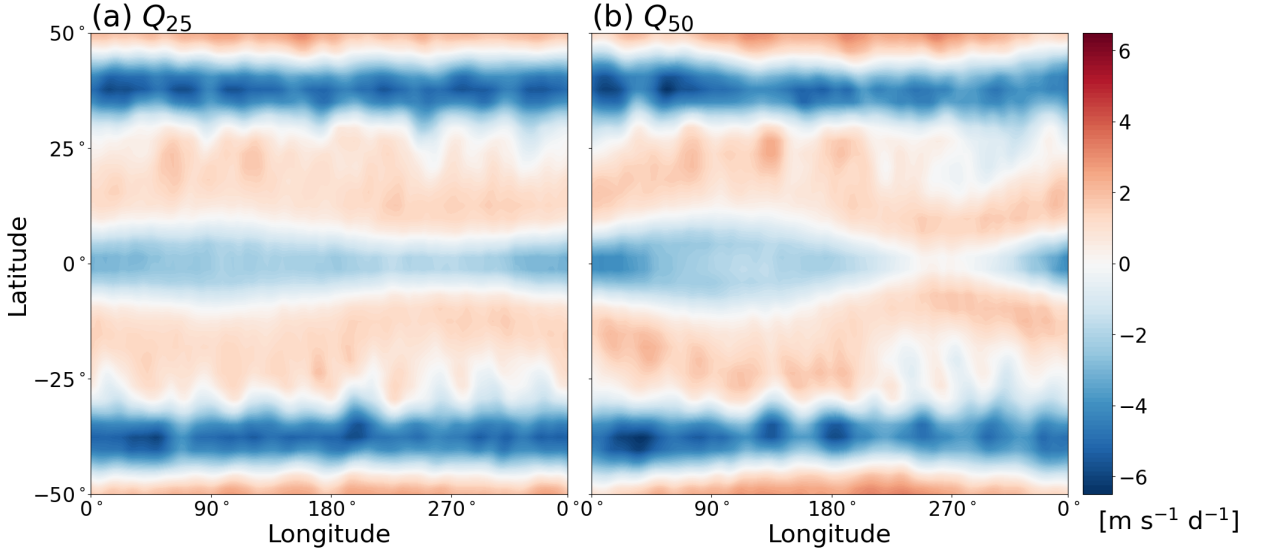


FIG. 14: 240-hPa transient eddy momentum flux divergence (S_T) for the (a) Q_{25} and (b) Q_{50} experiments.

however, is highly nonuniform. Strong zonal thermal gradients are confined to the boundary layer (below ~ 850 hPa) and the upper troposphere, while the temperature gradient in the lower free troposphere remains relatively small. Because our ZA forcings redistribute heat without altering the zonal-mean SST, this lack of bulk warming in the lower free troposphere is consistent with the finding that tropical-mean SST, rather than the maximum local SST, correlates with tropical tropospheric temperature (Sobel et al. 2002).

This vertical structure reflects two distinct mechanisms. From the top down, the transient eddy dynamics discussed previously act to modulate the thermodynamic profile over the CP: Rossby wave breaking imports cold extratropical air into the deep tropics, thereby destabilizing the column (Waugh and Polvani 2000). From the bottom up, the eastward shift of deep convection is anchored by extrema in the surface equivalent potential temperature change ($\Delta\theta_e$), which align with the upper-level extrema in ΔB and the temperature difference (not shown). This vertical alignment is consistent with quasi-equilibrium (QE) control on deep convection (Emanuel et al. 1994). Because surface temperatures remain tightly coupled to the prescribed ocean heat flux, the further eastward shift of $\Delta\theta_e$ is driven by specific humidity changes, likely tied to variations in boundary layer depth. Notably, this out-of-phase relationship between deep convection and maximum heating also emerges in aquaplanet simulations with zonal wavenumber-1 SST anomalies across multiple

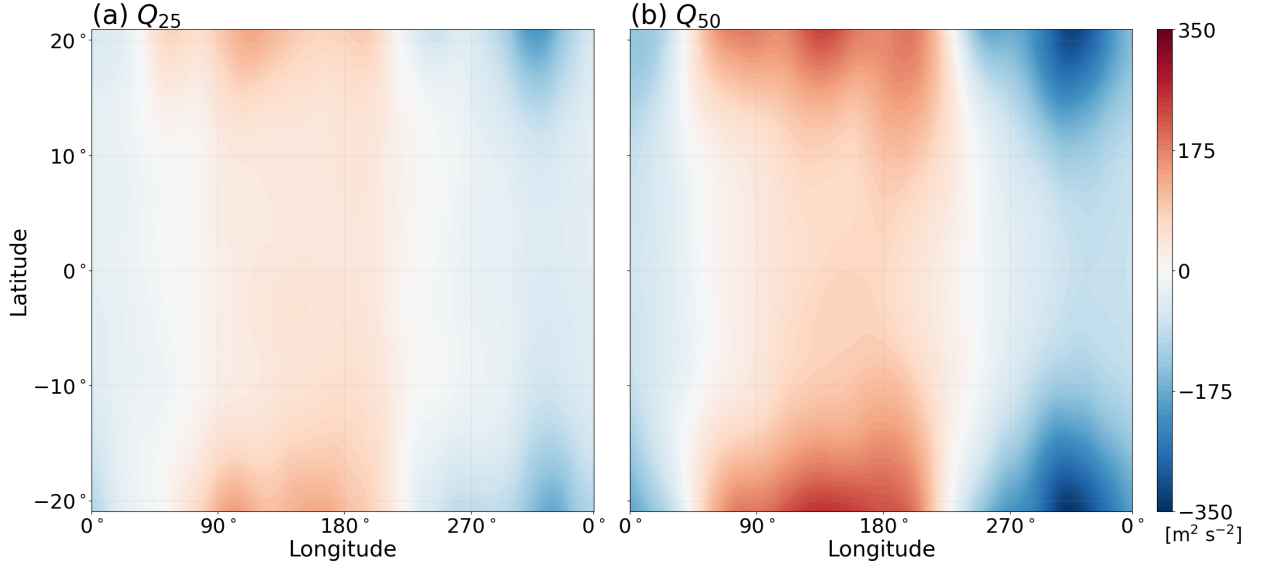


FIG. 15: 240-hPa Bernoulli function (B_λ) for the (a) Q_{25} and (b) Q_{50} experiments.

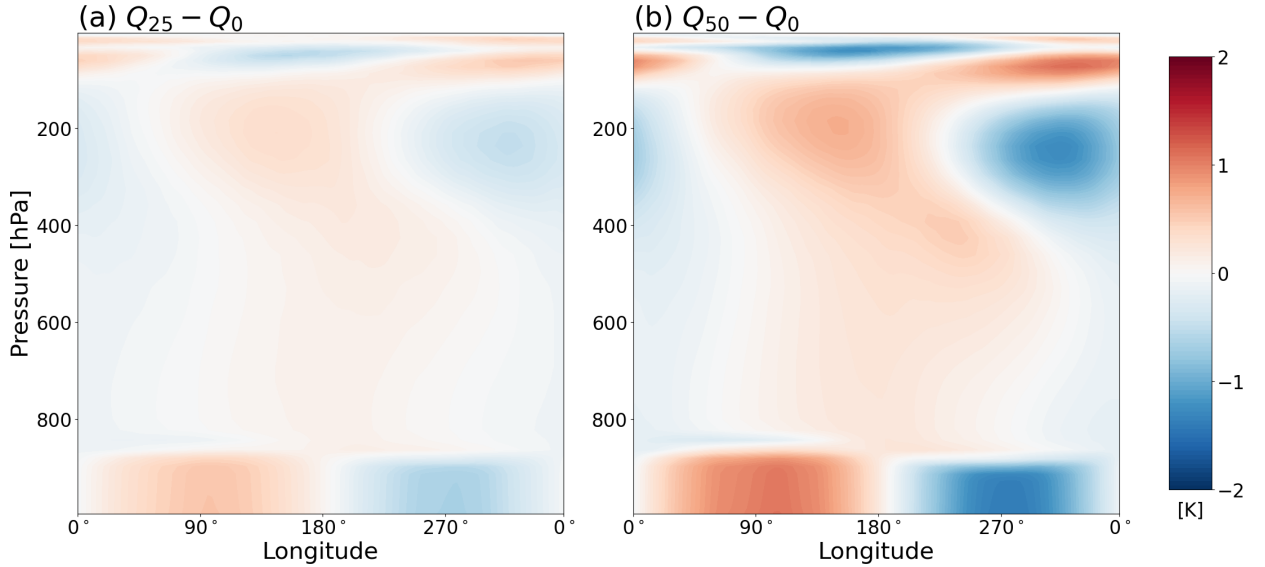


FIG. 16: Differences in deep tropical temperature (meridionally averaged from -2° to 2°) (a) $(Q_{25} - Q_0)$, and (b) $(Q_{50} - Q_0)$.

449 high-resolution GCM configurations, including one without parameterized convection (Lin et al.
450 2023; Clark et al. 2024).

451 Figure 17 summarizes all the processes discussed above in a schematic. The surface tempera-
452 ture anomalies in the ZA experiments generally give rise to more (less) precipitation or diabatic
453 heating over the WP (CP). The resulting Rossby waves move the off-equatorial upper-tropospheric

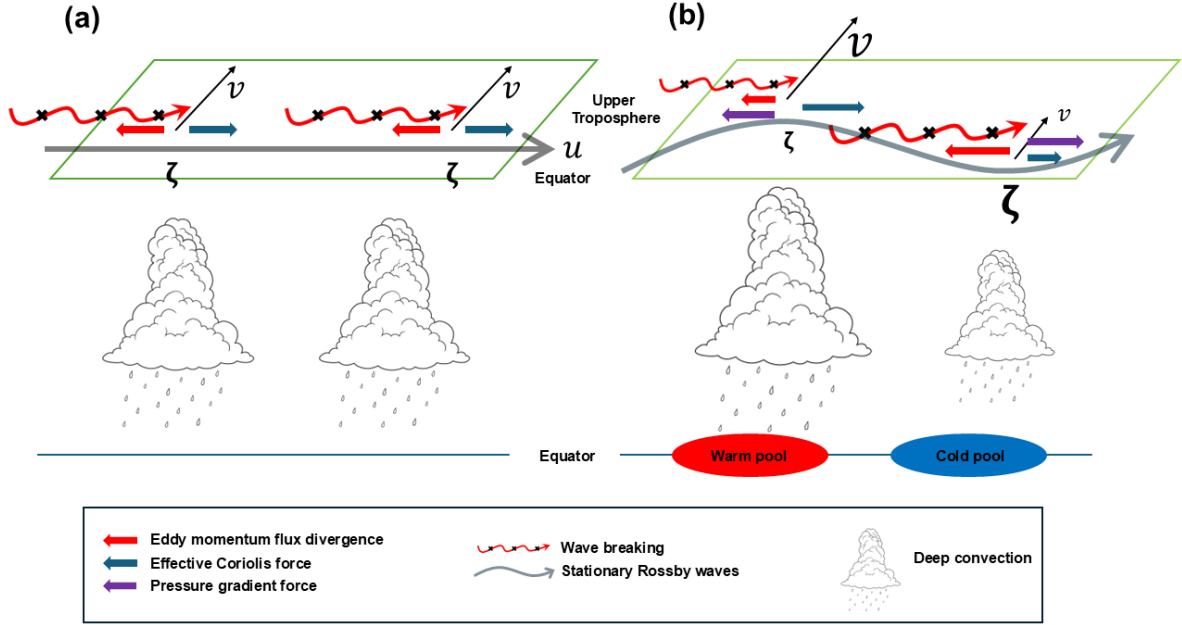


FIG. 17: Schematic of the zonal momentum balance and underlying processes in (a) the zonally symmetric case (Q_0), and (b) the zonally asymmetric cases (Q_{25} and Q_{50}), as is the case in the Northern Hemisphere.

westerlies and associated wave breaking poleward (equatorward) over the WP (CP). Relative to the zonally symmetric case, the eddy momentum flux divergence (S_T) is decreased (increased) over the WP (CP). The presence of Rossby waves also help explain the lower (higher) relative vorticity (ζ) over the WP (CP). Deep convection is located to the east of the center of the WP, and the resulting pressure gradient force acts on the WP and CP in opposite directions. Given the substantial decrease in S_T , and a slight decrease in ζ , the WP would experience a major LMC slowdown. This does not manifest because the westward PGF induces poleward flow and accelerates the LMC. Conversely, the LMC over the CP would speed up if the local increase in S_T acts alone, but the eastward PGF and a major increase in ζ are able to reverse the S_T effect, and lead to a net slowdown.

While we have presented a framework for understanding the dynamics of the LMC, it remains only one piece of the tropical circulation. In our experiments, the impacts of both the divergent zonal circulation and the rotational meridional wind are almost negligible. As is evident from Eqs. (12) and (13) and noted in the literature, length scales are crucial in determining the magnitude of the divergent zonal and meridional winds (Schwendike et al. 2014; Zurita-Gotor 2019). These length scales similarly modulate the rotational winds. Despite our forcings prescribing a large

zonal length scale, the results presented are still representative of trends found in observational reanalysis. Zurita-Gotor (2019) found that the WC has a negligible effect on the eddy momentum transport within the tropics. Complementarily, Schwendike et al. (2014) found that ω_Z is roughly an order of magnitude weaker than ω_M in observations. These results paint a consistent picture that brings into question the WC's role in the tropical climate, naturally inviting a reevaluation of how the three-dimensional tropical circulation is traditionally partitioned.

5. Summary and Conclusions

This study elucidates the dynamical changes that arise in the meridional circulation from zonally anomalous forcings. By perturbing a zonally symmetric control simulation with prescribed wavenumber-1 ocean heat fluxes, we find that the introduction of a relatively weak Walker circulation is concurrent with a significant weakening of the Hadley circulation. By modifying standard zonal-mean theories of the Hadley circulation to retain the full absolute vorticity profiles, we show that the changes in the zonal-mean relative vorticity are responsible for the weakening. We demonstrate that this Hadley circulation weakening can be accurately diagnosed in dynamical regimes where neither the eddy nor energetic constraints dominate. Furthermore, we find that the local meridional circulation responses vary with longitude: a strengthening over the warm pool is accompanied by a weakening over the cold pool. The Hadley circulation slowdown is the net effect after the cancellation of the disparate regional changes. Moreover, even in the presence of zonal asymmetries, local vertical motion remains overwhelmingly controlled by the meridional circulation rather than its zonal counterpart, highlighting the central role of the local meridional circulation in setting local climates.

Given this preeminent role of the local meridional circulation, it is crucial to unpack its dynamical drivers. To this end, we developed a diagnostic framework based on the upper-tropospheric zonal momentum balance. The meridional wind is dictated by a balance between the eddy momentum flux divergence and pressure gradient force, both modulated by the local absolute vorticity. The pressure gradient force drives an antisymmetric response (i.e., strengthening the meridional circulation over the warm pool and weakening it over the cold pool), which is countered by the eddy momentum flux divergence. The increased relative vorticity is a major factor in weakening the meridional circulation over the cold pool, consistent with the findings of the zonal-mean theory. This new

498 framework provides a basis for understanding the changes in the off-equatorial upper-tropospheric
499 meridional wind in response to a variety of perturbations.

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 501 of the Andromeda cluster made available by Boston College Research Computing / Information
 502 Technology Services.

503 *Data availability statement.* The numerical experiments performed in this study were carried
 504 out using ISCA’s Model of an Idealized Moist Atmosphere (MiMA). ISCA can be accessed at
 505 <https://execlim.github.io/IscaWebsite/>. A very similar modeling suite is accessible through the
 506 GFDL’s Flexible Modeling System (FMS) project at <https://www.gfdl.noaa.gov/fms/>.

507 APPENDIX A

508 Definitions of Local, Regional and Classical Streamfunctions

509 The local meridional and zonal mass streamfunctions (ψ_{LM} and ψ_{LZ} , respectively) are defined
 510 as:

$$\psi_{LM}(\lambda, \phi, p) = \frac{1}{g} \int_0^p \bar{v}_{\text{div}}(\lambda, \phi, p') dp', \quad (\text{A1})$$

$$\psi_{LZ}(\lambda, \phi, p) = \frac{1}{g} \int_0^p \bar{u}_{\text{div}}(\lambda, \phi, p') dp', \quad (\text{A2})$$

511 where u_{div} and v_{div} are the divergent zonal and meridional wind, respectively, based on the
 512 Helmholtz decomposition. The local streamfunctions (ψ_{LM} and ψ_{LZ}) can be interpreted as the
 513 mass flow in the meridional or zonal direction per unit length.

514 To explicitly connect ψ_{LM} and ψ_{LZ} to local vertical motion, we define the meridional and zonal
 515 components of the pressure velocity ($\bar{\omega}_M$ and $\bar{\omega}_Z$) as:

$$\bar{\omega}_M = -\frac{g}{a \cos \phi} \frac{\partial(\psi_{LM} \cos \phi)}{\partial \phi}, \quad (\text{A3})$$

$$\bar{\omega}_Z = -\frac{g}{a \cos \phi} \frac{\partial \psi_{LZ}}{\partial \lambda}. \quad (\text{A4})$$

516 It is straightforward to show that the summation of $\bar{\omega}_M$ and $\bar{\omega}_Z$ is equal to the total ω , and thus
 517 satisfied mass continuity:

$$\begin{aligned}\frac{\partial \bar{\omega}_M}{\partial p} + \frac{\partial \bar{\omega}_Z}{\partial p} &= -\nabla_p \cdot \bar{\mathbf{u}} \\ &= -\nabla_p \cdot \bar{\mathbf{u}}_{\text{div}},\end{aligned}\tag{A5}$$

518 where $\mathbf{u}$ is the horizontal wind vector (u, v) , $\mathbf{u}_{\text{div}}$ the divergent wind vector $(u_{\text{div}}, v_{\text{div}})$, and $\nabla_p \cdot$
 519 the divergence operator along a pressure surface. Specifically, $\bar{\omega}_M$ and $\bar{\omega}_Z$ correspond to the pure
 520 divergences of v and u , respectively.

521 It is important to distinguish ω_M and ω_Z from $[\omega]$ and ω^* , where $[\omega]$ is the zonal mean (ZM)
 522 pressure velocity and ω^* is the deviation from the ZM (i.e., the zonally asymmetric component).
 523 Note that ω^* has been used to represent the Walker circulation in previous studies (e.g., Vecchi and
 524 Soden 2007; Merlis and Schneider 2011). By definition, total vertical velocity can be decomposed
 525 as either $\omega = \omega_M + \omega_Z$ or $\omega = [\omega] + \omega^*$. Because the ZM of the vertical velocity is by definition
 526 driven by the ZM meridional circulation ($[\omega] = [\omega_M]$), equating the two decompositions yields
 527 $\omega_M + \omega_Z = [\omega_M] + \omega^*$. Since the local meridional circulation varies with longitude in a zonally
 528 asymmetric climate ($\omega_M(\lambda) \neq [\omega_M]$), it follows that $\omega_Z \neq \omega^*$. Thus, one should not equate the
 529 zonal divergent component (ω_Z) with the zonally asymmetric component (ω^*).

530 The regional meridional and zonal streamfunctions (ψ_{RM} and ψ_{RZ} , respectively) are integrals of
 531 their local counterparts over longitudinal and latitudinal bands, respectively:

$$\psi_{RM}(\phi, p) = a \cos \phi \int_{\lambda_1}^{\lambda_2} \psi_{LM}(\lambda, \phi, p) d\lambda,\tag{A6}$$

$$\psi_{RZ}(\lambda, p) = a \int_{\phi_1}^{\phi_2} \psi_{LZ}(\lambda, \phi, p) d\phi.\tag{A7}$$

532 where λ_1 and λ_2 denote a longitudinal band, and ϕ_1 and ϕ_2 a latitudinal band. While ψ_{RZ} is not
 533 explicitly referred to in this study, the classical zonal streamfunction referring to the WC is, by
 534 definition, a regional streamfunction.

535 The regional streamfunctions become the classical streamfunctions (ψ_{CM} and ψ_{CZ} if the bound-
 536 aries are taken to be $\lambda_1 = 0^\circ$ to $\lambda_2 = 360^\circ$ and $\phi_1 = -5^\circ$ to $\phi_2 = 5^\circ$ for the meridional and zonal

537 regional streamfunctions, respectively. They can also be written explicitly as:

$$\begin{aligned}\psi_{CM}(\phi, p) &= \frac{2\pi a \cos \phi}{g} \int_0^p [\bar{v}_{\text{div}}](\phi, p') dp' \\ &= \frac{2\pi a \cos \phi}{g} \int_0^p [\bar{v}](\phi, p') dp',\end{aligned}\tag{A8}$$

$$\psi_{CZ}(\lambda, p) = \frac{a\Delta\phi}{g} \int_0^p \langle \bar{u}_{\text{div}} \rangle_{\Delta\phi}(\lambda, p') dp',\tag{A9}$$

538 where $\langle \cdot \rangle_{\Delta\phi}$ denotes an average over the latitudinal band $\Delta\phi$ (e.g., -5° to 5°). ψ_{CM} corresponds to
539 the closed HC, and its contours are streamlines. This study uses ψ_{CZ} to represent the WC, which
540 is not precisely closed. Note that several studies also used $\langle \bar{u}_{\text{div}} \rangle_{\Delta\phi}$ to calculate streamfunctions
541 associated with the WC, but replaced $\Delta\phi$ in the prefactor of Eq. (A9) with 2π (Yu and Zwiers
542 2010; Bayr et al. 2014; Hartmann 2015; Eresanya and Guan 2021; Li et al. 2021).

543 APPENDIX B

544 Estimates of Local and Zonal-Mean Meridional Winds

545 The derivation starts with the inviscid zonal momentum equation in vector-invariant form:

$$\frac{\partial u}{\partial t} - (f + \zeta)v + \omega \frac{\partial u}{\partial p} = -\frac{1}{a \cos \phi} \frac{\partial B}{\partial \lambda},\tag{B1}$$

546 where f is the planetary vorticity, ζ the relative vorticity, $B = \Phi + \frac{1}{2}(u^2 + v^2)$ the Bernoulli
547 function, and Φ the geopotential.

548 To isolate the time-mean (climatological) state, we apply a temporal Reynolds decomposition
549 (e.g., $u = \bar{u} + u'$, where an overbar $\bar{\cdot}$ denotes a time-mean, and a prime $\cdot'$ denotes a deviation from
550 the time-mean) and take a time average of Eq. (B1). Assuming steady state yields:

$$-(f + \bar{\zeta})\bar{v} - \overline{\zeta'v'} + \bar{\omega} \frac{\partial \bar{u}}{\partial p} + \overline{\omega' \frac{\partial u'}{\partial p}} = -\frac{1}{2a \cos \phi} \frac{\partial}{\partial \lambda} (\overline{u'^2} + \overline{v'^2}) - \bar{B}_\lambda,\tag{B2}$$

551 where $\bar{B}_\lambda$ is the zonal gradient of the time-mean Bernoulli function.

552 We neglect the time-mean vertical advection term, as it is significant only at pressure levels
553 below the extrema of the HC in the outer tropics (Walker and Schneider 2006; Zurita-Gotor 2019).

Equation (B2) turns into:

$$(f + \bar{\zeta})\bar{v} = \bar{S}_T + \bar{B}_\lambda. \quad (\text{B3})$$

The time-mean transient eddy momentum flux divergence ($\bar{S}_T$) can be written as:

$$\bar{S}_T = \frac{1}{a \cos \phi} \frac{\partial (\overline{u' u'})}{\partial \lambda} + \frac{1}{a \cos^2 \phi} \frac{\partial (\overline{u' v'} \cos^2 \phi)}{\partial \phi} + \frac{\partial (\overline{\omega' u'})}{\partial p}. \quad (\text{B4})$$

TABLE B1: Approximate orders of magnitude of the components of the simulated time-mean relative vorticity ($\bar{\zeta}$) and zonal gradient of the time-mean Bernoulli function ($\bar{B}_\lambda$).

Component	Expression	Order of Magnitude
Meridional shear of zonal wind	$-\frac{1}{a} \frac{\partial (\bar{u} \cos \phi)}{\partial \phi}$	$\sim 10^{-5} \text{ s}^{-1}$
Zonal gradient of meridional wind	$\frac{1}{a \cos \phi} \frac{\partial \bar{v}}{\partial \lambda}$	$\sim 10^{-7} \text{ s}^{-1}$
Zonal geopotential gradient	$\frac{1}{a \cos \phi} \frac{\partial \Phi}{\partial \lambda}$	$\sim 10^{-5} \text{ m s}^{-2}$
Zonal gradient of zonal kinetic energy	$\frac{1}{a \cos \phi} \frac{\partial}{\partial \lambda} \left(\frac{\bar{u}^2}{2} \right)$	$\sim 10^{-6} \text{ m s}^{-2}$
Zonal gradient of meridional kinetic energy	$\frac{1}{a \cos \phi} \frac{\partial}{\partial \lambda} \left(\frac{\bar{v}^2}{2} \right)$	$\sim 10^{-7} \text{ m s}^{-2}$

Estimating $\bar{v}$ directly from Eq. (B3) is complicated by the fact that it also appears in the time-mean relative vorticity ($\bar{\zeta}$) and the time-mean Bernoulli gradient ($\bar{B}_\lambda$). As summarized in Table B1, standard atmospheric scaling (Vallis 2017) and diagnosis of our model simulations suggest that for both quantities, the components involving $\bar{v}$ are significantly smaller than the others. The meridional shear of the zonal wind is the main contributor to $\bar{\zeta}$, while the geopotential term dominates $\bar{B}_\lambda$. Thus, $\bar{v}$ can be estimated as:

$$\bar{v} = \frac{\bar{S}_T + \bar{B}_\lambda}{f + \bar{\zeta}}. \quad (\text{B5})$$

To make a connection with the ZM meridional wind, we apply a spatial Reynolds decomposition to the time-mean variables (e.g., $\bar{u} = [\bar{u}] + \bar{u}^*$, where a bracket $[\cdot]$ denotes a zonal-mean, and an asterisk $\cdot^*$ denotes a deviation from the zonal-mean) in Eq. (B2), which then becomes:

$$-f [\bar{v}] - [\bar{\zeta}] [\bar{v}] - [\bar{\zeta}^* \bar{v}^*] - [\bar{\zeta}' v'] + \left[\bar{\omega} \frac{\partial \bar{u}}{\partial p} \right] + \left[\overline{\omega' \frac{\partial u'}{\partial p}} \right] = - \left[\frac{1}{2a \cos \phi} \frac{\partial}{\partial \lambda} (\overline{u'^2} + \overline{v'^2}) \right]. \quad (\text{B6})$$

565 Note that the zonal-mean Bernoulli gradient ($[\bar{B}_\lambda]$) is zero by definition. We group all transient
 566 (temporal) covariances into a transient eddy momentum flux divergence term ($[\bar{S}_T]$), and all
 567 stationary (spatial) covariances into a stationary eddy momentum flux divergence term ($[\bar{S}_S]$):

$$[\bar{S}_T] = \frac{1}{a \cos^2 \phi} \frac{\partial \left([\overline{u'v'}] \cos^2 \phi \right)}{\partial \phi} + \frac{\partial \left([\overline{\omega'u'}] \right)}{\partial p}, \quad (\text{B7})$$

$$[\bar{S}_S] = \frac{1}{a \cos^2 \phi} \frac{\partial \left([\overline{u^*v^*}] \cos^2 \phi \right)}{\partial \phi} + \frac{\partial \left([\overline{\omega^*u^*}] \right)}{\partial p}. \quad (\text{B8})$$

568 The ZM meridional wind ($[\bar{v}]$) can be then be approximated as:

$$[\bar{v}] = \frac{[\bar{S}_T] + [\bar{S}_S]}{f + [\bar{\zeta}]}. \quad (\text{B9})$$

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