Wind Spatial Structure Triggers ENSO’s Oceanic Warm Water Volume Changes

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ABSTRACT: This study demonstrates that the generalization that strong anomalous equatorial Pacific westerly (easterly) winds during El Niño (La Niña) events display strong adjusted warm water volume (WWV) discharges (recharges) is often incorrect. Using ocean model simulations, we categorize the oceanic adjusted responses to strong anomalous equatorial winds into two categories: (i) transitioning (consistent with the above generalization) and (ii) neutral adjusted responses (with negligible WWV recharge and discharge). During the 1980–2016 period only 47% of strong anomalous equatorial winds are followed by transitioning adjusted responses, while the remaining are followed by neutral adjusted responses. Moreover, 55% (only 30%) of the strongest winds lead to transitioning adjusted responses during the pre-2000 (post-2000) period in agreement with the previously reported post-2000 decline of WWV lead time to El Niño–Southern Oscillation (ENSO) events. The prominent neutral adjusted WWV response is shown to be largely excited by anomalous wind stress forcing with a weaker curl (on average consistent with a higher ratio of off-equatorial to equatorial wind events) and weaker Rossby wave projection than the transitioning adjusted response. We also identify a prominent ENSO phase asymmetry where strong anomalous equatorial westerly winds (i.e., El Niño events) are roughly 1.6 times more likely to strongly discharge WWV than strong anomalous equatorial easterly winds (i.e., La Niña events) are to strongly recharge WWV. This ENSO phase asymmetry may be added to the list of mechanisms proposed to explain why El Niño events have a stronger tendency to be followed by La Niña events than vice versa.

KEYWORDS: ENSO; Asymmetry; Warm water volume; Wind Stress curl; Kelvin waves; Ocean dynamics; Rossby waves

1. Introduction

The tropical Pacific region provides unique conditions for the largest interannual climate fluctuation on our planet: El Niño–Southern Oscillation (ENSO). ENSO describes a coupled atmosphere–ocean phenomenon occurring every 2–7 years. It fluctuates between (i) anomalous warm states of equatorial eastern Pacific sea surface temperature (SST) accompanied by a weakening or a reversal of trade winds (El Niño), and (ii) anomalous cool states of equatorial eastern Pacific SST accompanied by a strengthening of trade winds (La Niña). Around four decades of intense research has tremendously enhanced our understanding of ENSO (e.g., Timmermann et al. 2018); however, we still have difficulties in unscrambling the subtleties of ENSO dynamics, and precisely predicting the ENSO state (e.g., Timmermann et al. 2018; Santoso et al. 2019). Enhancing the predictability of ENSO, which is expected to be partially reliant on enhancing our understanding of its dynamics, is sought after as the events are responsible for severe weather events and climatic conditions that affect societies around the globe (e.g., Diaz et al. 2001; McPhaden et al. 2006; Werner and Holbrook 2011; Loughran et al. 2017). Furthermore, CMIP5 models indicate a strengthening of these severe climatic events connected to ENSO under future climate change scenarios (Power et al. 2013; Cai et al. 2015; Perry et al. 2017; Power and Delage 2018).

Traditional conceptual models of ENSO, which were developed in the 1980s–1990s, are milestones in ENSO research that highlight our understanding of this phenomenon by summarizing its important mechanisms and feedbacks [Suarez and Schopf 1988; Weisberg and Wang 1997; Picaut et al. 1997; Jin 1997; summarized by Wang and Picaut (2004)]. These models are presented as several ordinary differential equations that solve for the eastern Pacific’s SST, and they describe ENSO as a self-sustained oscillation. In each of these models the growth of ENSO events is produced by the Bjerknes feedback (Bjerknes 1969), in which westerly (easterly) wind anomalies are generated due to the eastward extension (westward contraction) of the equatorial warm pool and its related shifts in anomalous deep convection that act to reinforce the original anomalously warm (cool) SSTs. However, each model suggests that different negative feedbacks are responsible for the end of the events, and due to the numerous assumptions and approximations, none of these conceptual models can fully describe ENSO’s complexity.

A leading paradigm for ENSO’s apparent oscillatory behavior is Jin’s (1997) recharge–discharge oscillator (RDO)
theory. According to this theory, during El Niño (La Niña) events, the Bjerknes feedback (Bjerknes 1969) causes equatorial westerly (easterly) wind anomalies and positive (negative) eastern equatorial Pacific SST anomalies to reinforce each other, at the same time as the oceanic adjustment to the westerly (easterly) wind anomalies acts to discharge (recharge) equatorial upper ocean heat content via diverging (converging) Sverdrup transports, conditioning the system for a change in phase of ENSO [see also Wang and Picaut (2004) for a comprehensive description of the RDO theory]. The main ideas of Jin’s (1997) theory have been confirmed by observational studies showing that the warm water volume (WWV; i.e., the thermocline depth integrated over 120°–280°E, 5°N–5°S), a variable describing the upper ocean heat content, is leading ENSO SST by 1–3 seasons (Meinen and McPhaden 2000; Meinen and McPhaden 2001; McPhaden 2012; Bunge and Clarke 2014). Consequently, the WWV is viewed as a precursor to ENSO events, and is largely thought to underlie its predictability.

Some aspects of ENSO, however, are not explained by the RDO theory such as (ii) the asymmetry between the magnitude of the WWV recharge preceding El Niño events compared to the magnitude of WWV discharge preceding La Niña events (e.g., Meinen and McPhaden 2000; Clarke and Zhang 2019); (ii) the asymmetry in duration and phase change transition with El Niño events lasting a shorter period and having a stronger tendency to be followed by La Niña events than vice versa (e.g., Kessler 2002; Larkin and Harrison 2002; Okumura and Deser 2010; Guan et al. 2019; Clarke and Zhang 2019); 3) the decrease observed in WWV/ENSO SST lead time from around three seasons in the 1980s and 1990s to only one season during the post-2000 period (McPhaden 2012; Horii et al. 2012; Bunge and Clarke 2014); and 4) the existence of different ENSO types: central Pacific (CP; with maximum SST anomaly found in the central equatorial Pacific) versus eastern Pacific (EP; with maximum SST anomaly found in the eastern equatorial Pacific) ENSO events (e.g., Kao and Yu 2009; McPhaden et al. 2011; Singh and Delcroix 2013). Moreover, the RDO theory considers only the adjusted oceanic response, which occurs when wind-forced equatorial Kelvin waves (KWs) have left the WWV region via the eastern boundary and wind-forced off-equatorial Rossby waves (RWs) reflect at the western boundary into the WWV region [for a more detailed understanding of the connection between Sverdrup transport as considered in Jin’s (1997) RDO theory and wave dynamics as considered in this study, see section 1.5.2 in Neske (2019)]. However, the instantaneous effect of wind-forced KWs that have not yet met the eastern boundary has also been shown to provide important short time scale (around 1–3 months) contributions to observed WWV changes determining ENSO event initiation and/or evolution (Weisberg and Wang 1997; McPhaden and Yu 1999; Boulanger et al. 2003; Bosc and Delcroix 2008; McGregor et al. 2016; Neske and McGregor 2018; Izumo et al. 2019).

Several studies have attempted to separate the WWV anomaly into its slower RW and faster KW components using various methods (Meinen and McPhaden 2000; Neske and McGregor 2018; Planton et al. 2018; Izumo et al. 2019). This partitioning allows more insight into ENSO dynamics that is not possible when considering the WWV anomaly as a whole. For example, Neske and McGregor (2018) demonstrate that the post-2000 WWV STD (standard deviation) decline is due to a 44% decline of post-2000 RW-driven (hereafter called adjusted) WWV STD, and in this post-2000 period the reduced adjusted contribution fails to work as a precursor for ENSO. Consequently, Neske and McGregor (2018) show that the dominance of the KW-driven (hereafter called instantaneous) WWV in the post-2000 period is responsible for the shortening of WWV/ENSO SST lead time reported by McPhaden (2012).

It is currently not understood, however, why the STD of the adjusted component of WWV decreased by around 44% from the pre- to post-2000 period, when over the same time period the STD of the instantaneous component of WWV increased by around 15% (Neske and McGregor 2018). Traditional ENSO theories assume that the strength of the anomalous equatorial westerly (easterly) winds was directly related to the strength of the upwelling (downwelling) RWs and its associated following WWV discharge (recharge) (e.g., Suarez and Schopf 1988; Jin 1997). Thus, we would expect that the post-2000 strengthening in the instantaneous contribution’s STD (the instantaneous contribution is highly correlated to equatorial wind stress; Fig. 2a) would lead to a strengthening in the adjusted response STD as well and not to a decline in the adjusted response STD as found by Neske and McGregor (2018).

What is not considered in most of the literature, and we propose may be key in explaining the post-2000 adjusted WWV decline, is the possible independence of the wind stress curl from the strength of equatorial wind stress. The wind stress curl around the equatorial region (i) changes due to the spatial structure and central latitude of the wind events and (ii) determines the strength and sign of the RWs through Ekman pumping (e.g., Tomczak and Godfrey 2013). McGregor et al. (2016) show, for instance, that the same amount of instantaneous WWV recharge leads to a lower adjusted response for off-equatorial wind events than for equatorial wind events as both produce a different wind stress curl around the WWV region.

An emphasis on considering wind stress curl for ENSO dynamics was given by Clarke et al. (2007), who theoretically derived WWV changes by wind stress curl at 5° latitude, which has a correlation of 0.85 to observed WWV changes between 1993 and 2001. However, when later developing their conceptual model of ENSO, Clarke et al. (2007) instead utilize the relationship between wind stress curl derived WWV and Niño-3.4 SST (correlation = 0.77) rather than wind stress curl itself. Thus, the developed conceptual model leaves out the possibility of strong equatorial wind stresses during ENSO events with weak wind stress curls around the boundaries of the WWV region. While this may not be essential for ENSO to operate (Clarke et al. 2007; Zhu et al. 2017), it may help to explain both the relatively subtle pre- to post-2000 changes in the WWV–ENSO SSTA relationship as well as ENSO asymmetries.

Our paper therefore aims to understand to what extent the spatial structure of the strongest westerly (easterly) equatorial wind stresses in the 1980–2016 period can influence the adjusted contribution. Given that the adjusted contribution is thought to underpin ENSO predictability on lead times of around 1 year (Neske and McGregor 2018) our analysis further
deeps our understanding of possibilities and limits of ENSO predictability.

This paper is organized as follows: section 2 provides a description of methods and data. Section 3 classifies the adjusted responses following the strongest equatorial wind stresses into two different categories. We then analyze (i) the differences of the SST and wind composites according to their category (section 4) and (ii) the WWV impact of the composite winds (section 5). The links between composite average winds and single wind events within the composite are then examined in section 6. Finally, we summarize and discuss the results in section 7.

2. Methods and data

a. Wind stress and SST data

1) WIND STRESS DATA

We use daily average anomalous wind stresses calculated from the 6-hourly 10-m surface winds (1979 to May 2016) of the European Centre of Medium-Range Weather Forecast (ERA-Interim) (Dee and Uppala 2009). The winds were first converted to wind stress using the quadratic stress law: \( \tau^x, \tau^y = \rho_C a (U_{10, t} + V_{10, t}) \), with \( \tau^x \) and \( \tau^y \) being zonal and meridional wind stresses, \( \rho_C = 1.2 \text{ kg m}^{-3} \) being the air density, \( a = 1.5 \times 10^{-3} \) being a dimensionless drag coefficient, and \( U_{10} \) and \( V_{10} \) being the zonal and meridional winds at 10-m height. Afterward the wind stresses are averaged to a daily output, prior to the long-term seasonal cycle and linear trend being removed. Throughout this paper, the term “wind stress” relates to the processed wind stress anomalies described in this section.

We further define the equatorial zonal wind stress time series averaged over 120°–280°E, 5°N–5°S as \( \tau_{eq} \).

2) WIND STRESS CURL

The curl of the horizontal wind stress is defined as \( \text{curl}(\tau) = (\partial \tau^y / \partial x) - (\partial \tau^x / \partial y) \). As the north–south scale of the wind in the equatorial region is much smaller than the east–west scale (e.g., Clarke 2008) we have chosen to neglect the first term of the wind stress curl throughout the paper. This choice is justified by the fact that the contribution of the \( \tau^y \) forcing to the control WWV shallow water model outcome (see section 2b for a definition of the control WWV) is negligible (Fig. S1 in the online supplemental material).

3) SST DATA

Additionally, we use the spatial ERSST.v5 monthly dataset between 1980 and 2016 (Smith et al. 2008, available at https://www1.ncdc.noaa.gov/pub/data/cmb/ersst/v5/netcdf/). From this dataset the Niño-3.4 SST time series is calculated by averaging over 190°–240°E, 5°N–5°S.

b. Shallow-water model simulations

Here, as in Neske and McGregor (2018), a 1.5-layer shallow water anomaly model (SWM) is utilized to split the WWV into its instantaneous and adjusted contributions. The SWM has a 1° horizontal grid spacing [the Arakawa C grid of Mesinger and Arakawa (1976)], while the model time step is 2 h, where Fischer (1965) numerical scheme is utilized for model time stepping. The SWM is forced with daily wind stress anomalies and the produced daily pycnocline depth anomalies are integrated over the WWV region (120°–280°E, 5°N–5°S) to obtain WWV anomalies. Western boundary reflection is governed by the conservation of mass. Further details about the SWM can be found in Neske and McGregor (2018, their Text S1).

The separation of the adjusted and instantaneous WWV contributions is done in the SWM by conducting hindcast experiments initialized on the first day of each month between January 1980 and March 2016 (as illustrated in Fig. 1a). Each of these hindcasts has a wind stress–forced spinup of one year before the SWM is left to freely evolve unforced (dashed line, Fig. 1a) for three months, whereas for the control simulation the model is kept forced over these three months (black circle, Fig. 1a). The monthly mean control WWV changes are consistent with those observed (Smith 1995; available at https://www.pmel.noaa.gov/tao/wwv/data/), as indicated by a correlation coefficient of 0.86 and a similar pre- to post-2000 STD decline (Fig. 1b).

Note that the correlation between observed and modeled WWV is strongest when forcing the SWM with the ERA-Interim wind product, as chosen in this paper, as compared to three other wind products (JRA-55, NCEP1, and NCEP2) utilizing the wind stress calculation described above (Fig. A2.4 in Neske 2019). Here, we also examine the SWM response to TropFlux daily wind stress forcing (available at https://incois.gov.in/tropflux/), which uses (i) bias-corrected ERA-Interim and ISCCP data as input and (ii) bulk estimates of air–sea fluxes to compute wind stresses (Kumar et al. 2013). We find SWM WWV differences that we consider negligible between the TropFlux and ERA-interim forced simulations (Fig. S2; correlation = 0.99). Thus, ERA-Interim wind stresses are retained for the analysis of this paper.

Interestingly, none of the SWM simulations forced by the aforementioned various wind products is able to reproduce all of the strongest observed WWV changes [e.g., the strong WWV trough of 1997/98 in Fig. 1b herein and in Fig. A2.4 in Neske (2019)]. One might contribute this to the simplicity of the SWM; however, a similar WWV underestimate results when forcing a more comprehensive global ocean–sea ice model with the ERA-Interim wind stresses (Fig. 3.1b in Neske 2019). Thus, we suspect that this discrepancy between the modeled control WWV and observed WWV is due to either (i) biases in wind strength and/or wind spatial structure that are apparent in all wind products or (ii) surface heat fluxes not utilized in the SWM and global ocean–sea ice model simulations that have been shown to significantly influence WWV (e.g., Huguenin et al. 2020). As our paper focuses on the observed pre- to post-2000 decline in WWV STD, and as this decline is reproduced in the SWM control simulation WWV, we believe that it is appropriate to use the SWM as a basis for our further analysis.

1) ADJUSTED CONTRIBUTION

The adjusted contribution (Fig. 1c) is the WWV anomaly after the SWM has been forced for 1 year and then it is run freely (without wind stress forcing) for three months (Fig. 1a). It largely consists of RWs reflected at the western boundary into the WWV region and some wind initiated RWs. This is because during the three months of free evolution (i) KWs reach the eastern Pacific boundary (within 1–3 months) where
most of their signal is taken poleward out of the equatorial region (McGregor et al. 2016) and (ii) RWs forced in the western/central Pacific (i.e., where the strongest coupling between atmosphere and ocean is found; Deser and Wallace 1990) typically reach the western boundary within three months where they reflect into the WWV region. The adjusted response has a peak correlation of 0.80 to the observed western Pacific (120°–205°E) WWV anomaly (WWVwest, taken from https://www.pmel.noaa.gov/elnino/upper-ocean-heat-content-and-enso) at a 3-month lag (Fig. 1d). The 3-month lag and the larger amplitude of WWVwest partially arise from the adjusted SWM run using a 3-month-long free evolution to mostly separate the reflected RW signal, whereas the WWVwest additionally contains the RW and KW signals directly forced in the western WWV region (Izumo et al. 2019). Furthermore, the adjusted WWV contribution has its peak correlation of 0.75 at a 10-month lag to τeq; however, the strong decline in the adjusted WWV contribution post-2000 STD (of 44%) is not found in τeq (cf. Figs. 1c and 2b).

2) INSTANTANEOUS CONTRIBUTION

The instantaneous contribution is the difference between the control WWV and the adjusted contribution (Figs. 1a,c). Therefore, the instantaneous contribution in any month is driven by winds over the 3-month period including the two months prior to and the month during the instantaneous contribution (Fig. 1a). It predominantly consists of the wind-forced KW signal in the eastern Pacific largely dominating the wind-forced RW signal in the western Pacific (Neske and McGregor 2018). The instantaneous contribution shows strong consistency with WWV changes due to meridional Ekman transport anomalies at 5°N and 5°S (Neske and McGregor 2018). Additionally, the instantaneous contribution is highly correlated with τeq (correlation = 0.86; Fig. 2a) and, consistent with τeq (Fig. 2b), it shows little change in post-2000 STD (Fig. 1c). We also separate the instantaneous contribution east (west) of 200°E and define this as WWVinst. est (WWVinst.west) in order to separate the KW WWV signal in the eastern Pacific from the RW WWV signal in the western Pacific.

If not stated differently all time series are smoothed by a 3-month running mean and correlations between time series are significant according to the method for autocorrelated time series of Ebisuzaki (1997). We further note that an overview of all model simulations conducted in this paper is provided in Table S1.
3. Categorizing adjusted WWV responses

To better understand why there is a strengthening of the post-2000 instantaneous WWV signal but a weakening in the post-2000 adjusted WWV signal (Fig. 1c), we examine the different adjusted WWV responses following the strongest anomalous equatorial wind stresses (we note that $\tau_{eq}$ is highly correlated to the instantaneous WWV contribution; Fig. 2a). We identify 16 peaks (14 troughs) where the magnitude of equatorial wind stress $\tau_{eq}$ is above 1 STD (below −1 STD) of $\tau_{eq}$ (0.007 N m$^{-2}$; horizontal dashed line). The correlation is calculated between both time series. $\tau_{eq}$ (N m$^{-2}$; brown line) and $\text{curl}(\tau)$ index (N m$^{-3}$; purple line). The brown (purple) number gives the STD of the post-2000 period divided by that of the pre-2000 period of the brown (purple) time series. The correlation is calculated between both time series. Colored dots (squares) mark the $\tau_{eq}$ peaks (troughs) above 1 STD (below −1 STD) of $\tau_{eq}$ according to (c) and (d).

Simulations with adjusted responses (averaged over day 60–90) in (c) and (d) are colored (i) green when falling below (exceeding) $-0.2 \times 10^{14}$ (0.2 $\times 10^{14}$) m$^{3}$ (transitioning adjusted responses), and (ii) gray when lying between 0.2 $\times 10^{14}$ and $-0.2 \times 10^{14}$ m$^{3}$ (neutral adjusted responses).

![Fig. 2.](image)

FIG. 2. (a) Instantaneous contribution (m$^{3}$; blue line) and $\tau_{eq}$ (N m$^{-2}$; brown line). Dots (squares) mark the $\tau_{eq}$ peaks (troughs) above 1 STD (below −1 STD) of $\tau_{eq}$ (0.007 N m$^{-2}$; horizontal dashed line). The correlation is calculated between both time series. (b) $\tau_{eq}$ (N m$^{-2}$; brown line) and $\text{curl}(\tau)$ index (N m$^{-3}$; purple line). The brown (purple) number gives the STD of the post-2000 period divided by that of the pre-2000 period of the brown (purple) time series. The correlation is calculated between both time series. Colored dots (squares) mark the $\tau_{eq}$ peaks (troughs) above 1 STD (below −1 STD) of $\tau_{eq}$ according to (c) and (d).

![Westerly wind forcing](image)

(c) Westerly wind forcing

![Easterly wind forcing](image)

(d) Easterly wind forcing

The strongest adjusted WWV changes in our simulations (which are up $-0.93$ and $0.57 \times 10^{-14}$ m$^{3}$; Figs. 2c and 2d) are smaller than the extrema of the adjusted time series shown in Fig. 1c (orange line). The reason for this disparity is that the adjusted time series displayed in Fig. 1c is the SWM WWV response to 12 months of anomalous wind stress forcing prior to the model's free evolution, rather than only the 3 months utilized here (see Fig. 1a, section 2). The 12-month and 3-month forced adjusted time series, however, are highly correlated (correlation = 0.8) and both time series show a clear pre- to post-2000 decline in STD (44% and 31% for the 12-month and 3-month adjusted responses, respectively; not shown here).
It is clear that fewer than half of the simulations (14 out of 30) produce a transitioning adjusted response that is expected from theory (green dots and squares in Fig. 2b). Furthermore, the ratio of transitioning responses to neutral responses is 9:7 for the westerly wind stress forcing of the $t_{x eq}$ peaks (dots in Fig. 2b), while it is only 5:9 for the easterly wind stress forcing of the $t_{x eq}$ troughs (squares in Fig. 2b). The composite adjusted WWV discharge following the $t_{x eq}$ peak forcing (averaged over days 60–90) is around 19% stronger than the composite adjusted WWV recharge of the $t_{x eq}$ trough forcing (cf. thick black lines in Figs. 2c and 2d). This asymmetry suggests that the strong westerly $t_{x eq}$ forcing has a larger potential as a long-term precursor for La Niña events compared to the strong easterly $t_{x eq}$ forcing for El Niño events. Additionally, during the pre-2000 period around 4 times more transitioning adjusted responses are found, compared to the post-2000 period (Fig. 2b). The dominance of neutral over transitioning responses in the post-2000 period (Fig. 2b) is consistent with the post-2000 reduction of the adjusted response STD found by Neske and McGregor (2018) (Fig. 1c).

4. Composite SSTAs and winds

To understand the dynamical differences between the transitioning and neutral WWV responses, we make composites of the 3-month average observed SSTAs and zonal winds around the identified $t_{x eq}$ peaks and troughs. These composites are further stratified according to their modeled adjusted WWV response, that is, (i) transitioning responses (green composite Figs. 3b,c and 4a,b) and (ii) neutral responses (gray composite Figs. 3d,e and 4c,d) as identified in section 3 (Figs. 2c,d).

a. Composite SSTAs

As expected from the strong coupling between ocean and atmosphere in the western (central) equatorial Pacific (e.g., Deser and Wallace 1990), $t_{x eq}$ displays a strong linear relationship (correlation of 0.83) to the Niño-3.4 SSTA (Fig. 3a); consequently, the majority [81% (71%)] of the $t_{x eq}$ peaks (troughs) are found during Niño-3.4 SSTAs $>$ 0.5°C (≥ −0.5°C). (See also Tables S2 and S3 for the dates and exact Niño-3.4 values.) The transitioning adjusted responses generally show a stronger Niño-3.4 SSTA magnitude compared to the neutral responses (Fig. 3a). Comparing the composite SSTA patterns, averaged over the 3-month period around the $t_{x eq}$ peaks and troughs (Figs. 3b–e), reveals that the transitioning (neutral) composite SSTA patterns display their strongest magnitudes in the eastern (central) Pacific consistent with the SSTA of EP (CP) ENSO events (e.g., McPhaden et al. 2011). This finding is largely consistent when looking at the SSTAs during the forcing periods of each single simulation: For 72% of the transitioning $t_{x eq}$ peaks and troughs the Niño-3 SSTA magnitude (averaged over 210°–270°E, 5°N–5°S) is larger compared to the Niño-4 SSTA magnitude (averaged over 160°–210°E, 5°N–5°S). In contrast, the Niño-3 SSTA magnitude is smaller than the Niño-4 SSTA magnitude for 69% of the neutral peaks and troughs patterns.
b. Wind stress composites

When looking at the composite zonal wind stress patterns (Figs. 4a–d) it is clear that the composite wind forcing of the transitioning adjusted discharge responses displays strong anomalous equatorial westerly wind stress between around $160^\circ$ and $230^\circ$E (Figs. 4a,e). The anomalous westerly wind stress decreases poleward abruptly at around $5^\circ$N and $5^\circ$S (Figs. 4a,f), inducing strong wind stress curls at the meridional boundaries of the WWV region (Figs. 4a,g). The composite wind forcing of the transitioning adjusted recharge responses is almost the mirror image (easterly wind stress), with the exception of the winds being shifted westward by around $20^\circ$ (Figs. 4b,e–g).

Comparing the transitioning and neutral zonal wind composites of $\tau_{seq}$ peaks reveals that the neutral composite (i) has also strong equatorial winds (shifted westward, Figs. 4a,c,e), (ii) displays a smaller zonal extent (Figs. 4a,c; compare gray and green lines in Fig. 4e), and (iii) is meridionally broader (Figs. 4a,c,f). Very similar differences are also seen when comparing the transitioning and neutral zonal wind composites of $\tau_{seq}$ troughs (Figs. 4b,d–f). This broader meridional pattern of the neutral zonal wind composite leads to weaker wind stress curls around the meridional boundaries of the WWV region than those seen in the transitioning composites (Fig. 4g; also compare Fig. 4c with Fig. 4a and Fig. 4d with Fig. 4b).

To understand the differences of each single forcing period between the transitioning and neutral simulations, we define a near-equatorial wind stress curl index [curl($\tau$) index], which is the wind stress curl averaged over $120^\circ$–$280^\circ$E, $2^\circ$–$8^\circ$S minus the wind stress curl averaged over $120^\circ$–$280^\circ$E, $2^\circ$–$8^\circ$S (vertical gray lines in Fig. 4g). The latitudes $2^\circ$–$8^\circ$ are chosen as (i) the projection on the first two equatorial symmetric meridional modes (i.e., $N = 1$ and $N = 3$) is strongest here (e.g., McGregor et al. 2016, their Fig. 5a), and (ii) the western boundary reflection efficiency of RWs has been shown to be inefficient poleward of $6^\circ$ (Kessler 1991). We first note, that despite the good correlation of 0.69 between $\tau_{seq}$ and the curl($\tau$) index, the curl($\tau$) index shows a pre- to post-2000 decline in STD of 27% not found for $\tau_{seq}$ (Fig. 2b). The curl($\tau$) index does not explain the full post-2000 adjusted WWV STD decline of 44% as the curl($\tau$) index does not consider the post-2000 cancellations of RW signals excited at different regions at different times, playing a role for the adjusted response as well [chapter 4 in Neske (2019)].

Consistent with the wind stress composites of Fig. 4, the average $\tau_{seq}$ magnitude of the transitioning peaks (troughs) forcings is slightly [7% (27%)], but not statistically significantly, stronger compared to that of the neutral peaks (troughs) forcings (Fig. 5a, applying a two-sample t test at a 95% confidence interval). The magnitude of the curl($\tau$) index, however, is significantly larger during the wind stress forcings of transitioning adjusted WWV responses compared to those wind stress forcings of neutral adjusted WWV responses (on average 49% and 52% stronger for the peaks and troughs simulations respectively; Fig. 5b).

Finally, it is worth noting that the small differences in equatorial wind stress curl and curl($\tau$) index magnitude between transitioning peak and trough simulations are not statistically significant [10% (3%) difference for the wind stress (curl) in Figs. 4a and 4b]. The same is true for the small differences between neutral peak and trough simulations [19% (9%) difference for the wind stress (curl) in Figs. 4c and 4d].
stronger average adjusted WWV change (discharge) following \( \tau_{eq} \) peaks relative to the average adjusted WWV change (recharge) following \( \tau_{eq} \) troughs (thick black lines in Figs. 2c,d) appears to be largely due to the more frequent occurrence of transitioning categories during the peaks compared to the troughs (Fig. 2b), rather than due to differences between peaks and troughs within the categories.

5. Understanding the warm water volume response

To get a better understanding of the WWV response of the different categories of adjusted WWV responses, we carry out another set of SWM experiments. To this end, one SWM simulation is carried out for each category of adjusted response, and similar to the previously utilized experiments, each SWM simulation is forced over three months before being allowed to freely evolve (i.e., with no wind forcing) for nine months. In this case, however, the simulation is forced with the composite mean zonal wind stress of each of the four categories of adjusted WWV response (Figs. 4a–d). The pycnocline outcome is averaged over (i) the last month of the forcing period (days \(-30\) to \(0\)), which gives the instantaneous response (Figs. 6a,d and 7a,d); (ii) the third month of the free evolution (days \(60\) to \(90\)), which gives the adjusted response (Figs. 6b,e and 7b,e); and (iii) the eighth month of the free evolution (days \(210\) to \(240\)), which gives the late adjusted response (Figs. 6c,f and 7c,f).

a. \( \tau_{eq} \) peaks

The equatorial region westerly wind stresses of the transitioning and neutral forcing composites (Figs. 4a,c) excite

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**Fig. 5.** Probability distribution of the magnitude of (a) \( \tau_{eq} \) (N m\(^{-2}\)), and (b) \( \text{curl}(\tau) \) index (N m\(^{-3}\)) of the 3-month forcing periods of the equatorial westerly and easterly transitioning (green bars) and neutral (gray bars) adjusted response as identified in Figs. 2c and 2d.

**Fig. 6.** Pycnocline anomaly output (m) averaged over days (a), (d) \(-30\) to \(0\) (instantaneous response), (b), (e) \(60\) to \(90\) (adjusted response), and (c), (f) \(210\) to \(240\) (late adjusted response) since the free evolution of the \( \tau_{eq} \) peaks SWM simulations. The SWM simulations are forced over 3 months by the (top) adjusted and (bottom) neutral wind stress composite as shown in Figs. 4a and 4c. Boxes mark the WWV region, and the vertical line in (a) and (d) divides the WWV region into east and west of \(200^\circ\)E.
downwelling KWs and an increase in WWV$_{\text{inst.east}}$ magnitude that does not differ between the two categories (Figs. 6a,d). This is expected, as the distribution of $\tau_{\text{eq}}$ magnitude exciting KWs is similar in both the transitioning and neutral composite forcings (Fig. 5a). In contrast, the WWV$_{\text{inst.west}}$ of the neutral composite response is around 51% smaller than that of the transitioning composite response (cf. Figs. 6a and 6d).

The differences in WWV$_{\text{inst.west}}$ among the two composites is primarily explained by differences in the strength and sign of each hemisphere’s wind stress curl (Figs. 4a,c), and is summarized by the stronger curl($\gamma$) index found for the transitioning forcing compared to the neutral forcings (Fig. 5b). Note that both dominant positive wind stress curl in the Northern Hemisphere (NH) and dominant negative wind stress curl in the Southern Hemisphere (SH) excite upwelling RWs. The differences of the curl patterns seen in Fig. 4 are also clearly apparent when the curl patterns are divided by the Coriolis parameter $f$ (Fig. 5). This is noted as the RW signal is determined by Ekman pumping, which divides the wind stress curl by $f$. We also note that we focus on the wind stress curl/RW signal between $2^\circ$ and $8^\circ$ of latitude for the reasons mentioned above (see section 4b). The transitioning forcing category (Fig. 4a) gives the “ideal” case (as expected from theory) with westerly winds accompanied by strong negative (positive) wind stress curl around the WWV regions northern (southern) boundaries, which excites strong upwelling RWs in both hemispheres (blue shading, Fig. 6a). In comparison, the weaker wind stress curl and the smaller zonal wind extent of the neutral forcing category (Fig. 4c) excite comparatively weaker upwelling RWs at a similar distance to the equator (blue shading, Fig. 6d).

The transitioning composite simulation clearly shows an adjusted (Fig. 6b) and a late adjusted discharge (Fig. 6c) when the strong near-equatorial upwelling RWs reflect at the western boundary into the WWV region (blue areas in Figs. 6b,c) and the forced downwelling KWs leave the WWV region via the eastern boundary to travel poleward (red areas in Figs. 6b,c). In comparison, the RW signal of the neutral composite simulation reflecting into the WWV region is much weaker (blue area in Fig. 6e) and leads to little or no adjusted or late adjusted discharge (Figs. 6e,f).

b. $\tau_{\text{eq}}$ troughs

The spatial patterns of the pycnocline in the $\tau_{\text{eq}}$ trough simulations virtually mirror those of the $\tau_{\text{eq}}$ peak simulations for instantaneous, adjusted and late adjusted outcomes (cf. Figs. 6 and 7; spatial correlations between 15$^\circ$N and 15$^\circ$S vary between $-0.78$ and $-0.97$). This again suggests that the weaker adjusted and late adjusted responses of neutral $\tau_{\text{eq}}$ trough simulations compared to those of transitioning $\tau_{\text{eq}}$ trough simulations are underpinned by the weaker near-equatorial RW response.

c. Comprehensive global ocean model simulations

The WWV magnitude and pycnocline spatial patterns for both the instantaneous and adjusted responses are largely consistent when conducting the experiments of this section with the comprehensive global coupled ocean–sea ice model GFDL-MOM025 [Fig. S4; see Zeller et al. (2019) for a model description]. Thus, we conclude that the results presented here are not significantly impacted by the simplicity of the ocean model utilized.

d. Idealized wind stress experiments

We conduct a simple experiment to better understand if the stronger adjusted WWV response of the $\tau_{\text{eq}}$ peaks transitioning forcing compared to the $\tau_{\text{eq}}$ peaks neutral forcing is due to (i) the larger longitudinal wind stress extent of the zonal transitioning forcing (Fig. 4a) or (ii) the larger latitudinal extent of zonal wind stress forcing of the neutral forcing (Fig. 4c).

To this end, we produce a zonal wind forcing that has the average latitudinal extent of the $\tau_{\text{eq}}$ neutral peak simulation and the average longitudinal extent of the $\tau_{\text{eq}}$ transitioning peak simulation (Fig. 8c; red line, Figs. 8e–g). For example, the pattern in Fig. 8c is produced by multiplying at each latitude the zonally averaged (averaged over 120$^\circ$–280$^\circ$E) zonal wind stress of the neutral simulation (gray line in Fig. 4f) by the longitudinal profile (averaged over 5$^\circ$N–5$^\circ$S) of the transitioning simulation (green line in Fig. 4e) and normalizing this pattern to fit the green line in Fig. 4e. Similarly, we produce a forcing that has the average longitudinal extent of the neutral peak simulation (gray line Fig. 4e) and the average latitudinal extent of the transitioning peak simulation (green line, Fig. 4f;
We also carry out two additional simulations that utilize the average longitudinal and latitudinal extent of the transitioning and neutral simulations for comparison (idealized transitioning and neutral forcings; Figs. 8a,b). We first note that the idealized transitioning wind stress curl pattern and its adjusted response are consistent with the nonidealized equivalent (cf. Figs. 4a and 8a; green solid and dashed lines in Fig. 8h). The wind stress curl pattern of the idealized neutral forcing does not display the longitudinal variability of its nonidealized equivalent (cf. Figs. 4c and 8b); however, its adjusted response is also very small (gray solid and dashed lines in Fig. 8h).

The consistency of the idealized and nonidealized forcing WWV changes suggests that this experiment will provide a reasonable base to better understand the relative importance of the wind stress forcing’s latitudinal and longitudinal extent. The adjusted response of the longitudinal neutral and latitudinal transitioning forcings shows a clear discharge that is only 19% weaker than that of the idealized transitioning composite forcing (cf. blue and green dashed lines in Fig. 8h).

Similarly, the differences between the adjusted responses of the transitioning longitudinal and neutral latitudinal forcing and the idealized neutral forcing are negligible (red and gray dashed lines in Fig. 8h). These results indicate that the latitudinal extent of the wind stress forcing plays the dominant role in determining the strength of the adjusted response and the longitudinal extent plays a secondary role. Summarizing, this means that when there is a strong wind stress curl equatorward of around ±8° latitude and the wind stress has a large longitudinal extent, more RWs are excited leading to a slightly stronger adjusted response (cf. blue dashed and green dashed line in Fig. 8h). However, when there is a weak wind stress curl equatorward of around ±8° latitude, weak RWs are excited and extending this forcing longitudinally has little impact on the adjusted response (cf. red and gray dashed lines in Fig. 8h).

6. Average winds versus high-frequency wind events

The link between the longer-term average composite wind stress (Fig. 4) and those of high-frequency wind events can be
understood by identifying the individual wind events in regions around the tropical Pacific (black boxes, Figs. 9a,d). We distinguish between equatorial wind events (located in the four boxes between 5°8N and 5°8S) and NH/SH off-equatorial wind events [located in the four (three) boxes within 5°–15°N (5°–15°S); Figs. 9a and 9d]. A westerly (easterly) wind event [WWE (EWE)] is identified when the zonal mean wind stress averaged over a given region and smoothed by a 3-day triangular filter exceeds (falls below) 0.04 N m⁻² for a period of three or more days [consistent with McGregor et al. (2016)].

Our results indicate that rather than the wind blowing constantly over three months with the same average strength (as seen in Figs. 4a–d), the response is episodic as, on average, each identified region displays two (up to eight) wind events per 3-month forcing period. We also find that these wind events last on average 7.1 days and have an average strength ranging between 0.042 and 0.13 N m⁻². The decomposition of the wind stress composites into individual high-frequency events in the different equatorial and off-equatorial regions allows us to explain why neutral simulations are meridionally broader and have weaker wind stress curls than transitioning simulations.

As each individual wind event has a different duration and magnitude (see also Puy et al. 2016), we further calculate the intensity of each wind event by multiplying its average strength by its duration in days. We note that the results presented in this section are similar when counting the number of wind events instead of calculating the intensity (not shown). In the following we only consider WWEs (EWEs) during the ξeq peak (trough) forcing periods as the role of EWEs (WWEs) is negligible (not shown).

The average ratio of off-equatorial to equatorial WWE intensity during these 3-month forcing periods is lowest for forcing periods that are categorized as transitioning (1.25 for the peaks and 0.93 for the troughs; Figs. 9a and 9d). Consistent with the wind stress calculations reported in section 4b (Fig. 5a), this equatorial intensity difference is not statistically significant (applying a t test at the 95% confidence interval). Further to this, looking for hemispheric differences we find that the WWE (EWE) intensity of almost each SH off-equatorial box is weaker for the average neutral forcings than for the average transitioning forcings (Figs. 9a,d). In contrast, the WWE (EWE) intensity of almost all NH off-equatorial boxes is stronger for the average neutral forcings than for the average transitioning forcings (Figs. 9a,d) and stronger differences are found for the peak simulations compared to the trough simulations. These results are consistent

**Fig. 9.** (a) Average number of westerly wind event (WWE) intensity [the average magnitude (N m⁻²) multiplied by the duration (days)] during the 3-month forcing periods of the ξeq peaks’ transitioning (green numbers) and neutral (gray numbers) simulations (Fig. 2c) for 11 different regions in the equatorial Pacific as marked by the boxes. (d) As in (a), but for easterly wind event (EWE) intensity during the trough forcing periods. (b),(c) As in (b), but for WWE (EWE) intensity during the 3-month forcing periods of each single ξeq peak transitioning (green) and neutral (gray) simulation. (e) As in (b), but for EWE intensity during the trough forcing periods. (f) As in (b) and (c), but for the SH instead of the NH.
with the composite wind stresses (Fig. 4): The strong wind stress curls equatorward of around \( \pm 8^\circ \) latitude of the transitioning simulations (Figs. 4a,b) are produced by weak off-equatorial wind stresses, while the weaker wind stress curls equatorward of around \( \pm 8^\circ \) latitude of the neutral simulations (Figs. 4c and 4d) are produced by a meridionally broader wind stress pattern that is most apparent in the NH. The calculation of wind event intensity in different regions now allows us to further understand the significance of the NH/SH difference. We find that the ratio of NH off-equatorial WWV (EWE) intensity to equatorial WWV (EWE) intensity is significantly larger for the neutral simulations than for the transitioning simulations (applying a \( t \) test at the 95% confidence interval; Figs. 9b,e), there is no significant difference for the ratio of SH off-equatorial WWV (EWE) intensity to equatorial WWV (EWE) intensity (Figs. 9c,f). This means that the increase in off-equatorial to equatorial wind events intensity ratio seen during neutral responses compared to transitioning responses arises due to an increase in the intensity of NH off-equatorial wind events.

7. Summary and discussion

This paper uses shallow water model (SWM) simulations to categorize the adjusted WWV responses following the strongest equatorial Pacific westerly/easterly wind stress between 1980 and 2016. As in Neske and McGregor (2018), the adjusted WWV component is the WWV changes that are driven by Rossby waves (RWs) and their reflection into the WWV region at the western boundary (McGregor et al. 2016). The importance of understanding the adjusted response arises as it is the slowly evolving component of WWV changes that is responsible for the longer lead time relationship to ENSO SSTs (Neske and McGregor 2018). Two categories of adjusted responses were then defined and utilized. These are (i) transitioning adjusted responses (i.e., equatorial westerly/easterly wind stress is followed by an adjusted discharge/recharge) and (ii) neutral adjusted responses (i.e., equatorial wind stress is followed by little to no adjusted recharge or discharge). Traditional conceptual models (e.g., Suarez and Schopf 1988; Jin 1997; Wang 2001) lead us to expect that strong equatorial westerly (easterly) winds during warm (cold) phases of ENSO initiate strong RWs, which eventually lead to strong transitioning adjusted discharges (recharges), conditioning the system for a change in ENSO phase. Recent findings, however, show that the STD of the instantaneous WWV [predominantly consisting of the wind-forced Kelvin waves (KWs); correlation to equatorial wind stress = 0.86; Fig. 2a] increased by 15% from the pre- to post-2000 period, while the STD of the adjusted WWV decreased by 44% at the same time (Neske and McGregor 2018). Thus, our study seeks to better understand what is driving the recent WWV changes as they are clearly at odds with traditional theories that link equatorial wind stress to the adjusted response.

a. Categorized adjusted WWV responses and their pre- and post-2000 importance

In our study, we analyze the 3-month forcing periods around the strongest anomalous positive and negative zonal equatorial wind stress (anomalies to the mean seasonal cycle) (Fig. 2a). These simulations show that slightly less than half (47%) of these strong anomalous equatorial wind stresses are followed by the transitioning adjusted responses that would be expected from ENSO theory (green simulations, Figs. 2b–d). The remaining identified strong anomalous equatorial wind stresses are followed by a neutral adjusted response (53%; gray simulations, Figs. 2b–d). Moreover, during the pre-2000 period 55% (11/20) of the strongest equatorial winds are followed by transitioning adjusted responses, and this reduces to only 30% (3/10) for the post-2000 period. As such, the post-2000 period is dominated by the neutral adjusted responses that have little to no adjusted recharge and discharge of WWV (Fig. 2b). Consequently, this finding is consistent with the post-2000 decline in adjusted WWV STD reported by Neske and McGregor (2018).

b. Dynamical differences between transitioning and neutral adjusted responses

The dynamical differences between transitioning and neutral responses are linked to the differences in wind stress patterns. The average neutral wind stress is shown to have a latitudinally broader and longitudinally narrower extent than the average transitioning wind stress forcing (Fig. 4). We use a series of idealized wind forced SWM experiments to further show that the negligible adjusted WWV response of the neutral simulations is primarily caused by the different latitudinal structure, which includes zonal wind stresses that are latitudinally broader and have a northward displacement of the maximum zonal wind stresses. This latitudinal difference creates a weaker wind stress curl equatorward of around \( \pm 8^\circ \) latitude, which diminishes the Ekman pumping and leads to a weaker RW signal when compared to the adjusted transitioning simulations (Figs. 5–8). The latitudinal differences between the average forcing of the transitioning and neutral simulations can be explained by looking at wind event intensity, as the average neutral forcing period displays a higher ratio of off-equatorial to equatorial wind event intensity than seen in the transitioning forcing period (Fig. 9).

Furthermore, the increase in off-equatorial to equatorial wind event intensity ratio seen during neutral responses compared to transitioning responses arises due to differences in the intensity of NH off-equatorial wind events (Fig. 9).

We further show that the differences in (i) WWV magnitude and (ii) pycnocline spatial patterns among the transitioning and neutral simulations of the SWM are largely consistent when repeating these simulations with the comprehensive global coupled ocean sea ice model GFDL-MOM025 (Fig. S4). Consequently, we conclude that the results presented in this study are not significantly impacted by the simplicity of the ocean model utilized.

The underlying average SSTA patterns during the transitioning (neutral) forcing periods are similar to EP (CP)-like ENSO SSTAs (Fig. 3) with patterns similar as shown for instance by McPhaden et al. (2011). Consistent with our finding of strong discharge (neutral) adjusted responses forced during EP (CP)-like El Niño SSTAs, Singh and Delcroix (2013) find an overall poleward mass transport during EP El Niños but not during CP El Niños. While our study links the neutral adjusted
response to weak RWs, Singh and Delcroix (2013) link the absence of mass transports during CP El Niño to a compensating effect of a poleward (equatorward) transport in the western (eastern) Pacific.

While there appears to be a relationship between the changing spatial structure of wind stresses and SSTAs, the relationship between SST and wind is complex and often nonlinear (e.g., Frauen and Dommenget 2010; Dommenget et al. 2013; DiNezio and Deser 2014; Izumo et al. 2020). Thus, we do not attempt to provide a clear physical explanation for this linkage here. Further numerical studies are needed in order to identify the role of SST spatial structure (EP/CP ENSO events) on the resulting winds, as there is too much interdependency to do this statistically.

c. ENSO asymmetry

Recent research has demonstrated that the boreal WWVwest discharge is a better predictor for La Niña events than the boreal WWVwest recharge for El Niño events (Planton et al. 2018). Note that WWVwest was shown to be closely linked to the adjusted response (Fig. 1d); and both the WWVwest has been shown by Izumo et al. (2019) and the adjusted response has been shown by Neske and McGregor (2018) to include the long-term lead time to ENSO SST. Planton et al. (2018) partially link the asymmetry in El Niño/La Niña predictability to the asymmetry in WWVwest that shows stronger discharges 13 months before a La Niña event compared to recharges 13 months before an El Niño event. Similarly, Neske and McGregor (2018) show stronger magnitudes of the adjusted discharge compared to the adjusted recharge. Planton et al. (2018) link this difference in recharge and discharge magnitudes to stronger westerly equatorial wind stress during El Niño phases compared to easterly wind stress during La Niña phases, which can be linked to two things: (i) stronger (weaker) positive (negative) SSTAs during El Niño (La Niña) events, and (ii) the fact that nonlinearities with a given positive SST during El Niño lead to equatorial westerly wind anomalies that are stronger than the equatorial easterly wind anomalies that would result from a negative SST (e.g., Frauen and Dommenget 2010; Dommenget et al. 2013; DiNezio and Deser 2014). Further to this, we demonstrate a wind stress curl difference between the forcing periods during El Niños compared to those during La Niñas, consistent with the findings of Im et al. (2015), which adds to the adjusted recharge and discharge asymmetry: 56% (9 out of 16) of the strongest 3-month equatorial westerly forcing periods have a strong wind stress curl that leads to a strong adjusted discharge, whereas this is the case for only 36% (5 out of 14) of the strongest 3-month equatorial easterly forcing periods and their associated adjusted recharge (green peaks and troughs, Fig. 2b). Consistently, the delayed thermoline feedback that is from the traditional ENSO view thought to transition the ENSO phase (Suarez and Schopf 1988; Battisti and Hirst 1989; Jin 1997) is stronger (weaker) during El Niño (La Niña) events (Guan et al. 2019).

Our findings of the wind stress curl strength asymmetry together with the mentioned findings of previous studies serve as an explanation for the asymmetry in phase change with El Niños having a stronger tendency to be followed by La Niñas than vice versa (e.g., Kessler 2002; Larkin and Harrison 2002) by explaining the asymmetry of recharged and discharged WWV phases. We note that beside recharge and discharge asymmetry other mechanisms are at work adding to the asymmetry in phase change transition, such as (i) the nonlinearities between subsurface temperature gradient and thermocline depth (DiNezio and Deser 2014) or (ii) the stronger (weaker) westward (eastward) currents during El Niño (La Niña) conditions induced by wind stress nonlinearities (Clarke and Zhang 2019).

d. Final remarks

To sum up, our work emphasizes the importance of wind stress curl (related to the ratio of off-equatorial wind event to equatorial wind event intensity) for understanding recent changes in the drivers of WWV variability and the asymmetry of ENSO. The pre- to post-2000 differences in wind stress curl and the wind stress curl asymmetry presented in this paper are in line with (i) the post-2000 failure of the adjusted WWV contribution being a long-term ENSO SST precursor (>9 months) and (ii) the larger (smaller) adjusted discharge (recharge) magnitudes [as both shown by Neske and McGregor (2018)]. All these findings give an explanation for the asymmetry in duration and phase change transition, with El Niño events having a stronger tendency to be followed by La Niña events than vice versa (e.g., Kessler 2002; Larkin and Harrison 2002; Okumura and Deser 2010; Guan et al. 2019); as well, they indicate why La Niña events tend to be predictable at longer lead times than El Niño events (Luo et al. 2008; Dommenget et al. 2013; DiNezio et al. 2017).

Future work aims to (i) quantify the amount of strong adjusted responses that are preceded by weak equatorial wind stresses compared to strong equatorial wind stresses; (ii) identify the relative importance of wind stress curls in different regions around the equatorial and off-equatorial Pacific during the pre- and post-2000 periods; and (iii) elucidate the importance of RW cancellations excited in different regions around the equatorial Pacific for explaining pre- and post-2000 differences.

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