Role of Stochastic Atmospheric Forcing from the South and North Pacific in Tropical Pacific Decadal Variability

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ABSTRACT

Stochastic variability of internal atmospheric modes, known as teleconnection patterns, drives large-scale patterns of low-frequency SST variability in the extratropics. To investigate how the decadal component of this stochastically driven variability in the South and North Pacific affects the tropical Pacific and contributes to the observed basinwide pattern of decadal variability, a suite of climate model experiments was conducted. In these experiments, the models are forced with constant surface heat flux anomalies associated with the decadal component of the dominant atmospheric modes, particularly the Pacific–South American (PSA) and North Pacific Oscillation (NPO) patterns. Both the PSA and NPO modes induce basinwide SST anomalies in the tropical Pacific and beyond that resemble the observed interdecadal Pacific oscillation. The subtropical SST anomalies forced by the PSA and NPO modes propagate to the equatorial Pacific mainly through the wind–evaporation–SST feedback. This atmospheric bridge is stronger from the South Pacific than the North Pacific due to the northward displacement of the intertropical convergence zone and the associated northward advection of momentum anomalies. The equatorial ocean dynamics is also more strongly influenced by atmospheric circulation changes induced by the PSA mode than the NPO mode. In the PSA experiment, persistent and zonally coherent wind stress curl anomalies over the South Pacific affect the zonal mean depth of the equatorial thermocline and weaken the equatorial SST anomalies resulting from the atmospheric bridge. This oceanic adjustment serves as a delayed negative feedback and may be important for setting the time scales of tropical Pacific decadal variability.

1. Introduction

Historical observations indicate that Pacific climate varies substantially on decadal to interdecadal time scales (Nitta and Yamada 1989; Trenberth and Hurrell 1994; Mantua et al. 1997; Zhang et al. 1997; Minobe 1997; Garreaud and Battisti 1999; Deser et al. 2004; Chen and Wallace 2015). The oceanic and atmospheric anomalies associated with Pacific decadal variability (PDV) are organized into a large-scale pattern characterized by opposite sign of sea surface temperature (SST) anomalies between the tropical and central North Pacific and changes in the intensity of the large-scale atmospheric circulation over the North and South Pacific, similar to the anomaly pattern associated with

El Niño–Southern Oscillation (ENSO). PDV was first identified in the North Pacific and called the Pacific decadal oscillation (PDO; Mantua et al. 1997). In recognition of the extension to the South Pacific, the PDO was later referred to as the interdecadal Pacific oscillation (IPO; Power et al. 1999). A few decades of research that followed show that PDV arises from interactions of the ocean and atmosphere [see recent reviews by Liu (2012) and Newman et al. (2016)]. However, it remains unclear what gives rise to the coherent decadal changes across the Pacific basin, as well as the turnabout of positive and negative phases of decadal variability. The importance of understanding the mechanisms of PDV is signified by the global impacts (e.g., Mantua et al. 1997; Power et al. 1999; Seager et al. 2005; Okumura et al. 2012; Partin et al. 2013). PDV may also affect the amplitude and other characteristics of ENSO (Fedorov and Philander 2000; Wang and An 2002; Choi et al. 2009; Okumura et al. 2017) and interfere with anthropogenic global warming (Kosaka and Xie 2013; England et al. 2014; Dai et al. 2015), causing large uncertainty in

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regional climate projection (Hawkins and Sutton 2009; Meehl et al. 2013; Wallace et al. 2014).

In contrast to ENSO, in which interannual SST variations in the tropical Pacific drive extratropical ocean–atmosphere anomalies, PDV is likely to result from multiple processes that originate in both the tropics and extratropics (Newman et al. 2016). Because of the similarities of anomaly patterns between PDV and ENSO, some early studies searched the origin of decadal variability in the tropics and proposed that slow propagation of off-equatorial Rossby waves plays an important role in setting the decadal time scales, much like the delayed oscillator theory for ENSO (Knutson and Manabe 1998; Yukimoto et al. 2000; Jin 2001). This hypothesis is consistent with the observations that tropical Pacific SST and wind anomalies show wider meridional structure for decadal than interannual time scales. Other studies argued that random changes in ENSO activities could induce decadal changes in the tropical Pacific mean state without explicit oscillation mechanisms (Rodgers et al. 2004; Vimont 2005). However, the SST variability in the tropical Pacific becomes weaker than that in the North Pacific on decadal time scales, which is not explained by the tropical origin of PDV (Zhang et al. 1997; Deser et al. 2004; Chen and Wallace 2015).

The difference in the relative magnitude of tropical and extratropical anomalies between PDV and ENSO led to an alternative hypothesis that PDV originates in the extratropics. In the extratropical oceans, large-scale patterns of SST variability are primarily driven by large-scale patterns of atmospheric circulation anomalies, so-called teleconnection patterns, through changes in surface heat fluxes and ocean currents (e.g., Cayan 1992; Marshall et al. 2001; Deser et al. 2010). The spatial organization of extratropical atmospheric circulation anomalies is caused mainly by nonlinear dynamical processes of the atmosphere and does not require external forcing (e.g., Branstator 1995). These intrinsic modes of atmospheric variability are stochastic by nature (e.g., Feldstein 2000) but are integrated to generate low-frequency variability by the ocean mixed layer due to its large thermal inertia. This mechanism, known as the stochastic climate model (Hasselmann 1976; Frankignoul and Hasselmann 1977), is widely adopted as the leading paradigm of low-frequency climate variability in the extratropics. Other oceanic processes, such as the reemergence of seasonally insulated ocean temperature anomalies (e.g., Alexander and Deser 1995; Hanawa and Sugimoto 2004) and wind-driven adjustment of the ocean gyre circulation (e.g., Miller et al. 1998; Taguchi et al. 2007), also contribute to the generation of low-frequency SST variability in the extratropical Pacific.

While decadal changes in tropical SSTs can readily affect the extratropical ocean and atmosphere through atmospheric teleconnections, it is less clear how decadal changes in the extratropical ocean and atmosphere are communicated to the tropics. Gu and Philander (1997) proposed oceanic pathways in which SST anomalies subducted in the subtropical North and South Pacific are advected adiabatically along the isopycnal surfaces to the equatorial thermocline on decadal time scales. However, the validity of this mechanism was questioned by several studies that showed that the water initially subducted in the North Pacific does not reach the equatorial Pacific (Schneider et al. 1999a; Nonaka et al. 2000; Nonaka and Xie 2000), likely due to the presence of a potential vorticity barrier north of the equator (Lu and McCreary 1995; Johnson and McPhaden 1999; Wang and Liu 2000). Nonetheless, both observations and models support the presence of an oceanic pathway from the subtropical South Pacific to the equatorial Pacific (Chang et al. 2001; Luo and Yamagata 2001; Giese et al. 2002; Yeager and Large 2004; Nonaka and Sasaki 2007; Tatebe et al. 2013), which may play an important role in setting the decadal time scales of variability.

The atmospheric bridge from the extratropics to the tropics, on the other hand, may contribute to the generation of concurrent decadal variability in the tropics. The surface wind anomalies associated with internal atmospheric variability extend into the subtropics. The resultant changes in subtropical SSTs force the same surface wind anomalies on the equatorward side and thus coupled wind–SST anomalies propagate toward the equator (Barnett et al. 1999; Pierce et al. 2000). Subsequent thermodynamic air–sea interactions may further organize a large-scale anomaly pattern in the tropics (Liu and Xie 1994; Zhou and Carton 1998; Dommenget 2010). Atmospheric models coupled to a mixed layer ocean simulate PDV that resembles the observed IPO in the absence of ocean dynamics, suggesting the critical role of stochastic atmospheric forcing and atmospheric bridge in basinwide decadal variability (Dommenget and Latif 2008; Clement et al. 2011; Okumura 2013). Yeh and Kirtman (2004) show that the IPO simulated in their model significantly weakens when stochastic atmospheric variability is suppressed. Studies based on idealized climate model experiments also support the importance of atmospheric bridge in communicating decadal changes in extratropical SSTs to the tropics (Liu and Yang 2003; Yang and Liu 2005; Matei et al. 2008).

Intriguingly, the atmospheric bridge to the tropical Pacific is suggested to be stronger from the South Pacific than the North Pacific (Liu and Yang 2003; Yang and

Liu 2005; Matei et al. 2008; Okumura 2013). Okumura (2013) suggests that the tropical component of the IPO is driven by stochastic variability of the southeasterly trade winds associated with the Pacific–South American (PSA) pattern, the southern counterpart of the well-known Pacific–North American (PNA) pattern. The resultant tropical Pacific variability may further extend the decadal anomalies into the North Pacific through atmospheric teleconnections. Okumura et al. (2017) also discuss that another mode of internal atmospheric variability over the South Pacific (PSA2; Ghil and Mo 1991) is potentially important for PDV different from the IPO. The strong influence of South Pacific atmospheric variability on the tropical Pacific has been attributed to the hemispheric asymmetry in mean climate. In the eastern tropical Pacific, the intertropical convergence zone (ITCZ) is displaced north of the equator and the southeasterly trades blow from the South Pacific into the equatorial region. Zhang et al. (2014b) argue that the direction of background winds controls the equatorward propagation of subtropical wind and SST anomalies although the detailed mechanisms remain unclear. There are, however, other studies that emphasize the role of stochastic atmospheric forcing from the North Pacific in the tropical component of PDV and spatial patterns of ENSO (e.g., Dommenget and Latif 2008; Di Lorenzo et al. 2015; Yu and Kim 2011; Yu et al. 2012).

The present study builds upon Okumura (2013) and Okumura et al. (2017) and investigates the role of extratropical atmospheric variability in forcing tropical Pacific decadal variability (TPDV). How do the modes of internal atmospheric variability over the South and North Pacific affect tropical Pacific climate on decadal time scales? What are the atmospheric and oceanic processes important for this linkage? Is the tropical Pacific more strongly influenced by atmospheric variability from the South Pacific than the North Pacific and what are the underlying mechanisms for the asymmetric linkages? Does the slow oceanic adjustment to South and North Pacific atmospheric variability play a role in setting the decadal time scales? To address these questions, we conduct a suite of climate model experiments by prescribing surface heat flux anomalies associated with the decadal component of the dominant atmospheric modes over the South and North Pacific while the ocean and atmosphere are fully coupled.

The rest of the paper is organized as follows. Section 2 introduces the models used in this study and describes the experiments conducted with these models. Section 3 presents the time mean response of Pacific climate to the two leading modes of atmospheric variability based on the model experiments. The key atmospheric and oceanic processes responsible for the mean climate

response are analyzed with focus on tropical air–sea interactions. Section 4 further discusses the relative importance of South and North Pacific atmospheric variability and the role of oceanic adjustment in TPDV. A summary is given in section 5.

2. Models and experiments

a. Models

In this study, we use the Community Climate System Model, version 4 (CCSM4), developed at the National Center for Atmospheric Research. The CCSM4 comprises four components (atmosphere, ocean, land, and cryosphere) linked by means of a flux coupler (Gent et al. 2011). The atmospheric component, the Community Atmosphere Model, version 4 (CAM4), uses a finite-volume dynamical core and has a horizontal resolution of 0.9° latitude $\times$ 1.25° longitude with 26 levels in the vertical. The oceanic component, the Parallel Ocean Program, version 2, has longitudinal spacing of 1.13° and latitudinal spacing varying from 0.27° at the equator to 0.65° at 60°N with 60 levels in the vertical. The CCSM4 simulates salient features of the observed tropical Pacific mean climate and variability, including asymmetry in the pattern and duration of El Niño and La Niña (Deser et al. 2012; Bellenger et al. 2014). The CCSM4 also reproduces observed diversity in the spatial patterns of ENSO, although equatorial Pacific SST variability is overestimated by 30% and extends too far into the western Pacific (Capotondi 2013). The PDV simulated in CCSM4 shows a pattern and periodicity similar to those in observations (Deser et al. 2012). However, both observations and paleoclimate reconstructions suggest that the CCSM4 underestimates the amplitude of PDV in the tropical Pacific (Ault et al. 2013; Newman et al. 2016). CCSM4 is an older version of the Community Earth System Model (CESM; Hurrell et al. 2013; Kay et al. 2015). We choose to use CCSM4 instead of CESM to be consistent with Okumura (2013) and Okumura et al. (2017). As pointed out in Okumura et al. (2017), CESM1 has larger cold SST bias in the equatorial Pacific although the simulated ENSO has more realistic amplitude.

In addition to CCSM4, we use the standalone CAM4 and CAM4 coupled to a slab ocean model (SOM). The SOM has the same horizontal grid as CAM4 and computes mixed layer ocean temperatures from surface heat fluxes and “ Q flux” that represents the effect of climatological ocean heat transport. The Q flux varies monthly, but the mixed layer depth is fixed at the annual mean value to better balance the annual cycle of heat (Bailey et al. 2009). We use three existing control

FIG. 1. Regression maps of global SST ($^{\circ}\text{C}$; shading), SLP (black contours at intervals of 0.1 hPa; zero contours thickened), and precipitation [positive (negative) contours in green solid (brown dashed) at intervals of 0.2 mm day^{-1}] anomalies on the (left) first and (right) second PCs of TPDV in the (top) CCSM4 and (bottom) CAM4-SOM control simulations. The PCs are standardized and all variables are smoothed with a 10-yr low-pass filter prior to the regression analysis. The first and second modes of TPDV explain 47% and 24% (38% and 18%) of the total low-pass-filtered variance in CCSM4 (CAM4-SOM), respectively.

simulations to compare with our experiments: a 1300-yr preindustrial integration of CCSM4, a 500-yr integration of CAM4-SOM, and a 300-yr integration of CAM4. The annual mean mixed layer depth and monthly Q flux in the CAM4-SOM run and monthly SST climatology in the CAM4 run are derived from the CCSM4 control simulation.

b. TPDV in the CCSM4 and CAM4-SOM control simulations

Figure 1 shows the global anomaly patterns associated with the two leading modes of TPDV in the CCSM4 and CAM4-SOM control simulations. These modes are defined based on empirical orthogonal function (EOF) analysis of 10-yr low-pass-filtered annual SSTs in the tropical Pacific (20°S – 20°N , 120°E – 80°W). In CCSM4, the leading mode of TPDV, which explains nearly half of the total variance, exhibits basinwide SST anomalies in the tropical Pacific. The global anomaly pattern strongly resembles the observed IPO. The second leading mode of TPDV, on the other hand, shows a zonal dipole pattern of SST anomalies in the tropical Pacific, with associated ocean–atmosphere anomalies in the extratropical North and South Pacific. As shown in Okumura (2013) and Okumura et al. (2017), these

global anomaly patterns associated with the two leading TPDV modes are surprisingly well reproduced in the CAM4-SOM control simulation in the absence of ocean dynamics. Note that tropical Pacific SST variability is considerably reduced in CAM4-SOM for the second TPDV mode, suggesting the important role of equatorial ocean dynamics in this mode. In both the CCSM4 and CAM4-SOM, extratropical atmospheric circulation anomalies project onto known teleconnection patterns: the PSA and PNA patterns for the first mode and the PSA2 and North Pacific Oscillation (NPO; Walker and Bliss 1932) patterns for the second mode. All these teleconnection patterns arise from internal atmospheric dynamics and can be identified as the leading modes of South and North Pacific atmospheric variability in the CAM4 control simulation (Fig. 2).

c. Model experiments

To investigate how the low-frequency variability of the internal atmospheric modes affect tropical Pacific climate, we conduct three sets of CCSM4 experiments by prescribing surface heat flux anomalies associated with the PSA, PSA2, and NPO patterns (the PSA, PSA2, and NPO experiments). We do not conduct a PNA experiment because the atmospheric circulation

FIG. 2. Regression maps of net surface heat flux (W m^{-2} ; shading) and SLP (contours at intervals of 1 hPa; zero contours thickened) anomalies on the (top) first and (middle) second PC of monthly SLP variability over the South Pacific ($15^{\circ}\text{--}70^{\circ}\text{S}$, $170^{\circ}\text{E}\text{--}70^{\circ}\text{W}$) and (bottom) the second PC of monthly SLP variability over the North Pacific ($15^{\circ}\text{--}70^{\circ}\text{N}$, $130^{\circ}\text{E}\text{--}110^{\circ}\text{W}$) in the CAM4 control simulation. The amplitude of surface heat flux anomalies is adjusted for decadal variability (see text for detailed methods). These surface heat flux anomalies are used to force the PSA, PSA2, and NPO experiments.

anomalies associated with the PNA pattern do not extend well into the subtropics in the absence of tropical SST forcing in the CAM4 control simulation. The PNA pattern also does not correlate with SST variability in the tropical Pacific on decadal time scales in the CAM4-SOM control simulation, suggesting that the enhanced linkage to the first TPDV mode likely results from tropical forcing (Okumura 2013). The PSA and PSA2 patterns are defined respectively as the first and second EOF modes of monthly sea level pressure (SLP) variability over the South Pacific ($15^{\circ}\text{--}70^{\circ}\text{S}$, $170^{\circ}\text{E}\text{--}70^{\circ}\text{W}$) and the NPO pattern as the second EOF mode of

monthly SLP variability over the North Pacific ($15^{\circ}\text{--}70^{\circ}\text{N}$, $130^{\circ}\text{E}\text{--}110^{\circ}\text{W}$), all based on the standalone CAM4 control simulation (Fig. 2). The PSA and PSA2 patterns explain 32% and 21% of the total monthly variance, respectively. The NPO pattern explains 21% of the total monthly variance. In this study, the patterns shown in Fig. 2 are defined as the positive phase of PSA, PSA2, and NPO. The surface heat flux forcing is obtained by regressing net surface heat flux anomalies (positive downward) onto the standardized monthly principal components (PCs). In CAM4, the PSA, PSA2, and NPO patterns emerge solely from internal nonlinear dynamical processes in the atmosphere without oceanic forcing and the associated surface heat flux anomalies do not include any effects of interactions with the ocean. The amplitude of surface heat flux anomalies is adjusted for decadal variability by multiplying them by the ratio of the standard deviations of the 10-yr low-pass-filtered PCs to the unfiltered PCs: 0.16 for PSA experiment and 0.13 for PSA2 and NPO experiments. Then these adjusted heat flux anomalies are increased by a factor of 4, 6, and 2 for the PSA, PSA2, and NPO experiments, respectively. We amplify the heat flux forcing to account for enhanced decadal variability of these atmospheric modes in CCSM4 compared to CAM4 (1.0, 1.5, and 1.4 times for PSA, PSA2, and NPO) and to ensure a robust response in the presence of internal decadal variability. We use different factors to amplify the PSA, PSA2, and NPO forcing so that the maximum heat flux anomalies in the subtropical Pacific are of similar magnitude ($\sim 6 \text{ W m}^{-2}$) for all the experiments. To mimic the effect of low-frequency variability of these stochastic atmospheric modes, we add these time-invariant heat flux anomalies to net surface heat flux calculated in the flux coupler when the atmosphere is coupled to the ocean once a day throughout the simulation. We conduct experiments only for the positive PSA, PSA2, and NPO forcing. Each experiment consists of three-member 100-yr simulations initialized with oceanic and atmospheric conditions obtained from different years of the CCSM4 control simulation. Preparatory experiments suggest that the model response is roughly linear to the amplitude and phase of the forcing. These experiments are analogous to the experiments conducted by Alexander et al. (2010), which explore the effect of wintertime NPO on subsequent ENSO events by imposing strong NPO heat flux forcing ($\sim 30 \text{ W m}^{-2}$ in the subtropical North Pacific) only during boreal winter.

The time mean ocean–atmosphere response in each experiment is calculated relative to the climatology of the last 500 years of the CCSM4 control simulation and averaged over the entire 100-yr simulations unless noted otherwise. The statistical significance of the mean SST

TABLE 1. List of the experiments conducted in this study.

Experiment name	Forcing	Model	Length (yr)	Ensemble size
PSA	PSA	CCSM4	100	3
NPO	NPO	CCSM4	100	3
PSA-slab	PSA	CAM4-SOM	100	1
PSA-15	PSA	CCSM4	15	17

response in each ensemble member is assessed by the percentile rank in the distribution of mean SSTs in 1201 overlapping 100-yr segments of the 1300-yr control simulation (Fig. S1 in the online supplemental material). The ensemble mean SST response is assessed by the percentile rank in the distribution of mean SSTs in 2000 composites of three randomly selected 100-yr segments. All the main results of ensemble mean SST response discussed in the paper are significant at the 95th percentile level (Figs. S1 and S2). The statistical significant tests based on a two-sided Student's t test yield similar results. Although Okumura et al. (2017) suggest that the PSA2 mode may force the second TPDV mode, the PSA2 experiment does not produce consistent response in the tropical Pacific among ensemble members (Fig. S3). Thus, we focus on the analysis of the PSA and NPO experiments in the following sections.

To assess the role of ocean dynamics in shaping the tropical Pacific response to the PSA forcing, we conduct a single 100-yr CAM4-SOM simulation forced with the same surface heat flux anomalies associated with the PSA pattern (PSA-slab experiment). The ocean–atmosphere response in this experiment is calculated relative to the climatology of the entire CAM4-SOM control simulation. Furthermore, to investigate the initial oceanic adjustment in the PSA experiment, we conduct 17-member 15-yr CCSM4 simulations forced with the same PSA surface heat flux anomalies (PSA-15 experiment). Each member of the PSA-15 experiment is initialized with oceanic and atmospheric conditions obtained from randomly chosen years of the CCSM4 control simulation. Combined with the three-member PSA experiment, we have a total of 20 ensemble members for the first 15 years under the PSA forcing, which is needed to distinguish forced transient response from large interannual variability in the tropical Pacific. Table 1 summarizes the information on the experiments conducted in this study.

3. Results

a. Annual mean response

Figure 3 shows annual mean anomalies of SST, SLP, and precipitation simulated in the PSA and NPO

experiments relative to the CCSM4 control simulation. The SST anomalies closely follow the surface heat flux forcing imposed over the South and North Pacific in the PSA and NPO experiments, respectively (Fig. 2). In both experiments, ocean–atmosphere anomalies are not confined to the forcing region but extend into the other hemisphere and outside the Pacific Ocean. In the PSA experiment, the ensemble mean response shows SST warming in the tropical Indo-Pacific. Precipitation increases over the warmer SSTs in the southeastern tropical Pacific and decreases to the west, indicative of an equatorward shift of the South Pacific convergence zone (SPCZ). These changes in tropical precipitation are associated with negative SLP anomalies and SST cooling in the central North Pacific. The ensemble mean anomaly pattern of the PSA experiment resembles the first TPDV mode in the CCSM4 control simulation (Fig. 1), although SST and precipitation anomalies are considerably smaller in the equatorial Pacific and SLP anomalies do not project onto the PSA pattern over the South Pacific. The Southern Hemispheric atmospheric circulation shows positive SLP anomalies in the midlatitude South Pacific and negative SLP anomalies over the Southern Ocean.

In the NPO experiment, the ensemble mean response shows SST cooling across the tropical Pacific, which is distinct from the zonal dipole pattern of the second TPDV mode in the CCSM4 control simulation (Fig. 1). Tropical precipitation anomalies are much smaller compared to the PSA experiment. Despite small changes in tropical precipitation, the Southern Hemispheric atmospheric circulation shows a meridional dipole pattern of SLP anomalies that projects onto the PSA pattern over the Pacific sector and extend farther into the Indian Ocean. The inferred increase in westerly winds over the Southern Ocean is associated with SST warming, for which changes in the ocean circulation may play an important role. The atmospheric circulation in the North Pacific has some projection onto the NPO pattern, although the centers of SLP anomalies are shifted to the east and positive SLP anomalies extend to the subtropical North Pacific where strong SST cooling occurs.

In both the PSA and NPO experiments, individual ensemble members show patterns of ocean–atmosphere anomalies similar to the ensemble mean, and thus the rest of the paper will focus on the ensemble mean response of both experiments. It should, however, be noted that there are significant variations among ensemble members due to internal variability independent of the surface heat flux forcing. Compared to the NPO experiment, the PSA experiment shows more variations among ensemble members, particularly over the tropical Pacific, and the North Pacific SST cooling is absent in one member.

FIG. 3. Annual mean response of SST ($^{\circ}\text{C}$; shading), SLP (black contours at intervals of 0.1 hPa; zero contours thickened), and precipitation [positive (negative) contours in green solid (brown dashed) at intervals of 0.2 mm day^{-1}] in the (top) ensemble mean and (bottom three rows) individual members of the (left) PSA and (right) NPO experiments.

b. Seasonality of the response: Atmospheric mechanisms

Ocean–atmosphere anomalies simulated in the PSA and NPO experiments show distinct seasonality outside the forcing region (Fig. 4). In the PSA experiment, the SST warming over the southeastern tropical Pacific expands into the equatorial Pacific from SON to DJF, but is confined to the south of the equator from MAM to JJA. The eastern equatorial Pacific SST anomalies change the sign and show cooling in JJA. The strong

seasonality of the equatorial Pacific SST response results in weak annual mean response in this experiment (Fig. 3). The enhanced equatorial SST warming in DJF is associated with a deepening of the Aleutian low and SST cooling in the central North Pacific, with the latter persisting into the following seasons. Compared to the PSA experiment, the seasonality of tropical Pacific response is weaker in the NPO experiment. The equatorial Pacific shows SST cooling year-round with the strongest cooling in MAM. Associated with the equatorial Pacific SST cooling, SLP anomalies show a meridional dipole

FIG. 4. Seasonal mean response of SST ($^{\circ}\text{C}$; shading) and SLP (contours at intervals of 0.1 hPa; zero contours thickened) in the ensemble mean of the (left) PSA and (right) NPO experiments.

pattern over the South Pacific and the Southern Ocean from JJA to DJF. It should also be noted that atmospheric circulation anomalies project onto the NPO pattern over the North Pacific in MAM, when the equatorial Pacific SST cooling becomes enhanced. The inferred decrease in westerly winds in the central North Pacific strengthens the underlying SST warming induced by the prescribed surface heat flux forcing.

To understand the seasonality of tropical Pacific response in the PSA and NPO experiments, we examine the seasonal evolution of SST, surface wind, and precipitation anomalies over the tropical Pacific (Fig. 5). In the PSA experiment, the SST warming induced by the

prescribed surface heat flux forcing increases precipitation over the southeastern tropical Pacific in all seasons, driving cross-equatorial northerly winds. The Coriolis force acting upon these northerly winds induces a westerly and easterly component south and north of the equator, respectively. To the south of the equator, the anomalous northwesterly winds weaken the climatological southeasterly trade winds and sea surface evaporation, which would act to enhance the SST warming in accordance with the wind–evaporation–SST (WES) feedback (Xie and Philander 1994). In SON, when the ITCZ is located at its northernmost position, the anomalous northwesterly winds extend toward the

FIG. 5. Seasonal mean response of SST ($^{\circ}\text{C}$; shading), surface wind (m s^{-1} ; vectors), and precipitation [positive (negative) contours in green solid (brown dashed) at intervals of 0.2 mm day^{-1}] in (left) the PSA-slab experiment and the ensemble mean of the (middle) PSA and (right) NPO experiments.

equatorial Pacific, concurrently with the equatorward expansion of the southeastern tropical Pacific warming. Both the equatorial surface wind and SST anomalies strengthen in DJF, suggestive of the WES feedback. In MAM, when the ITCZ migrates to its southernmost position, the surface wind and SST anomalies retreat to the southeastern tropical Pacific. In support of the role of WES feedback for the seasonality of the equatorial Pacific response, SST and surface wind response shows similar seasonal variations in the PSA-slab experiment, in which the ocean and atmosphere are coupled only through local surface heat fluxes. This experiment, however, does not simulate the eastern equatorial Pacific cooling in JJA, indicative of a role of equatorial ocean dynamics in the PSA experiment. The equatorial Pacific warming simulated in the PSA-slab experiment is stronger by a factor of 2 in SON and DJF compared to the PSA experiment.

In the NPO experiment, the SST cooling resulting from the surface heat flux forcing induces northeasterly winds over the northeastern tropical Pacific year-round, which would act to enhance the SST cooling. The anomalous northeasterly winds strengthen abruptly in JJA and SON, when the proximity of the ITCZ to this

region results in large negative precipitation anomalies. As the ITCZ migrates southward in DJF and MAM, the anomalous northeasterly winds extend south of the equator in the western-central basin. At the same time, the equatorial Pacific cooling intensifies across the basin, suggesting the role of both the WES feedback and equatorial ocean dynamics.

The equatorward expansion of subtropical Pacific SST and surface wind anomalies is closely tied to the seasonal migration of the ITCZ in both the PSA and NPO experiments, as well as in the PSA-slab experiment. Figure 6a shows latitude–time sections of the SST and surface wind anomalies averaged in the central-eastern tropical Pacific (180° – 90°W), along with the climatological latitudes of the ITCZ in the CCSM4 control simulation. In all the experiments, SST and surface wind anomalies in the subtropical forcing region extend toward the equatorial Pacific when the ITCZ migrates to the opposite hemisphere, and they retreat when the ITCZ moves back to the same hemisphere. While the ITCZ remains in the opposite hemisphere, both the equatorial SST and surface wind anomalies continue to intensify, in agreement with the WES feedback. The seasonal changes in equatorial SSTs lag those of surface

FIG. 6. (a) Latitude–time sections of monthly mean response of SST (°C; shading) and surface wind (m s^{-1} ; vectors) averaged in the central-eastern tropical Pacific (180° – 90° W) in (top) the PSA-slab experiment and the ensemble mean of the (middle) PSA and (bottom) NPO experiments. The green lines represent the position of the ITCZ (defined as the latitude where surface meridional wind vanishes) in the CCSM4 control simulation. (b) Latitude–time sections of monthly climatologies of precipitation (mm day^{-1} ; shading) and surface wind (m s^{-1} ; vectors) in (top) the CCSM4 control simulation and (bottom) observations. The white lines represent the position of the ITCZ in each dataset. The observations are based on the Global Precipitation Climatology Project, version 2 (Adler et al. 2003), and ERA-Interim (Dee et al. 2011) datasets for 1979–2015.

winds by 1–2 months, indicative of an atmospheric control of the seasonality of the equatorial response. In the PSA experiment, equatorial SSTs continue to cool into boreal summer and anomalous easterly winds develop even after the ITCZ moves north of the equator. In the NPO experiment, the equatorial cooling that intensifies during the brief excursion of the ITCZ to the Southern Hemisphere persists into summer.

It should be noted that the southward displacement of the Pacific ITCZ is a common bias in climate models (e.g., Lin 2007), including CCSM4. Figure 6b compares the seasonal cycle of climatological precipitation and surface winds averaged in the central-eastern tropical Pacific in the CCSM4 control simulation with observations. In CCSM4, the ITCZ migrates to the south of the equator from January to May along with the maximum precipitation. Unlike in CCSM4, the observed Pacific ITCZ is located north of the equator except the slight excursion in FMA, while the maximum precipitation band remains north of the equator year-round. Therefore, it is suggested that the equatorial Pacific is more strongly influenced by South Pacific atmospheric variability (Okumura 2013; Zhang et al. 2014a,b) and that the tropical linkage is likely underestimated in the PSA experiment and overestimated in the NPO experiment due to the model bias.

Similar impact of the seasonal migration of the ITCZ on the pattern of surface wind response and resultant WES feedback has been discussed in previous studies (Okajima et al. 2003; Wu et al. 2005). However, these studies did not explain why zonal wind anomalies in one hemisphere expand into the equatorial region when the ITCZ moves to the other hemisphere. This feature is not reproduced in a linear atmospheric model with a resting background state, in which zonal wind response to anomalous cross-equatorial SST gradient vanishes at the equator in the absence of the Coriolis force (Okajima et al. 2003). To understand the cause of equatorial zonal wind response in our experiments, we conduct momentum budget analysis. Following Okumura and Xie (2004), the surface zonal momentum equation near the equator may be written approximately as

$$\varepsilon u = -\frac{1}{\rho} \frac{\partial P}{\partial x} - v \frac{\partial u}{\partial y} + f v, \quad (1)$$

where the vertical eddy viscosity is parameterized as a simple damping term with a time scale of ε^{-1} . All the other variables have their conventional meanings. We have neglected the zonal advection since surface wind variations are much greater in the meridional than in the zonal direction. Linearizing Eq. (1) around the monthly climatology in the control simulation leads to the

FIG. 7. Longitude–time section of monthly mean response of (top row) surface zonal wind (m s^{-1} ; shading) and SST (contours at intervals of 0.05°C ; zero contours thickened) and (bottom three rows) zonal momentum acceleration (10^{-5} m s^{-2}) by (third row) pressure gradient force, (fourth row) meridional advection, and (second row) their sum averaged in the equatorial Pacific (5°N – 5°S) in (left) the PSA-slab experiment and the ensemble mean of the (middle) PSA and (right) NPO experiments.

following equation for the monthly mean anomaly of zonal wind velocity in the experiments:

$$\varepsilon u' = -\frac{1}{\rho} \frac{\partial P'}{\partial x} - \left(v' \frac{\partial \bar{u}}{\partial y} + \bar{v} \frac{\partial u'}{\partial y} \right) + f v', \quad (2)$$

where the overbar and prime denote the monthly climatology in the control simulation and the monthly mean anomaly in the experiments, respectively.

Figure 7 shows the longitude–time sections of the terms on the right-hand side of Eq. (2), as well as the zonal wind and SST anomalies along the equatorial

Pacific in the three experiments. Note that the meridional advection terms $v'(\partial\bar{u}/\partial y)$ and $\bar{v}(\partial u'/\partial y)$ are combined and the overbar and prime symbols are omitted. The sum of these terms agrees well with the seasonal cycle of zonal wind anomalies, particularly in the PSA and PSA-slab experiments, indicating that Eq. (2) is indeed a good approximation of the surface zonal momentum balance. In the PSA-slab experiment, the pressure gradient dominates the momentum budget west of 140°W , but the momentum advection plays a more important role to the east. During the equatorial cold season, the cross-equatorial winds transport westerly momentum into the eastern equatorial Pacific, weakening the climatological easterly winds. The resultant SST warming induces westward pressure gradient and hence westerly winds to the west, which would expand the SST warming farther to the west. Both the zonal wind and SST anomalies show distinct westward propagation during the equatorial cold season, analogous to the annual cycle of the equatorial Pacific and Atlantic (e.g., Mitchell and Wallace 1992; Xie 1994). During January through May, the cross-equatorial northerly winds transport easterly momentum into the equatorial Pacific, concurrently with the southward migration of the ITCZ (Fig. 6b). The easterly momentum reverses the westerly wind anomalies and weakens the SST warming. This advection of easterly momentum due to the unrealistic southward migration of ITCZ may lead to an underestimation of the impact of PSA forcing on the equatorial Pacific in CCSM4. The PSA experiment shows similar seasonal progression of the zonal momentum budget. However, despite the surge of anomalous westerly momentum during the equatorial cold season, the eastern equatorial Pacific exhibits cooling from June through September, which acts to decelerate anomalous westerly winds.

In the NPO experiment, the momentum advection contributes to the acceleration of easterly winds over the western equatorial Pacific during late boreal winter–spring, but its effect is much smaller compared to the PSA experiment. Equation (2) does not explain the persistent easterly wind anomalies near the date line in this experiment. The cause of the discrepancy is not clear, but it is possible that the momentum budget analysis based on monthly data underestimates the effect of momentum advection, which is already included in the monthly data.

c. Seasonality of the response: Oceanic mechanisms

How do these seasonally varying surface wind anomalies interact with the dynamics of the equatorial Pacific in the PSA and NPO experiments? Figure 8 shows longitude–depth sections of annual and seasonal mean ocean potential temperature anomalies in the equatorial

Pacific in both experiments, along with the thermocline depth in the control run approximated by the depth of the isopycnal $\sigma_\theta = 25.5 \text{ kg m}^{-3}$ (Fig. S4). In the PSA experiment, the annual mean warming of the equatorial Pacific seen in the SST field (Fig. 3) is confined to the surface layer. Underneath this surface warming is cooling along the climatological thermocline, which indicates basinwide shoaling of the thermocline. The subsurface cooling is stronger in the west than in the east, indicating a decrease in the east–west tilt of the thermocline consistent with the predominant westerly wind anomalies (Fig. 7). In SON and DJF, when the anomalous westerly winds intensify, the surface warming becomes stronger and the zonal thermocline tilt decreases further. In MAM and JJA, following a brief appearance of anomalous easterly winds, the subsurface cooling in the western Pacific extends eastward and into the surface layer in the eastern Pacific, where the shallow thermocline allows subsurface influences on the mixed layer.

In the NPO experiment, the annual mean cooling of the equatorial Pacific seen in the SST field (Fig. 3) is confined to the surface layer in the western Pacific but extends below the thermocline in the eastern Pacific, indicative of the Bjerknes feedback. Subsurface temperature anomalies show opposite sign between the western and eastern Pacific, indicating an increase in the east–west tilt of the thermocline consistent with the anomalous easterly winds (Fig. 7). In contrast to the PSA experiment, there is little annual mean change in the zonal mean depth of the thermocline in this experiment. The changes in the zonal thermocline tilt and surface temperatures are most pronounced in DJF and MAM, when the anomalous easterly winds intensify. In JJA and SON, when the anomalous easterly winds weaken, the subsurface warming in the western Pacific extends eastward and the surface cooling diminishes across the basin.

Although changes in the zonal tilt of the equatorial Pacific thermocline are dynamically consistent with local zonal wind anomalies in both the PSA and NPO experiments, it is not clear why changes in the zonal mean depth of the thermocline are evident only in the PSA experiment. To examine the role of remote wind forcing, Fig. 9 shows annual and seasonal mean anomalies of wind stress curl and depth of the isopycnal $\sigma_\theta = 25.5 \text{ kg m}^{-3}$ in the tropical Pacific. In the PSA experiment, negative wind stress curl anomalies located between 10° and 25°S persist year-round, which would drive southward Sverdrup transport and cause zonal mean shoaling of the equatorial thermocline. These persistent wind stress curl anomalies are associated with the cyclonic circulation induced

FIG. 8. Longitude–depth sections of (top) annual and (bottom four rows) seasonal mean ocean temperature response ($^{\circ}\text{C}$; shading) averaged in the equatorial Pacific (3°N – 3°S) in the ensemble mean of the (left) PSA and (right) NPO experiments. The white lines represent the climatological isopycnal $\sigma_{\theta} = 25.5 \text{ kg m}^{-3}$ in the CCSM4 control simulation. The vertical axis indicates depth in meters.

by the southeastern tropical Pacific warming (Fig. 4). The negative wind stress curl anomalies peak in SON and the resultant shoaling of the thermocline appears to propagate westward as Rossby waves and then eastward along the equator as Kelvin waves during DJF and MAM. In the NPO experiment, wind

stress curl anomalies are much smaller compared to those in the PSA experiment. Significant anomalies are found only during JJA and SON, when the precipitation and surface wind response intensify over the subtropical North Pacific (Fig. 5). These wind stress curl anomalies, however, have little impact on the

FIG. 9. (top) Annual and (bottom four rows) seasonal mean response of surface wind stress curl (10^{-8} N m^{-3} ; shading) and depth of the isopycnal $\sigma_\theta = 25.5 \text{ kg m}^{-3}$ (black contours at intervals of 1.5 m; zero contours thickened) in the ensemble mean of the (left) PSA and (right) NPO experiments.

equatorial thermocline as positive and negative anomalies of comparable magnitude drive opposite Sverdrup transport.

d. Ocean mixed layer heat budget

To quantify the relative contributions of thermodynamic and dynamic processes to the seasonal SST response in the PSA and NPO experiments, we conduct heat budget analysis of the ocean mixed layer (Fig. 10). By integrating the temperature equation over the seasonal mixed layer depth H , the heat budget of the ocean mixed layer can be expressed as

$$\begin{aligned} \left\langle \frac{\partial T}{\partial t} \right\rangle &= \left\langle -u \frac{\partial T}{\partial x} \right\rangle + \left\langle -v \frac{\partial T}{\partial y} \right\rangle + \left\langle -w \frac{\partial T}{\partial z} \right\rangle \\ &\quad + Q_{\text{net}} - \delta Q_{\text{SW}} + \epsilon \\ Q_{\text{net}} &= Q_{\text{LH}} + Q_{\text{SH}} + Q_{\text{SW}} + Q_{\text{LW}} \\ \langle a \rangle &= \rho C_p \int_{-H}^0 a dz, \end{aligned} \quad (3)$$

where T is the potential temperature; u , v , and w are the zonal, meridional, and vertical currents, respectively; Q_{net} is the net surface heat flux consisting of latent heat (Q_{LH}), sensible heat (Q_{SH}), shortwave radiation (Q_{SW}),

and longwave radiation (Q_{LW}) fluxes; δQ_{SW} is the fraction of Q_{SW} that penetrates the base of the mixed layer; ε is the residual term including the effect of turbulent mixing, horizontal diffusion, and other unresolved processes; and $\rho C_p = 4.1 \times 10^6 \text{ J m}^{-3} \text{ K}^{-1}$ is the product of the density and specific heat of seawater. The mixed layer depth H is calculated as the depth at which potential density changes by 0.125 kg m^{-3} from its surface value (Danabasoglu et al. 2012). The first four terms of Eq. (3) represent, from left to right, the heat storage and the zonal, meridional, and vertical ocean heat transport. Linearizing Eq. (3) around the seasonal climatology in the control simulation leads to the following equation for the seasonal mean anomaly of the heat budget in the PSA and NPO experiments:

$$\begin{aligned} \left\langle \frac{\partial T'}{\partial t} \right\rangle = & \left\langle -\bar{u} \frac{\partial T'}{\partial x} \right\rangle + \left\langle -u' \frac{\partial \bar{T}}{\partial x} \right\rangle + \left\langle -\bar{v} \frac{\partial T'}{\partial y} \right\rangle \\ & + \left\langle -v' \frac{\partial \bar{T}}{\partial y} \right\rangle + \left\langle -\bar{w} \frac{\partial T'}{\partial z} \right\rangle \\ & + \left\langle -w' \frac{\partial \bar{T}}{\partial z} \right\rangle + Q'_{\text{net}} + \epsilon', \end{aligned} \quad (4)$$

where the overbar and prime denote the seasonal climatology in the control run and the seasonal mean anomaly in the PSA and NPO experiments, respectively. The penetrating shortwave radiation term is neglected as it is an order of magnitude smaller than Q'_{SW} . In the equilibrium response, the heat storage becomes negligible and the terms on the right-hand side of Eq. (4) nearly balance. Following Xie et al. (2010), Q'_{LH} is further decomposed into Newtonian damping

$$Q'_{\text{LHT}} = \frac{L}{R_v \bar{T}^2} \overline{Q'_{\text{LH}} T'}, \quad (5)$$

and a residual that represents atmospheric forcing

$$Q'_{\text{LHa}} = Q'_{\text{LH}} - Q'_{\text{LHT}}, \quad (6)$$

where L is the latent heat of evaporation and R_v is the gas constant for water vapor. Note that Q'_{LHa} includes the effect of atmospheric adjustments in wind speed, relative humidity, and stability effects. We use this term to evaluate the role of WES feedback.

Figure 10 shows seasonal mean anomalies of the atmospheric heat fluxes and ocean heat transport components in the tropical Pacific calculated for the PSA and NPO experiments, along with low and high cloud fraction anomalies. In both experiments, Q'_{LHT} represents damping of seasonal SST anomalies, counteracting the other forcing terms. In the PSA experiment, Q'_{LHa} is

the primary mechanism to enhance and expand the southeastern tropical Pacific warming induced by the imposed surface heat flux anomalies, confirming the role of WES feedback. The warming effect of Q'_{LHa} is confined to the south of the equator in MAM and JJA but extends into the equatorial Pacific in SON and DJF, following the seasonal migration of westerly wind anomalies (Fig. 5). The resultant SST warming reduces low-level clouds over the central-eastern equatorial Pacific, which positively feeds back to the SST warming by increasing the incoming shortwave radiation. Over the southeastern tropical Pacific, on the other hand, increased deep convective clouds damp the SST warming by reducing the incoming shortwave radiation. In addition to the surface heat fluxes, the ocean heat transport also plays an important role in the seasonal heat budget of the ocean mixed layer. In particular, the vertical ocean heat transport contributes to equatorial cooling from DJF to JJA, when the subsurface cooling intensified in the western Pacific propagates to the eastern Pacific (Fig. 8). A linear decomposition of the vertical ocean heat transport shows that the cooling mostly results from climatological upwelling of subsurface temperature anomalies [the fifth term on the right-hand side of Eq. (4); not shown]. The zonal and meridional ocean heat transport also contributes to the cooling of the central and eastern equatorial Pacific, respectively. These oceanic feedbacks counteract the warming due to the WES feedback and lead to the reversal of SST anomalies in the eastern equatorial Pacific in JJA. The absence of these oceanic feedbacks explains the stronger equatorial Pacific warming in the PSA-slab experiment (Fig. 5).

In the NPO experiment, Q'_{LHa} acts to enhance the subtropical North Pacific cooling induced by the imposed surface heat flux anomalies during JJA and SON, in support of the WES feedback. Compared to the PSA experiment, however, the WES feedback is less effective in propagating the subtropical SST anomalies toward the equatorial Pacific. The cooling effect of Q'_{LHa} is confined to the north of the equator except in DJF, consistent with the seasonal migration of the easterly wind anomalies (Fig. 5). The weak but persistent cooling of the tropical Pacific in this experiment is rather maintained by positive feedback between low-level clouds and shortwave radiation. The changes in low-level clouds and shortwave radiation are pronounced over the northeastern tropical Pacific and the eastern equatorial Pacific, where the climatological stratus cloud deck forms over cold SSTs. The ocean heat transport, on the other hand, appears to play an important role in initiating the equatorial Pacific cooling in the eastern basin and spreading this

FIG. 10. Seasonal mean response of the ocean mixed layer heat budget (W m^{-2} ; shading) in the ensemble mean of the PSA and NPO experiments. See text for the definition of each term. The panels for Q'_{SW} and Q'_{LW} also show the seasonal mean response of low and high cloud fraction (contours at intervals of 0.005 and 0.01; zero contours thickened), respectively.

cooling to the whole tropical Pacific. In DJF, the enhanced anomalous easterly winds increase the east-west tilt of the thermocline (Figs. 7 and 8). The vertical ocean heat transport cools the eastern equatorial Pacific through both climatological and anomalous upwelling [the fifth and sixth terms on the right-hand side of Eq. (4)], while the zonal ocean heat transport cools

the western-central equatorial Pacific mainly through anomalous westward current [the second term on the right-hand side of Eq. (4)]. The resultant equatorial cooling is spread off the equator by the meridional ocean heat transport and the tropical-wide cooling is maintained by the cloud-radiation feedback throughout the year.

FIG. 10. (Continued)

4. Discussion

a. Relative importance of atmospheric forcing from the South and North Pacific

The heat budget analysis of the ocean mixed layer suggests that the WES feedback is more effective in propagating subtropical Pacific SST and surface wind anomalies into the equatorial Pacific in the PSA experiment than in the NPO experiment (Fig. 10). In both the

PSA and NPO experiments, the equatorward expansion of subtropical SST and wind anomalies mainly occurs during the seasons when the ITCZ is located in the opposite hemisphere. This is because the background cross-equatorial winds advect anomalous zonal momentum from upstream (Fig. 7), while subsequent thermodynamic air-sea interactions enhance the equatorial anomalies (Fig. 10). Since the ITCZ is located north of the equator most of the year, the equatorial

Pacific is more influenced by atmospheric forcing from the South Pacific than the North Pacific. The strong linkage of TPDV to the South Pacific is also supported by some observational studies that show pronounced tropical ocean–atmosphere anomalies south of the equator (Garreaud and Battisti 1999; Deser et al. 2004). Our analysis provides a mechanism for the apparent linkage between the equatorial asymmetry in tropical Pacific mean climate and the asymmetric impact of South and North Pacific atmospheric variability on the tropical Pacific, which has been suggested by previous modeling studies (Liu and Yang 2003; Yang and Liu 2005; Matei et al. 2008; Okumura 2013; Zhang et al. 2014b).

The impact of South and North Pacific atmospheric variability on TPDV analyzed in this study is analogous to the seasonal footprinting mechanism (SFM) proposed for the extratropical atmospheric forcing of ENSO events (Vimont et al. 2001; Anderson 2003; Chang et al. 2007; Alexander 2010). These studies show that the SST footprint imparted by the wintertime NPO drives surface wind changes in the western equatorial Pacific during the subsequent seasons, exciting oceanic equatorial Kelvin waves and triggering ENSO events. Ma et al. (2017) suggest that stochastic variability of the NPO is the primary source of uncertainty in the seasonal ENSO forecasts. Recent studies suggest that the same mechanism also works in the South Pacific and this South Pacific SFM may play an equally or more important role than the North Pacific counterpart (Zhang et al. 2014a; Min et al. 2015; Ding et al. 2017; You and Furtado 2017). The pattern of atmospheric variability associated with the South Pacific SFM resembles the PSA pattern, which exhibits maximum variability during austral winter. The concurrence of peak PSA variability and ENSO development makes it difficult to assess the causality of tropical and South Pacific variability and the relative importance of the South and North Pacific SFMs. Regardless, the significant role of the North Pacific SFM in ENSO needs to be reconciled with the weak impact of NPO on the equatorial Pacific shown in our study. One possible explanation is that thermodynamic air–sea interactions play a more important role than equatorial ocean dynamics on decadal time scales and thus the seasonal persistence of surface wind anomalies is required to induce robust response in the equatorial Pacific. On interannual time scales, on the other hand, a brief equatorward expansion of surface wind anomalies may be effective at triggering oceanic equatorial waves and ENSO events. A recent study by Liguori and Di Lorenzo (2019) indeed suggests that the relative importance of the North and South Pacific SFMs depends on the time scales of variability, with the

South Pacific SFM becoming more important on decadal than interannual time scales.

Our results indicate a strong linkage of the South Pacific to the tropical Pacific through thermodynamic air–sea interactions due to the northward position of the ITCZ and associated momentum advection, consistent with the multimodel analysis by Clement et al. (2011). However, some other studies using different atmospheric models coupled to a mixed-layer ocean show that ocean–atmosphere variability in the North Pacific strongly influences the equatorial Pacific on decadal time scales (Dommenges and Latif 2008; Yu et al. 2012; Di Lorenzo et al. 2015). Further studies are needed to understand the source of model discrepancy in the relative role of the South and North Pacific in TPDV.

b. Oceanic adjustment in the PSA experiment

In the PSA experiment, the equatorial Pacific thermocline shoals across the basin and the associated changes in the vertical ocean heat transport act to cool equatorial SSTs (Fig. 10). This cooling effect counteracts the warming induced by the WES feedback and weakens the equatorial Pacific SST response to the PSA forcing. In the time mean response, the equatorial thermocline shoaling is associated with negative wind stress curl anomalies over the tropical South Pacific and resultant southward Sverdrup transport (Fig. 9). Since it takes longer for the equatorial ocean to adjust to off-equatorial surface wind anomalies than for the WES feedback to extend subtropical SST anomalies toward the equator, the delay may set the time scale of decadal variability in the tropical Pacific, and hence the oscillation mechanism. We investigate the initial ocean adjustment to the PSA forcing by combining the PSA and PSA-15 experiments for the first 15 years. Figure 11 shows longitude–time sections of the 20-member ensemble mean response of the isopycnal depth along the equator and three different latitude bands in the tropical South Pacific, together with equatorial SSTs and tropical South Pacific wind stress curl. Note that the longitude axis is flipped in the tropical South Pacific panels. After the PSA surface heat flux forcing is imposed, negative wind stress curl anomalies develop quickly over the tropical South Pacific and persist throughout the experiments. These wind stress curl anomalies, in turn, force isopycnal shoaling in the central tropical South Pacific that successively propagates to the west. The time for the isopycnal depth anomalies to travel to the western boundary varies with latitudes from less than 1 year between 6° and 10°S to 3 years between 15° and 20°S, roughly in agreement with the phase speeds of the first baroclinic Rossby waves (e.g., White et al. 2003). Upon arriving at the western boundary, these Rossby

FIG. 11. Longitude–time sections of transient response of isopycnal depth ($\sigma_\theta = 25.5 \text{ kg m}^{-3}$; m; shading) averaged in (left three panels) three latitude bands in the tropical South Pacific (15°–20°S, 10°–15°S, and 6°–10°S) and (right) the equatorial Pacific (3°N–3°S) in the joint ensemble mean of the PSA and PSA-15 experiments. The stippled area indicates where the ensemble mean anomalies are significantly different from zero at the 90% confidence level based on a two-tailed Student's t test. The leftmost and rightmost panels additionally show, respectively, the monthly response of surface wind stress curl in the tropical South Pacific (6°–20°S; contours at intervals of $6 \times 10^{-9} \text{ N m}^{-3}$; zero contours thickened) and SST in the equatorial Pacific (3°N–3°S; contours at intervals of 0.3°C; zero contours thickened). The wind stress curl anomalies are smoothed with a 6-month running mean filter.

waves turn into Kelvin waves and propagate eastward along the equator, shoaling the equatorial thermocline. The arrival of initial upwelling Rossby waves from 10°–20°S in years 3–4 appears to trigger interannual oscillation along the equator with alternating sign of SST and isopycnal depth anomalies. The warm phase of this equatorial oscillation intensifies negative wind stress curl anomalies over the tropical South Pacific and forces strong upwelling Rossby waves, which subsequently propagate into the equatorial Pacific in years 7–8 and 10–11. The equatorial cooling caused by these successive upwelling Kelvin waves starting from years 3–4 would force a negative PSA pattern of atmospheric circulation anomalies and may trigger the phase reversal of decadal variability in the absence of the imposed surface heat flux forcing.

In addition to the wind-driven oceanic waves, isopycnal advection of temperature anomalies subducted in the subtropical South Pacific may also play a role in the equatorial ocean adjustment on decadal and longer time scales (Liu and Yang 2003; Yeager and Large 2004; Nonaka and Sasaki 2007; Matei et al. 2008; Tatebe et al. 2013). The isopycnal surfaces that feed into the lower equatorial thermocline ($\sigma_\theta \sim 25.5\text{--}26 \text{ kg m}^{-3}$; Fig. S4)

outcrop in the subtropical South Pacific where negative SST anomalies are present in the PSA and PSA-15 experiments (Fig. 12). The subduction and advection of these negative SST anomalies may contribute to the

FIG. 12. Annual mean SST response (°C; shading) and surface density climatology (black contours at intervals of 0.5 kg m^{-3}) in the ensemble mean of the PSA experiment.

equatorial ocean temperature cooling that maximizes in the lower thermocline in the PSA experiment (Fig. 8). These oceanic processes are, however, difficult to discern from the oceanic waves forced by tropical surface wind changes, which dominate the oceanic adjustment in the PSA and PSA-15 experiments. It is also noted that the subduction and advection processes do not cause a long-term drift in the equatorial Pacific state beyond the first 15 years of the PSA experiment, although substantial natural variability present in each member of this experiment makes it difficult to identify such oceanic adjustment processes. The importance of tropical surface winds in mediating the oceanic adjustment to subtropical/extratropical thermal forcing agrees with previous idealized model studies (Yang et al. 2004; Yang and Liu 2005). This result is also consistent with the finding that observed decadal variations of the tropical Pacific are mostly forced by changes in local winds (Schneider et al. 1999a,b; Hazeleger et al. 2001; Karspeck and Cane 2002).

The delayed negative oceanic feedback through wind-driven oceanic waves found in the PSA and PSA-15 experiments has important implications for the predictability of TPDV. While stochastically driven TPDV is fundamentally unpredictable, once the pattern of TPDV is established, the oceanic wave adjustment may trigger the phase reversal of TPDV. More observational and modeling analyses are needed to understand the role of wind-driven oceanic adjustment in the phase reversal of TPDV and the potential role of isopycnal advection of temperature anomalies in tropical Pacific variability on longer time scales.

c. Role of oceanic adjustment in TPDV in the CCSM4 control simulation

We examine whether the wind-driven oceanic adjustment identified in the PSA and PSA-15 experiments plays a role in setting the time scales of the leading mode of TPDV in the CCSM4 control simulation (Fig. 1). Figure 13 shows lagged regression of SST, surface wind, wind stress curl, and isopycnal depth anomalies on the leading PC of TPDV. During the positive phase of TPDV, the southeastern tropical Pacific warming is associated with basinwide cyclonic circulation anomalies over the tropical South Pacific, which force isopycnal shoaling in the underlying subsurface ocean. This isopycnal shoaling propagates from the central to the western tropical South Pacific and maximizes at the western boundary 2–3 years after the peak of the warm phase, consistent with the phase speeds of the first baroclinic Rossby waves. As the Rossby waves reach the western boundary, the equatorial thermocline shoals across the basin and equatorial SST anomalies transition

into the cold phase 4 years after the peak of the warm phase. This oceanic adjustment in the tropical South Pacific is generally consistent with the initial oceanic adjustment in the PSA and PSA-15 experiments (Fig. 11). The oceanic adjustment in the tropical North Pacific, in contrast, appears to play a minimal role in the phase transition of TPDV. Positive wind stress curl associated with anomalous equatorial westerly winds is limited to the south of 10°N and the resultant negative feedback to the equatorial thermocline would not provide sufficient lag for decadal variability. Outside the equatorial region, surface wind anomalies do not show a zonally coherent pattern over the tropical North Pacific, which is important in enhancing Rossby waves (Capotondi et al. 2003). The isopycnal deepening induced by negative wind stress curl anomalies over the northeastern tropical Pacific does not reach the western boundary due to the opposite wind stress curl anomalies to the west. Thus, the oceanic adjustment in the tropical South Pacific plays a more important role in setting the decadal time scale of TPDV. We note that the time scale of oceanic adjustment described here is somewhat sensitive to the cutoff frequency of the low-pass filter. Equatorial Pacific variability simulated in CCSM4 has a sharp spectral peak in the 3–6-yr band (Deser et al. 2012), and applying a low-pass filter tends to yield an artificial spectral peak near the cutoff frequency.

The importance of wind-driven oceanic adjustment in the tropical South Pacific in setting the decadal time scale of TPDV is consistent with previous modeling studies (Luo et al. 2003; Tatebe et al. 2013). The decadal oceanic linkage between the equatorial and tropical South Pacific has also been found in limited observational data (Chang et al. 2001; Luo and Yamagata 2001; Giese et al. 2002; Hasegawa et al. 2007). In support of these findings, Tatebe et al. (2013) show that ENSO-like TPDV in their model disappears when the oceanic linkage is partially blocked around 10°S while the oceanic linkage from the North Pacific does not significantly affect TPDV. Our analysis further suggests that this asymmetric oceanic linkage of the equatorial Pacific with the South and North Pacific may have its origin in the atmosphere. The tropical atmospheric circulation anomalies driving the oceanic linkage are not symmetrical about the equator and basinwide circulation anomalies are present only in the tropical South Pacific. The larger atmospheric circulation anomalies over the tropical South Pacific may reflect the leading role of South Pacific atmospheric variability in driving TPDV (Okumura 2013). Alternatively, decadal changes in tropical Pacific SSTs may force asymmetric atmospheric response due to the asymmetry in tropical Pacific mean climate. In particular, tropical Pacific precipitation

FIG. 13. (a) Lagged regression maps of (left) SST (°C; shading) and surface wind (m s⁻¹; vectors) and (right) surface wind stress curl (10⁻⁹ N m⁻³; shading) and isopycnal depth ($\sigma_{\theta} = 25.5 \text{ kg m}^{-3}$; contours at intervals of 1 m; zero contours omitted) anomalies on the first PC of TPDV in the CCSM4 control simulation. The maps are shown for 0–6-yr lags at intervals of 2-yr lag. (b) Longitude–time sections of regression of isopycnal depth anomalies ($\sigma_{\theta} = 25.5 \text{ kg m}^{-3}$; m; shading) in the (left) tropical South Pacific (10°–20°S), (middle) equatorial Pacific (3°N–3°S), and (right) tropical North Pacific (10°–20°N) on the first PC of TPDV in the CCSM4 control simulation. The left and middle panels also show, respectively, the regression of surface wind stress curl (contours at intervals of $1 \times 10^{-9} \text{ N m}^{-3}$; zero contours omitted) and SST (black contours at intervals of 0.05°C; zero contours omitted) in the same latitude bands. In both (a) and (b), the PC is standardized and all variables are smoothed with a 10-yr low-pass filter prior to the regression analysis.

anomalies associated with TPDV are aligned with the climatological precipitation bands and extend further away from the equator over the South Pacific than the North Pacific due to the presence of the SPCZ (Fig. 1, top left). The resultant atmospheric diabatic heating is likely to induce larger circulation anomalies over the tropical South Pacific than the tropical North Pacific. Thus, the strong linkages between the equatorial and South Pacific suggested in previous studies may result from the interplay of oceanic and atmospheric processes.

5. Summary

A suite of CCSM4 experiments was conducted to investigate the influence of stochastic atmospheric variability over the South and North Pacific on TPDV. In particular, we focused on the role of internal atmospheric modes known as the PSA and NPO patterns, which strongly affect the trade winds over the subtropical South and North Pacific, respectively. The results show that both the PSA and NPO modes induce basinwide ocean–atmosphere anomalies in the tropical Pacific and beyond that resemble the observed IPO pattern. The diagnostic analysis suggests the following mechanisms by which the PSA and NPO modes affect the tropical Pacific. The surface wind anomalies associated with the PSA and NPO modes force subtropical SST anomalies through changes in surface heat fluxes. The subtropical SST anomalies, in turn, act to expand the wind anomalies toward the equator. The coupled wind–SST anomalies propagate into the equatorial Pacific through the WES feedback during the seasons when the ITCZ is located in the opposite hemisphere. The equatorial SST anomalies are also maintained by the low-level cloud-radiation feedback. The equatorial ocean dynamics play different roles in the PSA and NPO experiments. In the NPO experiment, a change in the zonal tilt of the equatorial thermocline forced by local zonal wind anomalies enhances the equatorial SST anomalies through the Bjerknes feedback. In the PSA experiment, on the other hand, a change in the zonal mean depth of the equatorial thermocline forced by wind stress curl anomalies in the tropical South Pacific dampens the equatorial SST anomalies induced by the WES feedback.

Our experiments show that on decadal time scales, the equatorial Pacific is more strongly influenced by the PSA than the NPO mode through both the atmospheric and oceanic pathways. Regarding the atmospheric pathway, the WES feedback communicates the subtropical wind and SST anomalies to the equatorial Pacific more effectively from the South Pacific than the North Pacific. This is because the background cross-equatorial

southerly winds associated with the northward displacement of the ITCZ advect zonal momentum anomalies from the South Pacific into the equatorial regions. Regarding the oceanic pathway, the equatorial Pacific thermocline is more strongly influenced by the tropical atmospheric circulation anomalies associated with the PSA mode than the NPO mode. This asymmetric oceanic linkage has an origin in the atmosphere: the tropical atmospheric circulation anomalies associated with the PSA mode are larger and more zonally coherent compared to those associated with the NPO mode, presumably due to the stronger equatorial linkage and the presence of the SPCZ. The resultant changes in surface wind stress curl force off-equatorial oceanic Rossby waves, which negatively feed back to the equatorial SST anomalies induced by the WES feedback several years behind. This delayed negative oceanic feedback from the South Pacific may be important for setting the time scales of TPDV, as supported by the analysis of the CCSM4 control simulation. The key atmospheric and oceanic processes analyzed in this study provide insights into not only the origin but also the potential predictability of TPDV.

Okumura et al. (2017) suggest that TPDV forced by stochastic atmospheric variability interacts with the ENSO dynamics and causes decadal modulation in the amplitude and other characteristics of the ENSO. In support of this hypothesis, the CCSM4 experiments conducted in this study show significant changes in the ENSO. It is important to note, however, that neither the NPO nor the PSA2 experiment reproduces the second leading mode of TPDV in the CCSM4 control simulation, which is strongly linked to the decadal modulation of the ENSO amplitude. This result suggests that the second TPDV mode is more likely induced by decadal changes in the ENSO amplitude, rather than the other way around. The detailed analysis of the ENSO modulation in the CCSM4 experiments will be reported in a separate paper.

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REFERENCES

- Adler, R. F., and Coauthors, 2003: The version-2 Global Precipitation Climatology Project (GPCP) monthly precipitation analysis (1979–present). *J. Hydrometeorol.*, **4**, 1147–1167, [https://doi.org/10.1175/1525-7541\(2003\)004<1147:TVGPCP>2.0.CO;2](https://doi.org/10.1175/1525-7541(2003)004<1147:TVGPCP>2.0.CO;2).
- Alexander, M. A., 2010: Extratropical air–sea interaction, sea surface temperature variability, and the Pacific decadal oscillation. *Climate Dynamics: Why Does Climate Vary?* *Geophys. Monogr.*, Vol. 189, Amer. Geophys. Union, 123–148.
- , and C. Deser, 1995: A mechanism for the recurrence of wintertime midlatitude SST anomalies. *J. Phys. Oceanogr.*, **25**, 122–137, [https://doi.org/10.1175/1520-0485\(1995\)025<0122:AMFTRO>2.0.CO;2](https://doi.org/10.1175/1520-0485(1995)025<0122:AMFTRO>2.0.CO;2).
- , D. J. Vimont, P. Chang, and J. D. Scott, 2010: The impact of extratropical atmospheric variability on ENSO: Testing the seasonal footprinting mechanism using coupled model experiments. *J. Climate*, **23**, 2885–2901, <https://doi.org/10.1175/2010JCLI3205.1>.
- Anderson, B. T., 2003: Tropical Pacific sea-surface temperatures and preceding sea level pressure anomalies in the subtropical North Pacific. *J. Geophys. Res.*, **108**, 4732, <https://doi.org/10.1029/2003JD003805>.
- Ault, T. R., C. Deser, M. Newman, and J. Emile-Geay, 2013: Characterizing decadal to centennial variability in the equatorial Pacific during the last millennium. *Geophys. Res. Lett.*, **40**, 3450–3456, <https://doi.org/10.1002/grl.50647>.
- Bailey, D., C. Hannay, M. Holland, and R. Neale, 2009: Slab ocean model forcing. NCAR Tech. Note, 2 pp., <http://www.cesm.ucar.edu/models/cesm1.2/data8/doc/SOM.pdf>.
- Barnett, T. P., D. W. Pierce, M. Latif, D. Dommenges, and R. Saravanan, 1999: Interdecadal interactions between the tropics and midlatitudes in the Pacific basin. *Geophys. Res. Lett.*, **26**, 615–618, <https://doi.org/10.1029/1999GL900042>.
- Bellenger, H., E. Guilyardi, J. Leloup, M. Lengaigne, and J. Vialard, 2014: ENSO representation in climate models: From CMIP3 to CMIP5. *Climate Dyn.*, **42**, 1999–2018, <https://doi.org/10.1007/s00382-013-1783-z>.
- Branstator, G., 1995: Organization of storm track anomalies by recurring low-frequency circulation anomalies. *J. Atmos. Sci.*, **52**, 207–226, [https://doi.org/10.1175/1520-0469\(1995\)052<0207:OOSTAB>2.0.CO;2](https://doi.org/10.1175/1520-0469(1995)052<0207:OOSTAB>2.0.CO;2).
- Capotondi, A., 2013: ENSO diversity in the NCAR CCSM4 climate model. *J. Geophys. Res.*, **118**, 4755–4770, <https://doi.org/10.1002/jgrc.20335>.
- , M. A. Alexander, and C. Deser, 2003: Why are there Rossby wave maxima in the Pacific at 10°S and 13°N? *J. Phys. Oceanogr.*, **33**, 1549–1563, <https://doi.org/10.1175/2407.1>.
- Cayan, D. R., 1992: Latent and sensible heat flux anomalies over the northern oceans: Driving the sea surface temperature. *J. Phys. Oceanogr.*, **22**, 859–881, [https://doi.org/10.1175/1520-0485\(1992\)022<0859:LASHFA>2.0.CO;2](https://doi.org/10.1175/1520-0485(1992)022<0859:LASHFA>2.0.CO;2).
- Chang, P., B. Giese, L. Ji, H. Seidel, and F. Wang, 2001: Decadal change in the south tropical Pacific in a global assimilation analysis. *Geophys. Res. Lett.*, **28**, 3461–3464, <https://doi.org/10.1029/2001GL013018>.
- , L. Zhang, R. Saravanan, D. J. Vimont, J. C. Chiang, L. Ji, H. Seidel, and M. K. Tippett, 2007: Pacific meridional mode and El Niño–Southern Oscillation. *Geophys. Res. Lett.*, **34**, L16608, <https://doi.org/10.1029/2007GL030302>.
- Chen, X., and J. M. Wallace, 2015: ENSO-like variability: 1900–2013. *J. Climate*, **28**, 9623–9641, <https://doi.org/10.1175/JCLI-D-15-0322.1>.
- Choi, J., S.-I. An, B. Dewitte, and W. W. Hsieh, 2009: Interactive feedback between the tropical Pacific decadal oscillation and ENSO in a coupled general circulation model. *J. Climate*, **22**, 6597–6611, <https://doi.org/10.1175/2009JCLI2782.1>.
- Clement, A., P. DiNezio, and C. Deser, 2011: Rethinking the ocean’s role in the Southern Oscillation. *J. Climate*, **24**, 4056–4072, <https://doi.org/10.1175/2011JCLI3973.1>.
- Dai, A., J. C. Fyfe, S.-P. Xie, and X. Dai, 2015: Decadal modulation of global surface temperature by internal climate variability. *Nat. Climate Change*, **5**, 555–559, <https://doi.org/10.1038/nclimate2605>.
- Danabasoglu, G., S. C. Bates, B. P. Briegleb, S. R. Jayne, M. Jochum, W. G. Large, S. Peacock, and S. Yeager, 2012: The CCSM4 ocean component. *J. Climate*, **25**, 1361–1389, <https://doi.org/10.1175/JCLI-D-11-00091.1>.
- Dee, D. P., and Coauthors, 2011: The ERA-Interim reanalysis: Configuration and performance of the data assimilation system. *Quart. J. Roy. Meteor. Soc.*, **137**, 553–597, <https://doi.org/10.1002/qj.828>.
- Deser, C., A. S. Phillips, and J. W. Hurrell, 2004: Pacific interdecadal climate variability: Linkages between the tropics and the North Pacific during boreal winter since 1900. *J. Climate*, **17**, 3109–3124, [https://doi.org/10.1175/1520-0442\(2004\)017<3109:PICVLB>2.0.CO;2](https://doi.org/10.1175/1520-0442(2004)017<3109:PICVLB>2.0.CO;2).
- , M. A. Alexander, S.-P. Xie, and A. S. Phillips, 2010: Sea surface temperature variability: Patterns and mechanisms. *Annu. Rev. Mar. Sci.*, **2**, 115–143, <https://doi.org/10.1146/annurev-marine-120408-151453>.
- , and Coauthors, 2012: ENSO and Pacific decadal variability in the Community Climate System Model version 4. *J. Climate*, **25**, 2622–2651, <https://doi.org/10.1175/JCLI-D-11-00301.1>.
- Di Lorenzo, E., G. Liguori, N. Schneider, J. C. Furtado, B. T. Anderson, and M. A. Alexander, 2015: ENSO and meridional modes: A null hypothesis for Pacific climate variability. *Geophys. Res. Lett.*, **42**, 9440–9448, <https://doi.org/10.1002/2015GL066281>.
- Ding, R., J. Li, Y. Tseng, C. Sun, and F. Xie, 2017: Joint impact of North and South Pacific extratropical atmospheric variability on the onset of ENSO events. *J. Geophys. Res. Atmos.*, **122**, 279–298, <https://doi.org/10.1002/2016JD025502>.
- Dommenges, D., 2010: The slab ocean El Niño. *Geophys. Res. Lett.*, **37**, L20701, <https://doi.org/10.1029/2010GL044888>.
- , and M. Latif, 2008: Generation of hyper climate modes. *Geophys. Res. Lett.*, **35**, L02706, <https://doi.org/10.1029/2007GL031087>.
- England, M. H., and Coauthors, 2014: Recent intensification of wind-driven circulation in the Pacific and the ongoing warming hiatus. *Nat. Climate Change*, **4**, 222–227, <https://doi.org/10.1038/nclimate2106>.
- Fedorov, A. V., and S. G. Philander, 2000: Is El Niño changing? *Science*, **288**, 1997–2002, <https://doi.org/10.1126/science.288.5473.1997>.
- Feldstein, S. B., 2000: The timescale, power spectra, and climate noise properties of teleconnection patterns. *J. Climate*, **13**, 4430–4440, [https://doi.org/10.1175/1520-0442\(2000\)013<4430:TTPSAC>2.0.CO;2](https://doi.org/10.1175/1520-0442(2000)013<4430:TTPSAC>2.0.CO;2).
- Frankignoul, C., and K. Hasselmann, 1977: Stochastic climate models. Part II: Application to sea-surface temperature anomalies and thermocline variability. *Tellus*, **29**, 289–305, <https://doi.org/10.3402/tellusa.v29i4.11362>.
- Garreaud, R., and D. S. Battisti, 1999: Interannual (ENSO) and interdecadal (ENSO-like) variability in the Southern Hemisphere tropospheric circulation. *J. Climate*, **12**,

- 2113–2123, [https://doi.org/10.1175/1520-0442\(1999\)012<2113:IEAIEL>2.0.CO;2](https://doi.org/10.1175/1520-0442(1999)012<2113:IEAIEL>2.0.CO;2).
- Gent, P. R., and Coauthors, 2011: The Community Climate System Model version 4. *J. Climate*, **24**, 4973–4991, <https://doi.org/10.1175/2011JCLI4083.1>.
- Ghil, M., and K. Mo, 1991: Intraseasonal oscillations in the global atmosphere. Part II: Southern Hemisphere. *J. Atmos. Sci.*, **48**, 780–790, [https://doi.org/10.1175/1520-0469\(1991\)048<0780:IOITGA>2.0.CO;2](https://doi.org/10.1175/1520-0469(1991)048<0780:IOITGA>2.0.CO;2).
- Giese, B. S., C. S. Urizar, and N. S. Fučkar, 2002: Southern Hemisphere origins of the 1976 climate shift. *Geophys. Res. Lett.*, **29**, 1014, <https://doi.org/10.1029/2001GL013268>.
- Gu, D., and S. G. H. Philander, 1997: Interdecadal climate fluctuations that depend on exchanges between the tropics and extratropics. *Science*, **275**, 805–807, <https://doi.org/10.1126/science.275.5301.805>.
- Hanawa, K., and S. Sugimoto, 2004: ‘Reemergence’ areas of winter sea surface temperature anomalies in the world’s oceans. *Geophys. Res. Lett.*, **31**, L10303, <https://doi.org/10.1029/2004GL019904>.
- Hasegawa, T., T. Yasuda, and K. Hanawa, 2007: Generation mechanism of quasidecadal variability of upper ocean heat content in the equatorial Pacific Ocean. *J. Geophys. Res.*, **112**, C08012, <https://doi.org/10.1029/2006JC003755>.
- Hasselmann, K., 1976: Stochastic climate models. Part I: Theory. *Tellus*, **28**, 473–485, <https://doi.org/10.3402/tellusa.v28i6.11316>.
- Hawkins, E., and R. Sutton, 2009: The potential to narrow uncertainty in regional climate predictions. *Bull. Amer. Meteor. Soc.*, **90**, 1095–1108, <https://doi.org/10.1175/2009BAMS2607.1>.
- Hazeleger, W., M. Visbeck, M. Cane, A. Karspeck, and N. Naik, 2001: Decadal upper ocean temperature variability in the tropical Pacific. *J. Geophys. Res.*, **106**, 8971–8988, <https://doi.org/10.1029/2000JC000536>.
- Hurrell, J., and Coauthors, 2013: The Community Earth System Model: A framework for collaborative research. *Bull. Amer. Meteor. Soc.*, **94**, 1339–1360, <https://doi.org/10.1175/BAMS-D-12-00121.1>.
- Jin, F., 2001: Low-frequency modes of tropical ocean dynamics. *J. Climate*, **14**, 3874–3881, [https://doi.org/10.1175/1520-0442\(2001\)014<3874:LFMOTO>2.0.CO;2](https://doi.org/10.1175/1520-0442(2001)014<3874:LFMOTO>2.0.CO;2).
- Johnson, G., and M. McPhaden, 1999: Interior pycnocline flow from the subtropical to the equatorial Pacific Ocean. *J. Phys. Oceanogr.*, **29**, 3073–3089, [https://doi.org/10.1175/1520-0485\(1999\)029<3073:IPFFTS>2.0.CO;2](https://doi.org/10.1175/1520-0485(1999)029<3073:IPFFTS>2.0.CO;2).
- Karspeck, A. R., and M. A. Cane, 2002: Tropical Pacific 1976–77 climate shift in a linear, wind-driven model. *J. Phys. Oceanogr.*, **32**, 2350–2360, [https://doi.org/10.1175/1520-0485\(2002\)032<2350:TPCSIA>2.0.CO;2](https://doi.org/10.1175/1520-0485(2002)032<2350:TPCSIA>2.0.CO;2).
- Kay, J., and Coauthors, 2015: The Community Earth System Model (CESM) large ensemble project: A community resource for studying climate change in the presence of internal climate variability. *Bull. Amer. Meteor. Soc.*, **96**, 1333–1349, <https://doi.org/10.1175/BAMS-D-13-00255.1>.
- Knutson, T. R., and S. Manabe, 1998: Model assessment of decadal variability and trends in the tropical Pacific Ocean. *J. Climate*, **11**, 2273–2296, [https://doi.org/10.1175/1520-0442\(1998\)011<2273:MAODVA>2.0.CO;2](https://doi.org/10.1175/1520-0442(1998)011<2273:MAODVA>2.0.CO;2).
- Kosaka, Y., and S.-P. Xie, 2013: Recent global-warming hiatus tied to equatorial Pacific surface cooling. *Nature*, **501**, 403–407, <https://doi.org/10.1038/nature12534>.
- Liguori, G., and E. Di Lorenzo, 2019: Separating the North and South Pacific meridional modes contributions to ENSO and tropical decadal variability. *Geophys. Res. Lett.*, **46**, 906–915, <https://doi.org/10.1029/2018GL080320>.
- Lin, J.-L., 2007: The double-ITCZ problem in IPCC AR4 coupled GCMs: Ocean–atmosphere feedback analysis. *J. Climate*, **20**, 4497–4525, <https://doi.org/10.1175/JCLI4272.1>.
- Liu, Z., 2012: Dynamics of interdecadal climate variability: A historical perspective. *J. Climate*, **25**, 1963–1995, <https://doi.org/10.1175/2011JCLI3980.1>.
- , and S.-P. Xie, 1994: Equatorward propagation of coupled air–sea disturbances with application to the annual cycle of the eastern tropical Pacific. *J. Atmos. Sci.*, **51**, 3807–3822, [https://doi.org/10.1175/1520-0469\(1994\)051<3807:EPOCAD>2.0.CO;2](https://doi.org/10.1175/1520-0469(1994)051<3807:EPOCAD>2.0.CO;2).
- , and H. Yang, 2003: Extratropical control of tropical climate, the atmospheric bridge and oceanic tunnel. *Geophys. Res. Lett.*, **30**, 1230, <https://doi.org/10.1029/2002GL016492>.
- Lu, P., and J. McCreary, 1995: Influence of the ITCZ on the flow of thermocline water from the subtropical to the equatorial Pacific Ocean. *J. Phys. Oceanogr.*, **25**, 3076–3088, [https://doi.org/10.1175/1520-0485\(1995\)025<3076:IOTIOT>2.0.CO;2](https://doi.org/10.1175/1520-0485(1995)025<3076:IOTIOT>2.0.CO;2).
- Luo, J., and T. Yamagata, 2001: Long-term El Niño–Southern Oscillation (ENSO)-like variation with special emphasis on the South Pacific. *J. Geophys. Res.*, **106**, 22 211–22 227, <https://doi.org/10.1029/2000JC000471>.
- , S. Masson, S. Behera, P. Delecluse, S. Gualdi, A. Navarra, and T. Yamagata, 2003: South Pacific origin of the decadal ENSO like variation as simulated by a coupled GCM. *Geophys. Res. Lett.*, **30**, 2250, <https://doi.org/10.1029/2003GL018649>.
- Ma, J., S.-P. Xie, and H. Xu, 2017: Contributions of the North Pacific meridional mode to ensemble spread of ENSO prediction. *J. Climate*, **30**, 9167–9181, <https://doi.org/10.1175/JCLI-D-17-0182.1>.
- Mantua, N. J., S. R. Hare, Y. Zhang, J. M. Wallace, and R. C. Francis, 1997: A Pacific interdecadal climate oscillation with impacts on salmon production. *Bull. Amer. Meteor. Soc.*, **78**, 1069–1080, [https://doi.org/10.1175/1520-0477\(1997\)078<1069:APICOW>2.0.CO;2](https://doi.org/10.1175/1520-0477(1997)078<1069:APICOW>2.0.CO;2).
- Marshall, J., H. Johnson, and J. Goodman, 2001: A study of the interaction of the North Atlantic oscillation with ocean circulation. *J. Climate*, **14**, 1399–1421, [https://doi.org/10.1175/1520-0442\(2001\)014<1399:ASOTIO>2.0.CO;2](https://doi.org/10.1175/1520-0442(2001)014<1399:ASOTIO>2.0.CO;2).
- Matei, D., N. Keenlyside, M. Latif, and J. Jungclauss, 2008: Sub-tropical forcing of tropical Pacific climate and decadal ENSO modulation. *J. Climate*, **21**, 4691–4709, <https://doi.org/10.1175/2008JCLI2075.1>.
- Meehl, G. A., A. Hu, J. M. Arblaster, J. Fasullo, and K. E. Trenberth, 2013: Externally forced and internally generated decadal climate variability associated with the interdecadal Pacific oscillation. *J. Climate*, **26**, 7298–7310, <https://doi.org/10.1175/JCLI-D-12-00548.1>.
- Miller, A., D. Cayan, and W. White, 1998: A westward-intensified decadal change in the North Pacific thermocline and gyre-scale circulation. *J. Climate*, **11**, 3112–3127, [https://doi.org/10.1175/1520-0442\(1998\)011<3112:AWIDCI>2.0.CO;2](https://doi.org/10.1175/1520-0442(1998)011<3112:AWIDCI>2.0.CO;2).
- Min, Q., J. Su, R. Zhang, and X. Rong, 2015: What hindered the El Niño pattern in 2014? *Geophys. Res. Lett.*, **42**, 6762–6770, <https://doi.org/10.1002/2015GL064899>.
- Minobe, S., 1997: A 50–70 year climatic oscillation over the North Pacific and North America. *Geophys. Res. Lett.*, **24**, 683–686, <https://doi.org/10.1029/97GL00504>.
- Mitchell, T., and J. Wallace, 1992: The annual cycle in equatorial convection and sea-surface temperature. *J. Climate*, **5**, 1140–1156, [https://doi.org/10.1175/1520-0442\(1992\)005<1140:TACIEC>2.0.CO;2](https://doi.org/10.1175/1520-0442(1992)005<1140:TACIEC>2.0.CO;2).

- Newman, M., and Coauthors, 2016: The Pacific decadal oscillation, revisited. *J. Climate*, **29**, 4399–4427, <https://doi.org/10.1175/JCLI-D-15-0508.1>.
- Nitta, T., and S. Yamada, 1989: Recent warming of tropical sea surface temperature and its relationship to the Northern Hemisphere circulation. *J. Meteor. Soc. Japan*, **67**, 375–383, https://doi.org/10.2151/jmsj1965.67.3_375.
- Nonaka, M., and S.-P. Xie, 2000: Propagation of North Pacific interdecadal subsurface temperature anomalies in an ocean GCM. *Geophys. Res. Lett.*, **27**, 3747–3750, <https://doi.org/10.1029/2000GL011488>.
- , and H. Sasaki, 2007: Formation mechanism for isopycnal temperature–salinity anomalies propagating from the eastern South Pacific to the equatorial region. *J. Climate*, **20**, 1305–1315, <https://doi.org/10.1175/JCLI4065.1>.
- , S.-P. Xie, and K. Takeuchi, 2000: Equatorward spreading of a passive tracer with application to North Pacific interdecadal temperature variations. *J. Oceanogr.*, **56**, 173–183, <https://doi.org/10.1023/A:1011135113079>.
- Okajima, H., S.-P. Xie, and A. Numaguti, 2003: Interhemispheric coherence of tropical climate variability: Effect of the climatological ITCZ. *J. Meteor. Soc. Japan*, **81**, 1371–1386, <https://doi.org/10.2151/jmsj.81.1371>.
- Okumura, Y., 2013: Origins of tropical Pacific decadal variability: Role of stochastic atmospheric forcing from the South Pacific. *J. Climate*, **26**, 9791–9796, <https://doi.org/10.1175/JCLI-D-13-00448.1>.
- , and S.-P. Xie, 2004: Interaction of the Atlantic equatorial cold tongue and the African monsoon. *J. Climate*, **17**, 3589–3602, [https://doi.org/10.1175/1520-0442\(2004\)017<3589:IOTAEC>2.0.CO;2](https://doi.org/10.1175/1520-0442(2004)017<3589:IOTAEC>2.0.CO;2).
- , D. Schneider, C. Deser, and R. Wilson, 2012: Decadal–interdecadal climate variability over Antarctica and linkages to the tropics: Analysis of ice core, instrumental, and tropical proxy data. *J. Climate*, **25**, 7421–7441, <https://doi.org/10.1175/JCLI-D-12-00050.1>.
- , T. Sun, and X. Wu, 2017: Asymmetric modulation of El Niño and La Niña and the linkage to tropical Pacific decadal variability. *J. Climate*, **30**, 4705–4733, <https://doi.org/10.1175/JCLI-D-16-0680.1>.
- Partin, J., and Coauthors, 2013: Multidecadal rainfall variability in South Pacific convergence zone as revealed by stalagmite geochemistry. *Geology*, **41**, 1143–1146, <https://doi.org/10.1130/G34718.1>.
- Pierce, D. W., T. P. Barnett, and M. Latif, 2000: Connections between the Pacific Ocean tropics and midlatitudes on decadal timescales. *J. Climate*, **13**, 1173–1194, [https://doi.org/10.1175/1520-0442\(2000\)013<1173:CBTPOT>2.0.CO;2](https://doi.org/10.1175/1520-0442(2000)013<1173:CBTPOT>2.0.CO;2).
- Power, S., T. Casey, C. Folland, A. Colman, and V. Mehta, 1999: Inter-decadal modulation of the impact of ENSO on Australia. *Climate Dyn.*, **15**, 319–324, <https://doi.org/10.1007/s003820050284>.
- Rodgers, K. B., P. Friederichs, and M. Latif, 2004: Tropical Pacific decadal variability and its relation to decadal modulations of ENSO. *J. Climate*, **17**, 3761–3774, [https://doi.org/10.1175/1520-0442\(2004\)017<3761:TPDVAI>2.0.CO;2](https://doi.org/10.1175/1520-0442(2004)017<3761:TPDVAI>2.0.CO;2).
- Schneider, N., A. J. Miller, M. A. Alexander, and C. Deser, 1999a: Subduction of decadal North Pacific temperature anomalies: Observations and dynamics. *J. Phys. Oceanogr.*, **29**, 1056–1070, [https://doi.org/10.1175/1520-0485\(1999\)029<1056:SODNPT>2.0.CO;2](https://doi.org/10.1175/1520-0485(1999)029<1056:SODNPT>2.0.CO;2).
- , S. Venzke, A. Miller, D. Pierce, T. Barnett, C. Deser, and M. Latif, 1999b: Pacific thermocline bridge revisited. *Geophys. Res. Lett.*, **26**, 1329–1332, <https://doi.org/10.1029/1999GL900222>.
- Seager, R., Y. Kushnir, C. Herweijer, N. Naik, and J. Velez, 2005: Modeling of tropical forcing of persistent droughts and pluvials over western North America: 1856–2000. *J. Climate*, **18**, 4065–4088, <https://doi.org/10.1175/JCLI3522.1>.
- Taguchi, B., S. Xie, N. Schneider, M. Nonaka, H. Sasaki, and Y. Sasai, 2007: Decadal variability of the Kuroshio Extension: Observations and an eddy-resolving model hindcast. *J. Climate*, **20**, 2357–2377, <https://doi.org/10.1175/JCLI4142.1>.
- Tatebe, H., Y. Imada, M. Mori, M. Kimoto, and H. Hasumi, 2013: Control of decadal and bidecadal climate variability in the tropical Pacific by the off-equatorial South Pacific Ocean. *J. Climate*, **26**, 6524–6534, <https://doi.org/10.1175/JCLI-D-12-00137.1>.
- Trenberth, K. E., and J. W. Hurrell, 1994: Decadal atmosphere–ocean variations in the Pacific. *Climate Dyn.*, **9**, 303–319, <https://doi.org/10.1007/BF00204745>.
- Vimont, D. J., 2005: The contribution of the interannual ENSO cycle to the spatial pattern of decadal ENSO-like variability. *J. Climate*, **18**, 2080–2092, <https://doi.org/10.1175/JCLI3365.1>.
- , D. S. Battisti, and A. C. Hirst, 2001: Footprinting: A seasonal connection between the tropics and mid-latitudes. *Geophys. Res. Lett.*, **28**, 3923–3926, <https://doi.org/10.1029/2001GL013435>.
- Walker, G. T., and E. W. Bliss, 1932: World Weather V. *Mem. Roy. Meteor. Soc.*, **4**, 53–84.
- Wallace, J. M., I. M. Held, D. W. J. Thompson, K. E. Trenberth, and J. E. Walsh, 2014: Global warming and winter weather. *Science*, **343**, 729–730, <https://doi.org/10.1126/science.1243617>.
- Wang, B., and S. An, 2002: A mechanism for decadal changes of ENSO behavior: Roles of background wind changes. *Climate Dyn.*, **18**, 475–486, <https://doi.org/10.1007/s00382-001-0189-5>.
- Wang, D., and Z. Liu, 2000: The pathway of the interdecadal variability in the Pacific Ocean. *Chin. Sci. Bull.*, **45**, 1555–1561, <https://doi.org/10.1007/BF02886211>.
- White, W. B., Y. M. Tourre, M. Barlow, and M. Dettinger, 2003: A delayed action oscillator shared by biennial, interannual, and decadal signals in the Pacific basin. *J. Geophys. Res.*, **108**, 3070, <https://doi.org/10.1029/2002JC001490>.
- Wu, L., F. He, and Z. Liu, 2005: Coupled ocean–atmosphere response to north tropical Atlantic SST: Tropical Atlantic dipole and ENSO. *Geophys. Res. Lett.*, **32**, L21712, <https://doi.org/10.1029/2005GL024222>.
- Xie, S.-P., 1994: On the genesis of the equatorial annual cycle. *J. Climate*, **7**, 2008–2013, [https://doi.org/10.1175/1520-0442\(1994\)007<2008:OTGOTE>2.0.CO;2](https://doi.org/10.1175/1520-0442(1994)007<2008:OTGOTE>2.0.CO;2).
- , and S. G. H. Philander, 1994: A coupled ocean–atmosphere model of relevance to the ITCZ in the eastern Pacific. *Tellus*, **46A**, 340–350, <https://doi.org/10.3402/tellusa.v46i4.15484>.
- , C. Deser, G. Vecchi, J. Ma, H. Teng, and A. Wittenberg, 2010: Global warming pattern formation: Sea surface temperature and rainfall. *J. Climate*, **23**, 966–986, <https://doi.org/10.1175/2009JCLI3329.1>.
- Yang, H., and Z. Liu, 2005: Tropical–extratropical climate interaction as revealed in idealized coupled climate model experiments. *Climate Dyn.*, **24**, 863–879, <https://doi.org/10.1007/s00382-005-0021-8>.
- , —, and H. Wang, 2004: Influence of extratropical thermal and wind forcings on equatorial thermocline in an ocean GCM. *J. Phys. Oceanogr.*, **34**, 174–187, [https://doi.org/10.1175/1520-0485\(2004\)034<0174:IOETAW>2.0.CO;2](https://doi.org/10.1175/1520-0485(2004)034<0174:IOETAW>2.0.CO;2).
- Yeager, S. G., and W. G. Large, 2004: Late-winter generation of spiciness on subducted isopycnals. *J. Phys. Oceanogr.*, **34**, 1528–1547, [https://doi.org/10.1175/1520-0485\(2004\)034<1528:LGOSOS>2.0.CO;2](https://doi.org/10.1175/1520-0485(2004)034<1528:LGOSOS>2.0.CO;2).

- Yeh, S.-W., and B. Kirtman, 2004: The impact of internal atmospheric variability on the North Pacific SST variability. *Climate Dyn.*, **22**, 721–732, <https://doi.org/10.1007/s00382-004-0399-8>.
- You, Y., and J. C. Furtado, 2017: The role of South Pacific atmospheric variability in the development of different types of ENSO. *Geophys. Res. Lett.*, **44**, 7438–7446, <https://doi.org/10.1002/2017GL073475>.
- Yu, J.-Y., and S. T. Kim, 2011: Relationships between extratropical sea level pressure variations and the central Pacific and eastern Pacific types of ENSO. *J. Climate*, **24**, 708–720, <https://doi.org/10.1175/2010JCLI3688.1>.
- , M.-M. Lu, and S. T. Kim, 2012: A change in the relationship between tropical central Pacific SST variability and the extratropical atmosphere around 1990. *Environ. Res. Lett.*, **7**, 034025, <https://doi.org/10.1088/1748-9326/7/3/034025>.
- Yukimoto, S., M. Endoh, Y. Kitamura, A. Kitoh, T. Motoi, and A. Noda, 2000: ENSO-like interdecadal variability in the Pacific Ocean as simulated in a coupled general circulation model. *J. Geophys. Res.*, **105**, 13 945–13 963, <https://doi.org/10.1029/2000JC900034>.
- Zhang, H., A. Clement, and P. DiNezio, 2014a: The South Pacific meridional mode: A mechanism for ENSO-like variability. *J. Climate*, **27**, 769–783, <https://doi.org/10.1175/JCLI-D-13-00082.1>.
- , C. Deser, A. Clement, and R. Tomas, 2014b: Equatorial signatures of the Pacific meridional modes: Dependence on mean climate state. *Geophys. Res. Lett.*, **41**, 568–574, <https://doi.org/10.1002/2013GL058842>.
- Zhang, Y., J. M. Wallace, and D. S. Battisti, 1997: ENSO-like interdecadal variability: 1900–93. *J. Climate*, **10**, 1004–1020, [https://doi.org/10.1175/1520-0442\(1997\)010<1004:ELIV>2.0.CO;2](https://doi.org/10.1175/1520-0442(1997)010<1004:ELIV>2.0.CO;2).
- Zhou, Z., and J. A. Carton, 1998: Latent heat flux and interannual variability of the coupled atmosphere–ocean system. *J. Atmos. Sci.*, **55**, 494–501, [https://doi.org/10.1175/1520-0469\(1998\)055<0494:LHFAIV>2.0.CO;2](https://doi.org/10.1175/1520-0469(1998)055<0494:LHFAIV>2.0.CO;2).