Potential Impact of Preceding Aleutian Low Variation on El Niño–Southern Oscillation during the Following Winter

SHANGFENG CHEN AND WEN CHEN

Center for Monsoon System Research, Institute of Atmospheric Physics, Chinese Academy of Sciences, Beijing, China

RENGUANG WU

School of Earth Sciences, Zhejiang University, Hangzhou, and Center for Monsoon System Research, Institute of Atmospheric Physics, Chinese Academy of Sciences, Beijing, China

BIN YU

Climate Research Division, Environment and Climate Change Canada, Toronto, Ontario, Canada

HANS-F. GRAF

Center for Atmospheric Science, University of Cambridge, Cambridge, United Kingdom

(Manuscript received 24 September 2019, in final form 10 January 2020)

ABSTRACT

The present study reveals a close relation between the interannual variation of Aleutian low intensity (ALI) in March and the subsequent winter El Niño–Southern Oscillation (ENSO). When March ALI is weaker (stronger) than normal, an El Niño (a La Niña)–like sea surface temperature (SST) warming (cooling) tends to appear in the equatorial central-eastern Pacific during the subsequent winter. The physical process linking March ALI to the following winter ENSO is as follows. When March ALI is below normal, a notable atmospheric dipole pattern develops over the North Pacific, with an anticyclonic anomaly over the Aleutian region and a cyclonic anomaly over the subtropical west-central Pacific. The formation of the anomalous cyclone is attributed to feedback of the synoptic-scale eddy-to-mean-flow energy flux and associated vorticity transportation. Specifically, easterly wind anomalies over the midlatitudes related to the weakened ALI are accompanied by a decrease in synoptic-scale eddy activity, which forces an anomalous cyclone to its southern flank. The accompanying westerly wind anomalies over the tropical west-central Pacific induce SST warming in the equatorial central-eastern Pacific during the following summer–autumn via triggering eastward-propagating warm Kelvin waves, which may sustain and develop into an El Niño event during the following winter via positive air–sea feedback. The relation of March ALI with the following winter ENSO is independent of the preceding tropical Pacific SST, the preceding-winter North Pacific Oscillation, and the spring Arctic Oscillation. The results of this analysis may provide an additional source for the prediction of ENSO.

1. Introduction

El Niño–Southern Oscillation (ENSO) is characterized by large sea surface temperature (SST) anomalies in the tropical central-eastern Pacific and related atmospheric mass oscillation between tropical western and eastern Pacific. It is the strongest air–sea coupled mode over the tropics on the interannual time scale and has pronounced influences on climate and weather over

the Pacific and remote regions via atmospheric teleconnections (e.g., [Lau and Nath 1996](#); [Trenberth 1997](#); [Zhang et al. 1997](#); [Wang et al. 2000](#); [Alexander et al. 2002](#); [Wu et al. 2003](#); [McPhaden et al. 2006](#); [Yu and Zwiers 2007](#); [Luo et al. 2010](#); [Song et al. 2017](#); [Chen and Song 2019](#), and references therein).

Substantial efforts have been devoted over the past few decades to understanding of the dynamics underlying ENSO. On one hand, previous studies showed that oceanic dynamics and air–sea interaction within the tropical Pacific play an important role in the occurrence, development, and decaying of an ENSO event (e.g.,

Corresponding author: Dr. Shangfeng Chen, chenshangfeng@mail.iap.ac.cn

Bjerknes 1969; Battisti 1988; Schopf and Suarez 1988; Jin 1997; Wang et al. 1999). The Bjerknes positive air–sea feedback is crucial for the development of the SST anomalies in the tropical central-eastern Pacific related to ENSO events (Bjerknes 1969). Tropical oceanic waves (westward-propagating Rossby waves and eastward-propagating Kelvin waves) can weaken SST anomalies in the tropical central-eastern Pacific, and contribute to the transition of ENSO phases (Battisti 1988; Schopf and Suarez 1988; Jin 1997; Wang et al. 1999).

On the other hand, a number of previous studies indicated that forcings from the extratropics play non-negligible roles in modulating winter ENSO outbreaks (e.g., Li 1990; Nakamura et al. 2006, 2007; Wang et al. 2011; Wang et al. 2012; Ham et al. 2013; Zhu et al. 2016; Chen et al. 2014, 2015, 2017, 2018). For example, Li (1990) showed that a stronger East Asian winter monsoon (EAWM) could induce pronounced atmospheric convections over the tropical western North Pacific via intrusion of cold surges. The resultant atmospheric convection induces significant westerly wind bursts over the tropical western Pacific via a Gill-type atmospheric response, which further influences the occurrence of ENSO events in the subsequent winter. Westerly wind bursts over the tropical western Pacific are an important trigger for the ENSO outbreak via inducing eastward-propagating warm Kelvin waves (e.g., Barnett et al. 1989; McPhaden 1999; Huang et al. 2001; Lengaigne et al. 2004; Lian et al. 2014; Lai et al. 2015). Nakamura et al. (2006) found that the spring (March–April average) Arctic Oscillation (AO; Thompson and Wallace 1998) can influence the following winter ENSO via modulating the westerly wind anomalies over the tropical western Pacific. Chen et al. (2014) indicated that the formation of westerly anomalies over the tropical western Pacific during the positive phase of the spring AO is attributed to the interaction between synoptic-scale eddy activity and low-frequency mean flow over the North Pacific.

The extratropical North Pacific plays a crucial role in modulating ENSO occurrence (Vimont et al. 2001, 2003; Alexander et al. 2010; Yu and Kim 2011; Yu et al. 2012). Vimont et al. (2001, 2003) found that the winter North Pacific Oscillation (NPO) can affect the occurrence of ENSO in the following winter via the so-called seasonal footprinting mechanism. NPO is the second empirical orthogonal function (EOF) mode of sea level pressure (SLP) anomalies over the extratropical North Pacific and features an oscillation in SLP anomalies between the subtropics and midlatitudes of the North Pacific (Rogers 1981; Linkin and Nigam 2008; Yu and Kim 2011; Chen and Wu 2018). As demonstrated by previous studies, the first EOF mode of SLP anomalies over the

extratropical North Pacific generally represents the variation of the Aleutian low intensity (ALI) (e.g., Linkin and Nigam 2008; Yu and Kim 2011; Song and Duan 2015). Several studies have demonstrated that variations in ALI correlate well with the climate anomalies over the tropics and subtropics (e.g., Zhu and Wang 2010; Song and Duan 2015; Choi and Cha 2017). For instance, Zhu and Wang (2010) showed that the interannual variation of boreal winter ALI has a close relationship with the Australian summer monsoon via atmospheric teleconnection. Choi and Cha (2017) found that the summer tropical cyclone activity over the tropical western Pacific is influenced by the preceding winter's ALI variation.

Previous studies generally indicated the strong impact of ENSO on the ALI variation, rather than the opposite influence of ALI on ENSO (Horel and Wallace 1981; Gershunov and Barnett 1998; Straus and Shukla 2000; Alexander et al. 2002; Mo 2010; Yu and Kim 2011). It is unknown whether there is a possible influence of ALI on ENSO. In this study, we will present observational evidences to reveal a close connection between the ALI variation in March and the subsequent winter ENSO. We will show that the influence of March ALI on the subsequent winter ENSO is independent of the ENSO cycle (i.e., it still exists when previous winter and simultaneous ENSO signals have been removed), preceding winter NPO, and spring AO. The March ALI variation may provide an additional source of predictability to improve the prediction of the following winter ENSO. In particular, previous studies indicated that predicting an ENSO event prior to the spring is difficult due to the so-called spring predictability barrier phenomenon of ENSO forecasts (Webster and Yang 1992; McPhaden 2003; Luo et al. 2008). The result of this study may have implications for the unresolved issue of the spring predictability barrier.

The rest of the paper is organized as follows. Section 2 describes the datasets and analysis methods. Section 3 presents observational evidence to demonstrate the connection between March ALI and following winter ENSO. Section 4 investigates the physical process connecting March ALI to subsequent winter ENSO. Section 5 provides a discussion of the findings. The key findings of this study are summarized in section 6.

2. Data and methodology

a. Data

The present study employs the monthly mean SST from the National Oceanic and Atmospheric Administration (NOAA) Extended Reconstructed SST, version 3b

(ERSSTv3b) (Smith et al. 2008). This SST dataset has a horizontal resolution of $2^\circ \times 2^\circ$ and is available from 1854 to the present. The monthly mean SLP, surface wind stress, horizontal winds, geopotential height, surface heat fluxes (including surface latent and sensible heat fluxes, surface shortwave and longwave radiation), and daily mean geopotential height are obtained from the National Centers for Environmental Prediction–Department of Energy (NCEP–DOE) from 1979 to the present (Kanamitsu et al. 2002). Atmospheric variables from the NCEP–DOE reanalysis are available on horizontal $2.5^\circ \times 2.5^\circ$ grids and surface heat fluxes are on T62 Gaussian grids. The monthly mean precipitation data are obtained from the Global Precipitation Climatology Project from 1979 to the present (Adler et al. 2003). In addition, the monthly mean AO index is extracted from the Climate Prediction Center (<http://www.cpc.ncep.noaa.gov>).

This analysis focuses on variations on the interannual time scale. Thus, all monthly mean variables are subjected to a 2–7-yr bandpass Lanczos filter (Duchon 1979) to obtain their interannual components. Applying a 2–9-yr bandpass filter provides similar results. Significant levels of correlation and regression coefficients are estimated according to a two-tailed Student's *t* test.

b. Kelvin wave forcing

Previous studies suggested that the dynamic response of tropical ocean to the surface wind stress anomaly can be characterized by the equatorial oceanic Kelvin wave forcing function (KWF) (e.g., Battisti 1988; Vimont et al. 2003). Following these studies, equatorial oceanic KWF is written as follows:

$$K_f(x, \lambda) = \int_{30^\circ\text{S}}^{30^\circ\text{N}} \tau_x(x, y, \lambda) \psi_o(y) dy, \quad (1)$$

where x and y denote longitude and latitude, respectively; $\tau_x(x, y, \lambda)$ represents anomalous surface zonal wind stress at a lag time λ , and $\psi_o(y)$ denotes the meridional structure of the equatorial Kelvin waves (Battisti 1988; Vimont et al. 2003). The meridional structure of the equatorial Kelvin waves could be derived by solving the dynamic equation of the upper-layer ocean in the equatorial β plane (Gill 1980; Battisti 1988; Vimont et al. 2003). The equatorial Kelvin waves could only move eastward and their amplitudes decrease exponentially away from the equator. Positive (negative) values of KWF indicate that surface zonal wind stress anomalies trigger the eastward-propagating warm (cold) equatorial Kelvin waves that contribute to SST warming (cooling) (Battisti 1988; Vimont et al. 2003).

c. Extended EP flux

This study employs the extended Eliassen–Palm (EP) flux (Hendon and Hartmann 1985; Lau 1988; Chen et al. 2014) to qualitatively evaluate the dynamical interaction between synoptic-scale eddy activity and low-frequency mean flow. Following previous studies (Hendon and Hartmann 1985; Lau 1988), the horizontal component of the extended EP flux is written as follows:

$$E_u = \left[\frac{1}{2} (\overline{v'^2} - \overline{u'^2}) \mathbf{i}, -\overline{u'v'} \mathbf{j} \right] \times \cos \varphi, \quad (2)$$

where u' and v' are synoptic-scale zonal and meridional winds, respectively, and φ denotes the latitude. Synoptic-scale winds are obtained by applying a band-pass filter to the raw daily fields to isolate variation on the 2–8-day time scale. The overbar represents the March average.

d. Geopotential height tendency

As indicated by previous studies (Lau and Holopainen 1984; Lau 1988; Cai et al. 2007), feedback of the synoptic-scale eddy activity to the mean flow can be quantitatively estimated by the feedback term in the geopotential height tendency equation. The geopotential height tendency due to the eddy vorticity flux forcing is expressed as follows (Lau 1988; Cai et al. 2007):

$$F = \frac{f}{g} \nabla^{-2} [-\nabla \cdot (\nabla' \zeta')]. \quad (3)$$

Here g is the acceleration of gravity, f denotes the Coriolis parameter; ∇' and ζ' correspond to the synoptic-scale winds and vorticity, respectively, and ζ' is calculated according to the synoptic-scale winds.

3. Lead-lag correlation between Aleutian low variation and tropical Pacific SST

The EOF method is employed to identify the dominant patterns of interannual variation of monthly SLP anomalies over the extratropical North Pacific (20° – 70°N , 120°E – 100°W) during 1979–2016. SLP anomalies in the EOF analysis are weighted by the cosine of latitude to account for the decrease of area toward the North Pole (North et al. 1982a). Figure 1 shows the first and second EOF modes (EOF1 and EOF2) of SLP anomalies and the related principal component (PC) time series. EOF1 and EOF2 explain 29% and 22% of the SLP interannual variance, respectively. These two modes are separated from each other and from the other EOF modes according to the method of North et al. (1982b).

FIG. 1. The (a) first and (b) second EOF modes of interannual variations of monthly mean SLP anomalies from January 1979 to December 2016. The climatological annual cycle has been removed before the EOF analysis. (c),(d) The principal component (PC) time series corresponding to EOF1 and EOF2, respectively. The box in (a) is used to define the Aleutian low intensity (ALI) in this analysis.

EOF1 is characterized by large loading around the Aleutian region (Fig. 1a), representing interannual variability of the Aleutian low intensity, consistent with previous studies (e.g., Yu and Kim 2011; Song and Duan 2015). Based on the spatial distribution of the loading in Fig. 1a, an ALI index is defined as area-averaged SLP anomalies over 30° – 65° N, 160° E– 140° W (i.e., the boxed region in Fig. 1a). EOF2 features out-of-phase SLP anomalies between the subtropics (20° – 45° N) and mid-latitudes (50° – 70° N) (Fig. 2a), which bear a close resemblance to the NPO structure identified in previous studies (Walker and Bliss 1932; Wallace and Gutzler 1981; Rogers 1981; Linkin and Nigam 2008).

Previous studies have demonstrated that winter NPO can influence the following winter ENSO (Vimont et al. 2001, 2003; Alexander et al. 2010). Chen et al. (2013) further demonstrated that the Pacific center of the spring AO has a notable modulation of the relation of the winter NPO and the following winter ENSO. The Pacific center of the AO may have several relations with the Aleutian low (AL) (Thompson and Wallace 1998; Chen et al. 2014). However, it is unknown whether or

not the ALI variation is linked to the SST variations in the tropical central-eastern Pacific during the following winter. To address this issue, we calculate lead-lag correlation coefficients between the monthly ALI and the Niño-3.4 index, which is presented in Fig. 2. The Niño-3.4 index is defined as the area-average SST anomalies over 5° S– 5° N, 170° – 120° W, which is often used to represent ENSO variability (e.g., Anderson 2007; Deser et al. 2012). From Fig. 2, it is evident that ALI in March has a significant positive correlation with the Niño-3.4 index in the following autumn and winter, with the correlation coefficient between them exceeding 0.4, which is significant at the 99% confidence level. This indicates that ALI in March has a connection with the subsequent winter ENSO. In addition, significant negative correlations are found between July ALI and the Niño-3.4 index from the preceding April to the following winter (Fig. 2). This implies a quasi-simultaneous influence of SST anomalies in the tropical central-eastern Pacific on the boreal summer July ALI variation. Furthermore, marked connections can be observed between ALI in January and February and the preceding

FIG. 2. Lead-lag correlation coefficients between the monthly ALI and the Niño-3.4 index during 1979–2016. The y axis represents the month of ALI. The x axis denotes the month by which the ALI leads or lags the Niño-3.4 index. Positive (negative) value represents that the ALI index leads (lags) the Niño-3.4 index. The shading indicates the correlation coefficient passing the 95% confidence level.

Niño-3.4 index (Fig. 2), indicating the influence of the tropical Pacific SST anomalies on the ALI variation. The influence of the SST anomalies in the tropical Pacific on the ALI variation is consistent with previous studies (Gershunov and Barnett 1998; Alexander et al. 2002; Yu and Kim 2011). The present study is concerned with positive correlation between March ALI and the following autumn and winter Niño-3.4 index. As the ENSO-related SST anomalies have a phase-locking feature with a mature peak in winter (e.g., Wang et al. 2000; Chen 2002), the above positive correlation implies that the March ALI variability may have potential influence on the winter ENSO occurrence. It should be mentioned that ALIs in both FMA(0) (February–April average) and MA(0) (March–April average) have a statistically significant correlation with the following winter Niño-3.4 index. In addition, the results obtained in the following analysis are largely similar if based on the FMA or MA ALI, but with slight weaker amplitudes of the signals. In the following analysis, we will focus on the connection between March ALI and the subsequent winter ENSO.

Figure 3a displays the normalized March [Mar(0)] ALI and following winter [N(0)DJ(1) average] Niño-3.4 index during 1979–2016. In this study, the notations (0) and (1) refer to the year during and after the March ALI year, respectively. The correlation coefficient between these two indices reaches 0.49, significant at the 99% confidence level. In addition, most of the large values in the ND(0)J(1) Niño-3.4 index correspond to extreme Mar(0) ALI. For example, strong 1982/83, 1997/98, and

FIG. 3. (a) Normalized time series of interannual variations of March [Mar(0)] ALI and its following winter [ND(0)J(1) average] Niño-3.4 index during 1979–2016. (b) As in (a), but preceding winter ENSO signals have been subtracted from the Mar(0) ALI and ND(0)J(1) Niño-3.4 index by means of linear regression with respect to the D(-1)JFM(0) Niño-3.4 index. The time notations (-1), (0), and (1) refer to the year before, during, and after the March ALI year, respectively. Years are labeled according to the Mar(0) date.

2015/16 El Niño events are preceded by positive values of March ALI (larger than 0.5 standard deviations). Note that the correlation between Mar(0) ALI and subsequent winter Niño-3.4 index is 0.43 when preceding winter ENSO signals, represented by the Niño-3.4 index averaged from December(-1) to March(0) [D(-1)JFM(0)], have been removed by means of linear regression (Fig. 3b). This indicates that the connection between Mar(0) ALI and subsequent winter ENSO is independent of the ENSO cycle. In the following analysis, the preceding ENSO signal has been excluded from all the variables.

4. Factors contributing to the ALI-ENSO connection

a. Atmospheric circulation anomalies in March

This section analyzes the possible physical processes connecting Mar(0) ALI to the subsequent winter ENSO. We first inspect simultaneous atmospheric circulation anomalies. Figure 4 displays the patterns of 850-hPa winds and 500- and 300-hPa geopotential height anomalies in Mar(0) regressed onto the normalized ALI in

FIG. 4. Regression of (a) 850-hPa winds (ms^{-1}), and geopotential height anomalies (m) at (b) 500 and (c) 300 hPa in Mar(0) onto the normalized Mar(0) ALI. Shading in (a) indicates either component of the wind anomalies significant at the 95% confidence level. Shading in (b) and (c) indicates geopotential height anomalies significant at the 95% confidence level.

Mar(0). Atmospheric circulation anomalies over the North Pacific display a quasi-barotropic vertical structure in the troposphere (Fig. 4). A marked anticyclonic anomaly and above-average positive geopotential heights are observed over the midlatitudes (Fig. 4), indicating a weakened ALI. In addition, a significant cyclonic anomaly and below-average geopotential heights are present over the subtropical west-central North Pacific. Accordingly, there are strong easterly wind anomalies around 20°–40°N and westerly wind anomalies over high latitudes around 50°–60°N and the tropical west-central Pacific (Fig. 4a). Anomalous

westerly winds over the tropical west-central Pacific are crucial to the occurrence of an El Niño event via triggering eastward-propagating warm Kelvin waves (Barnett et al. 1989; Huang et al. 2001; Lengaigne et al. 2004; Chen et al. 2014, 2016). Therefore, negative geopotential height and cyclonic anomalies over the subtropical Pacific and associated westerly wind anomalies over the tropical west-central Pacific play an important role in connecting the Mar(0) ALI to the subsequent winter ENSO.

How are the cyclonic circulation anomaly and associated negative geopotential height anomalies generated over the subtropical North Pacific? To address this issue, SST anomalies in the North Pacific in Mar(0) are examined. Figure 5a displays the regression of SST anomalies in Mar(0) onto the normalized Mar(0) ALI. SST anomalies are weak and statistically insignificant in the tropical Pacific. This indicates that the formation of the cyclonic anomaly over the subtropical western North Pacific cannot be explained by tropical Pacific SST changes. Over the extratropical North Pacific, by contrast, a horseshoe-like SST anomaly pattern is observed, with significant negative SST anomalies along the west coast of North America and in the subtropical northeastern Pacific, accompanied by positive SST anomalies in the midlatitudes and off the east coast of East Asia (Fig. 5a). The generation of the SST anomalies in the North Pacific is primarily attributed to the surface heat flux changes induced by the Mar(0) ALI-related atmospheric circulation anomalies, as described below.

Figures 5b–f display the anomalies of surface net heat flux (NHF), latent heat flux (LHF), sensible heat flux (SHF), net shortwave radiation (SWR), and net longwave radiation (LWR) in Mar(0), respectively, regressed upon the standardized Mar(0) ALI. Values of surface heat flux anomalies are taken to be positive when they act to warm the ocean surface. A significant horseshoe-like NHF anomaly pattern is seen over the extratropical North Pacific (Fig. 5b), confirming the crucial role of the NHF changes in the formation of positive SST anomalies. Changes in NHF over the North Pacific are dominated by LHF anomalies (Figs. 6b,c). SHF and LWR have a positive contribution and SWR has a negative contribution to the NHF change (Figs. 5b,d). Significant increases in the LHF and SHF over midlatitudes of the North Pacific and subtropical central North Pacific (Figs. 5b,c) may be related to the anomalous easterly winds (Fig. 4a), which reduce the total wind speed there. Decrease in LHF and SHF along the east coast of North America and over the subtropical North Pacific (Figs. 5b,c) is likely attributed to anomalous northwesterly and northeasterly winds, which carry

FIG. 5. Regression of (a) SST ($^{\circ}\text{C}$), (b) surface net heat flux, (c) latent heat flux, (d) sensible heat flux, (e) shortwave radiation flux, and (f) longwave radiation flux anomalies in Mar(0) onto the normalized simultaneous Mar(0) ALI. Units for the surface heat fluxes in (b)–(f) are W m^{-2} . Shading denotes anomalies significance at the 95% confidence level.

colder and drier air from higher latitudes (Fig. 4a). A decrease (an increase) in SWR (LWR) over the subtropical west-central North Pacific (Figs. 5e,f) is probably due to an increase in total cloud cover (not shown) related to the anomalous cyclone. The above results suggest that simultaneous SST anomalies in the North Pacific are mainly induced by the atmospheric circulation anomalies related to the Mar(0) ALI.

The atmospheric circulation anomalies over the North Pacific in association with the Mar(0) ALI display a quasi-barotropic vertical structure (Fig. 4). This implies that atmospheric internal dynamics, usually associated with synoptic-scale eddy activity, may play a role in the formation of the circulation anomalies over the subtropical North Pacific (Lau 1988; Cai et al. 2007). Previous studies suggested that the interaction between the synoptic-scale eddy (also called the storm track) and

mean flow is a crucial internal source for the monthly mean atmospheric circulation anomalies (Lau 1988; Hartmann and Lo 1998; Limpasuvan and Hartmann 1999, 2000; Lorenz and Hartmann 2001, 2003; Cai et al. 2007). In the following, we analyze synoptic-scale eddy anomalies related to the Mar(0) ALI variation.

Figure 6 displays 300-hPa synoptic-scale eddy anomalies in Mar(0) regressed upon the normalized Mar(0) ALI. Synoptic-scale eddy activity is calculated as the root-mean-square of the 2–8-day bandpass-filtered geopotential height anomalies at a specific pressure level, following previous studies (Chang and Fu 2002; Lee et al. 2012; Chen et al. 2014). A significant increase (decrease) in the synoptic-scale eddy activity appears over the North Pacific along 50° – 60°N (30° – 40°N), corresponding to westerly (easterly) anomalies. As demonstrated by previous studies (Trenberth 1986; Lau

FIG. 6. Regression of 300-hPa storm track anomalies (m) in Mar(0) onto the normalized Mar(0) ALI. The stippled region indicates anomalies significant at the 95% confidence level.

1988; Cai et al. 2007), easterly (westerly) wind anomalies are accompanied by below (above) average synoptic-scale eddy activity as well as negative (positive) geopotential height tendency immediately to its south and positive (negative) geopotential height anomalies to its north. Thus, the marked decrease in the synoptic-scale eddy activity around 30°–40°N over the North Pacific contributes to the formation of the negative geopotential and cyclonic circulation anomalies over the subtropical North Pacific (Figs. 4 and 6). In addition, the significant increase of the synoptic-scale eddy activity around 50°–65°N (Fig. 6) is accompanied by a negative geopotential height tendency to its north and positive geopotential height tendency to its south. This contributes to the maintenance of anticyclonic circulation and positive geopotential height anomalies over the midlatitudes. It should be mentioned that easterly wind anomalies to the south of the midlatitude anticyclonic anomaly related to the enhanced synoptic-scale eddy activity can also impact synoptic-scale eddy activity around 30°–40°N. This implies that the positive synoptic-scale eddy anomalies along 50°–65°N over the North Pacific may also partly contribute to the formation of anomalous cyclonic circulation and negative geopotential height anomalies over the subtropical North Pacific. These results confirm a positive interaction between synoptic-scale eddy activity and atmospheric circulation anomalies over the North Pacific, as identified by previous studies (Lau 1988; Cai et al. 2007; Chen et al. 2014).

Figure 7 displays atmospheric circulation anomalies in Mar(0) regressed upon the Mar(0) North Pacific storm track intensity index (STI). The STI is defined as follows:

$$STI = ST_{[B]} - ST_{[A]}, \quad (4)$$

where $ST_{[B]}$ and $ST_{[A]}$ represent the regional mean storm track anomalies over 50°–65°N, 160°E–140°W and

FIG. 7. As in Fig. 4, but for anomalies regressed upon the normalized Mar(0) STI. Definition of the STI is provided in the text.

30°–40°N, 160°E–150°W, respectively. These regions are chosen according to the distribution of storm track anomalies in Fig. 6. Atmospheric circulation anomalies over the North Pacific related to Mar(0) STI are similar to those associated with the Mar(0) ALI, both displaying a barotropic dipole structure over the North Pacific. In particular, significant negative geopotential height and cyclonic circulation anomalies are apparent over the subtropical North Pacific (Fig. 7).

To further confirm the crucial role of the synoptic-scale eddy feedback in the formation of the cyclonic and negative geopotential height anomalies over the subtropical North Pacific, we display regressions of extended EP flux, divergence of the EP flux and geopotential height tendency anomalies at 300 hPa in

March onto the normalized Mar(0) STI during 1979–2016 in Fig. 8. Previous studies (Hendon and Hartmann 1985; Hoskins et al. 1983; Lau 1988; Cai et al. 2007) have demonstrated that the dynamical interaction between synoptic-scale eddies and the low-frequency mean flow could be described by the extended EP flux. Atmospheric anomalies attributed to the synoptic-scale eddy feedback could be qualitatively evaluated by the divergence of EP flux and quantitatively estimated by the eddy-induced geopotential height tendency (Trenberth 1986; Hendon and Hartmann 1985; Lau 1988). In particular, convergence (divergence) of EP flux is accompanied by anticyclonic (cyclonic) vorticity forcing to the north and cyclonic (anticyclonic) vorticity forcing to the south of the convergence (divergence) region. Strong and significant EP flux convergence anomalies are observed over the North Pacific around 30°–40°N, 150°E–130°W (Fig. 8a). These EP flux convergences over the North Pacific are accompanied by pronounced positive geopotential height tendencies to the north and negative geopotential height tendencies to the south with a weaker amplitude (Figs. 8a,b). In general, the spatial structure of the geopotential height tendency anomalies shown in Fig. 8b bears a resemblance to the patterns of geopotential height anomalies displayed in Figs. 4c and 7c. Hence, this diagnostic analysis confirms that the North Pacific synoptic-scale eddy activity anomalies and the related wave–mean flow interaction play an important role in the formation of the March ALI-related atmospheric circulation anomalies, including the cyclonic and negative geopotential height anomalies over the subtropical North Pacific.

We have also examined SST anomalies in the Pacific basin from the preceding winter to following winter in association with the Mar(0) STI (not shown). Results indicate that SST anomalies in the tropical Pacific in the preceding winter and simultaneous March are fairly weak and statistically insignificant. This suggests that the formation of the meridional dipolar anomaly pattern of storm track in Fig. 6 has a weak relation with preceding winter ENSO events.

b. Evolution of SST, winds, and atmospheric heating anomalies

Figure 9 displays the evolution of SST anomalies regressed upon the normalized ALI in Mar(0). Figure 10 shows the evolution of 850-hPa wind and precipitation anomalies. Here, precipitation anomalies are used to represent vertically integrated diabatic heating anomalies as in previous studies (Yu and Zwiers 2007; Chen et al. 2014, 2016). In general, positive (negative) precipitation anomalies correspond to enhanced (suppressed) atmospheric heating. In AM(0), a significant

FIG. 8. Regression of (a) extended EP flux (vectors; $\text{m}^2 \text{s}^{-2}$) and divergence of the EP flux (shading; m s^{-2}), and (b) geopotential height tendency (m day^{-1}) at 300 hPa in March onto the normalized Mar(0) STI during 1979–2016. Stippled regions in (a) and (b) indicate anomalies of EP flux divergence and geopotential height tendency significant at the 95% confidence level, respectively.

horseshoe-like SST anomaly pattern is present over the extratropical North Pacific, similar to that in Mar(0) (Figs. 5a and 9a). In addition, pronounced positive SST anomalies extend northeastward and southeastward from the tropical central Pacific to the subtropics (Fig. 9a), probably due to the southeasterly wind anomalies there (Fig. 4a). The anomalous southeasterly winds at the southern flank of the anomalous cyclone reduce the northeasterly winds, leading to the reduction of upward surface latent heat flux (Figs. 5b,c), contributing to local SST warming (Fig. 9a). Pronounced positive atmospheric heating anomalies (indicated by positive precipitation anomalies) appear over the subtropical North Pacific in AM(0) corresponding to the described SST warming (Figs. 9a and 10a). The subtropical atmospheric heating anomalies play a crucial role in maintaining the westerly wind anomalies over the

FIG. 9. Regression of SST anomalies ($^{\circ}\text{C}$) in (a) AM(0), (b) JJA(0), (c) SON(0), and (d) D(0)JF(1) onto the normalized Mar(0) ALI. Stippled regions indicate anomalies significant at the 95% confidence level.

tropical west-central Pacific via a Gill-type atmospheric response (Fig. 10a). These westerly wind anomalies subsequently trigger eastward-propagating warm Kelvin waves (as shown below), leading to equatorial central-eastern Pacific warming during the following summer and autumn (Figs. 9b,c). The tropical Pacific SST warming, atmospheric heating, and westerly wind anomalies maintain and develop into the following winter via a Bjerknes-like positive air–sea feedback process. Finally, El Niño-like SST warming and atmospheric circulation anomalies are induced in the following winter (Figs. 9d and 10d).

As indicated above, the appearance of tropical central-eastern Pacific SST warming during the following summer and autumn may partly be due to the eastward-propagating warm Kelvin waves induced by anomalous westerly winds over the tropical western Pacific. To confirm this, Fig. 11 displays the evolution of the tropical Kelvin wave forcing anomalies from spring to winter. As demonstrated in previous studies (Battisti 1988; Vimont et al. 2003), positive (negative) values of Kelvin wave forcing correspond to eastward-propagating warm (cold) Kelvin waves induced by surface zonal wind stress anomalies, which lead to SST warming (cooling). Pronounced positive anomalies of Kelvin wave forcing appear over the equatorial western Pacific in spring

(Fig. 11a), migrate eastward in summer (Fig. 11b), and propagate to the tropical central-eastern Pacific in autumn (Fig. 11c), with the amplitude being gradually intensified. This confirms that anomalous westerly wind anomalies over the tropical western Pacific contribute to SST warming in the tropical central-eastern Pacific via triggering eastward-propagating warm Kelvin waves.

5. Discussion

Previous studies have demonstrated that winter [ND(−1)JFM(0)] NPO can exert significant influences on the following winter ENSO via the seasonal footprinting mechanism (Vimont et al. 2001, 2003). The correlation coefficient between the winter NPO index and the following winter Niño-3.4 index reaches 0.37 during 1980–2016, significant at the 95% confidence level. Here, the NPO index is defined as the PC time series corresponding to the second EOF of SLP anomalies over 20° – 85°N , 120°E – 120°W , following Linkin and Nigam (2008). An immediate question is whether the impact of March ALI on the subsequent winter ENSO is independent of ND(−1)JFM(0) NPO. To answer this question, we have compared the Mar(0) ALI and ND(−1)JFM(0) NPO index. The correlation coefficient between Mar(0) ALI and ND(−1)JFM(0) NPO index is

FIG. 10. Regression of 850-hPa winds (vectors; m s^{-1}) and precipitation (shading; mm day^{-1}) anomalies in (a) AM(0), (b) JJA(0), (c) SON(0), and (d) D(0)JF(1) onto the normalized Mar(0) ALI. Stippled regions indicate precipitation anomalies significant at the 95% confidence level.

only 0.03, indicating that the variation of Mar(0) ALI is independent of ND(−1)JFM(0) NPO. In addition, we have recalculated the correlation coefficient between Mar(0) ALI and its subsequent winter Niño-3.4 index after removing the ND(−1)JFM(0) NPO signal by means of linear regression. It turns out that the correlation between Mar(0) ALI and following winter Niño-3.4 index is still significant ($R = 0.46$) at the 99% confidence level. This strongly suggests that the impact of Mar(0) ALI on the subsequent winter ENSO is independent of ND(−1)JFM(0) NPO.

Previous studies have demonstrated that spring (March–April average) AO can influence the occurrence of ENSO in the following winter via modulating wind anomalies over the tropical west-central Pacific (Nakamura et al. 2006, 2007; Chen et al. 2014). The correlation coefficient between spring AO index and the following winter Niño-3.4 index reaches 0.48 during 1979–2016, consistent with previous findings (Nakamura et al. 2006; Chen et al. 2014). Thus, another question is whether the connection of the Mar(0) ALI on the subsequent winter ENSO is attributed to the impact of spring AO. The correlation coefficient between Mar(0) ALI and spring AO is 0.26 during 1979–2016, which is below the 90%

confidence level. In addition, after removing the spring AO index from the Mar(0) ALI and its subsequent winter Niño-3.4 index, the correlation coefficient between the Mar(0) ALI and subsequent winter Niño-3.4 index is still significant ($R = 0.38$) at the 95% confidence level. In particular, the Mar(0) ALI-related cyclonic circulation and negative geopotential height anomalies are still apparent over the subtropical North Pacific (Fig. 12). These results indicate that the connection between Mar(0) ALI and its subsequent winter ENSO is independent of spring AO.

In addition, from Fig. 3, it seems that the in-phase variation of the Mar(0) ALI with the following winter Niño-3.4 index is stronger after than before the early 2000s. This indicates that the impact of the March ALI on the following winter ENSO may experience an enhancement around the early 2000s. By contrast, Chen et al. (2015) indicated that the influence of the spring AO on the following winter ENSO is weak after the mid-1990s (see their Figs. 2b and 2d). In particular, the correlation coefficient between the Mar(0) ALI and the following winter Niño-3.4 index is as high as 0.62 during 2001–16. By contrast, the correlation of the spring AO with following winter Niño-3.4 index is statistically

FIG. 11. Anomalies of Kelvin wave forcing (N m^{-1}) in (a) AM(0), (b) JJA(0), (c) SON(0), and (d) D(0)JF(1) regressed upon the normalized Mar(0) ALI. Anomalies that significantly different from zero at the 95% confidence level are indicated by red circles.

insignificant during 2001–16 ($R = 0.3$). This result can also be confirmed by Fig. 13, which displays regression patterns of winter SST anomalies onto the Mar(0) ALI and spring AO index after the early 2000s. A clear El Niño-like SST warming pattern appears in the tropical Pacific associated with a weakened ALI during preceding March (Fig. 13a). However, the impact of spring AO on the following winter ENSO appears weak after the early 2000s (Fig. 13b), which is consistent with Chen et al. (2015). These results suggest that the interdecadal variation of the ALI–ENSO relation is different from that of the AO–ENSO connection. The mechanism for the interdecadal enhancement of the March ALI–winter ENSO connection is not yet known and will be further investigated.

In addition, it should be mentioned that the March ALI-generated winter SST anomaly pattern in Fig. 13b is more like the central Pacific (CP) ENSO (Kao and Yu 2009; Kug et al. 2009). This suggests that the strengthened March ALI–winter ENSO connection after the early 2000s may partly contribute to the more frequent occurrence of CP ENSO events in recent decades. In particular, from Fig. 3b, most of the recent CP El Niño events (e.g., 2002/03, 2004/05, 2009/10, and 2015/16) (Yu and Kim 2013) are followed by a positive phase of the March ALI (i.e., weakened ALI). Several previous studies have indicated that springtime (FMA average) SST anomalies in the northern tropical Atlantic (NTA) have an impact on the CP ENSO events in the

following winter (Ham et al. 2013; Wang et al. 2017). In addition, it is found that connection of the springtime NTA SST anomalies with the following winter ENSO experienced a significant enhancement around the early 1980s (Chen and Wu 2017). To examine whether there exists a connection between the NTA SST and ALI, the correlation coefficient between the March ALI and the spring NTA SST index is calculated. Here, the spring (FMA average) NTA SST index is defined as area-mean SST anomalies over the region of 0° – 15°N , 90°W – 20°E following the method proposed by Ham et al. (2013). The correlation coefficient is about -0.24 during 1979–2016, which is statistically insignificant. Hence, it suggests that interannual variation of the March ALI is independent of the spring NTA SST.

The correlation coefficients among the ND(−1) JFM(0) NPO, MA(0) AO, and Mar(0) AL indices are generally weak, and statistically insignificant. This suggests that these three indices are relatively independent from each other. In the following, an empirical prediction model of the winter [ND(0)JF(1)] Niño-3.4 index is developed via a linear regression method using the ND(−1)JFM(0) NPO, MA(0) AO, and Mar(0) AL indices, which is shown as follows:

$$\begin{aligned} \text{ENSO}_{\text{WIN}+1}(t) = & \alpha \text{NPO}_{\text{WIN}-1}(t) + \beta \text{AO}_{\text{SPR}0}(t) \\ & + \gamma \text{AL}_{\text{MAR}0}(t) \end{aligned} \quad (5)$$

FIG. 12. As in Fig. 4, but the spring [MA(0) average] AO-related signals have been subtracted from Mar(0) ALI before constructing this figure.

where $ENSO_{WIN+1}(t)$, $NPO_{WIN-1}(t)$, $AO_{SPR0}(t)$, and $AL_{MAR0}(t)$ represent the ND(0)JF(1) Niño-3.4, ND(-1)JFM(0) NPO, MA(0) AO, and Mar(0) AL indices, respectively. From Fig. 14, using the three indices to hindcast the following winter Niño-3.4 index yields a correlation skill of 0.66, which is considerably larger than that using a single factor. This results suggest that preceding Mar(0) ALI variations can provide an additional source in the prediction of subsequent winter ENSO. Actually, studies found that winter NPO-like and spring AO-like atmospheric forcings do not always induce El Niño events (e.g., Alexander et al. 2010; Park et al. 2013; Chen et al. 2014, 2015). For example, Park et al. (2013) indicated that the occurrence rate of El

FIG. 13. Regression of winter SST anomalies onto the preceding normalized March ALI and spring AO index during 2001–16. Stippled regions indicate anomalies significant at the 95% confidence level.

Niño under the condition of a positive NPO in previous winter is about 41% during 1949–2009.

6. Conclusions

The present study reveals a close statistical connection between the interannual variation of March ALI and subsequent winter ENSO variability during 1979–2016. When March ALI is weaker (stronger) than normal, a significant El Niño (La Niña)-like SST warming (cooling) pattern occurs in the tropical central-eastern Pacific during the following winter. We examined the physical processes linking March ALI to the subsequent winter ENSO variability, which are summarized schematically in Fig. 15.

When March ALI is weaker than normal, pronounced anticyclonic circulation and positive geopotential height anomalies appear over the Aleutian region and significant cyclonic circulation and negative geopotential height anomalies are observed over the subtropical west-central North Pacific (Fig. 15). Anomalous cyclonic

FIG. 14. Normalized time series of the ND(−1)JFM(0) NPO (green line), MA(0) AO (blue line), Mar(0) AL (red line), and ND(0)JF(1) Niño-3.4 (bars) indices during 1980–2016. The black line indicates the hindcast ND(0)JF(1) Niño-3.4 index using three predictors [i.e., the ND(−1)JFM(0) NPO, MA(0) AO, and Mar(0) AL indices].

vortices and negative geopotential height anomalies over the subtropical North Pacific are likely attributable to the interaction between synoptic-scale eddies and mean flow and associated vorticity transportation. Easterly wind anomalies over the midlatitudes of the North Pacific are accompanied by weakened synoptic-scale eddy activity that induces a negative geopotential height tendency to its south and a positive geopotential height tendency to its north (Fig. 15). Hence, a cyclonic circulation anomaly and negative geopotential height anomalies are induced over the subtropical North Pacific (Fig. 15).

Southeasterly wind anomalies at the western flank of the subtropical cyclonic anomaly reduce the trade winds and lead to positive SST anomalies in the subtropical North Pacific in spring via a reduction of upward surface latent heat flux. Significant atmospheric heating associated with the SST warming over the subtropical North Pacific maintains the westerly wind anomalies over the tropical western Pacific via a Gill-type atmospheric response. The anomalous westerly winds trigger eastward-propagating warm Kelvin waves, leading to large SST warming in the tropical central-eastern Pacific (Fig. 15). The associated atmospheric heating, atmospheric circulation, and SST warming over the tropics sustain and develop into the following autumn and winter via a Bjerknes-like positive air–sea feedback (Fig. 15). Finally, El Niño-like SST warming and associated atmospheric circulation anomalies are generated in the following winter (Fig. 15).

The present study indicates that the wintertime NPO, spring AO, and March ALI are three independent factors for the following winter ENSO variability. The combination of winter NPO, March ALI, and spring

FIG. 15. A mechanism diagram showing the processes linking the March ALI to the subsequent winter ENSO. AC (C) indicates anticyclonic (cyclonic) anomalies. SWNP, TWP, and TCEP denote subtropical western North Pacific, tropical western Pacific, and tropical central-eastern Pacific, respectively.

AO can improve the prediction of winter ENSO. The physical processes for the joint impacts of winter NPO, March ALI, and spring AO on ENSO will be investigated in a separate study.

Acknowledgments. We thank the anonymous reviewers for their constructive suggestions and comments, which help to improve the paper. This study is supported by the National Natural Science Foundation of China grants (41605050, 41530425, and 41775080) and the Young Elite Scientists Sponsorship Program by the China Association for Science and Technology (2016QNRC001). The NCEP–DOE reanalysis data are obtained from <https://www.esrl.noaa.gov/psd/data/gridded/data.ncep.reanalysis2.html>. The ERSSTv3b SST data are obtained from <https://www.esrl.noaa.gov/psd>. The AO index is derived from <http://www.cpc.ncep.noaa.gov>.

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., I. Bladé, M. Newman, J. R. Lanzante, N.-C. Lau, and J. D. Scott, 2002: The atmospheric bridge: The influence of

- ENSO teleconnections on air–sea interaction over the global oceans. *J. Climate*, **15**, 2205–2231, [https://doi.org/10.1175/1520-0442\(2002\)015<2205:TABTIO>2.0.CO;2](https://doi.org/10.1175/1520-0442(2002)015<2205:TABTIO>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., 2007: On the joint role of subtropical atmospheric variability and equatorial subsurface heat content anomalies in initiating the onset of ENSO events. *J. Climate*, **20**, 1593–1599, <https://doi.org/10.1175/JCLI4075.1>.
- Barnett, T., L. Dümenil, U. Schlese, E. Roeckner, and M. Latif, 1989: The effect of Eurasian snow cover on regional and global climate variations. *J. Atmos. Sci.*, **46**, 661–686, [https://doi.org/10.1175/1520-0469\(1989\)046<0661:TEOESC>2.0.CO;2](https://doi.org/10.1175/1520-0469(1989)046<0661:TEOESC>2.0.CO;2).
- Battisti, D. S., 1988: Dynamics and thermodynamics of a warming event in a coupled tropical atmosphere–ocean model. *J. Atmos. Sci.*, **45**, 2889–2919, [https://doi.org/10.1175/1520-0469\(1988\)045<2889:DATOAW>2.0.CO;2](https://doi.org/10.1175/1520-0469(1988)045<2889:DATOAW>2.0.CO;2).
- Bjerknes, J., 1969: Atmospheric teleconnections from the equatorial Pacific. *Mon. Wea. Rev.*, **97**, 163–172, [https://doi.org/10.1175/1520-0493\(1969\)097<0163:ATFTEP>2.3.CO;2](https://doi.org/10.1175/1520-0493(1969)097<0163:ATFTEP>2.3.CO;2).
- Cai, M., S. Yang, H. van den Dool, and V. Kousky, 2007: Dynamical implications of the orientation of atmospheric eddies: A local energetics perspective. *Tellus*, **59A**, 127–140, <https://doi.org/10.1111/j.1600-0870.2006.00213.x>.
- Chang, E. K. M., and Y. F. Fu, 2002: Interdecadal variations in Northern Hemisphere winter storm track intensity. *J. Climate*, **15**, 642–658, [https://doi.org/10.1175/1520-0442\(2002\)015<0642:IVINHW>2.0.CO;2](https://doi.org/10.1175/1520-0442(2002)015<0642:IVINHW>2.0.CO;2).
- Chen, S. F., and R. Wu, 2017: An enhanced influence of sea surface temperature in the tropical northern Atlantic on the following winter ENSO since the early 1980s. *Atmos. Oceanic Sci. Lett.*, **10**, 175–182, <https://doi.org/10.1080/16742834.2016.1259542>.
- , and —, 2018: Impacts of winter NPO on subsequent winter ENSO: Sensitivity to the definition of NPO index. *Climate Dyn.*, **50**, 375–389, <https://doi.org/10.1007/s00382-017-3615-z>.
- , and L. Song, 2019: The leading interannual variability modes of winter surface air temperature over Southeast Asia. *Climate Dyn.*, **52**, 4715–4734, <https://doi.org/10.1007/s00382-018-4406-x>.
- , W. Chen, B. Yu, and H.-F. Graf, 2013: Modulation of the seasonal footprinting mechanism by the boreal spring Arctic Oscillation. *Geophys. Res. Lett.*, **40**, 6384–6389, <https://doi.org/10.1002/2013GL058628>.
- , B. Yu, and W. Chen, 2014: An analysis on the physical process of the influence of AO on ENSO. *Climate Dyn.*, **42**, 973–989, <https://doi.org/10.1007/s00382-012-1654-z>.
- , —, and —, 2015: An interdecadal change in the influence of the spring Arctic Oscillation on the subsequent ENSO around the early 1970s. *Climate Dyn.*, **44**, 1109–1126, <https://doi.org/10.1007/s00382-014-2152-2>.
- , R. Wu, W. Chen, B. Yu, and X. Cao, 2016: Genesis of westerly wind bursts over the equatorial western Pacific during the onset of the strong 2015–2016 El Niño. *Atmos. Sci. Lett.*, **17**, 384–391, <https://doi.org/10.1002/asl.669>.
- , W. Chen, and B. Yu, 2017: The influence of boreal spring Arctic Oscillation on the subsequent winter ENSO in CMIP5 models. *Climate Dyn.*, **48**, 2949–2965, <https://doi.org/10.1007/s00382-016-3243-z>.
- , —, and —, 2018: Modulation of the relationship between spring AO and the subsequent winter ENSO by the preceding November AO. *Sci. Rep.*, **8**, 6943, <https://doi.org/10.1038/s41598-018-25303-0>.
- Chen, W., 2002: Impacts of El Niño and La Niña on the cycle of the East Asian winter and summer monsoon (in Chinese). *Chin. J. Atmos. Sci.*, **26**, 595–610.
- Choi, J., and Y. Cha, 2017: Anomalous variation in summer tropical cyclone activity by preceding winter Aleutian low oscillation. *Atmos. Sci. Lett.*, **18**, 268–275, <https://doi.org/10.1002/asl.752>.
- Deser, C., 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>.
- Duchon, C. E., 1979: Lanczos filtering in one and two dimensions. *J. Appl. Meteor.*, **18**, 1016–1022, [https://doi.org/10.1175/1520-0450\(1979\)018<1016:LFIOAT>2.0.CO;2](https://doi.org/10.1175/1520-0450(1979)018<1016:LFIOAT>2.0.CO;2).
- Gershunov, A., and T. P. Barnett, 1998: Interdecadal modulation of ENSO teleconnections. *Bull. Amer. Meteor. Soc.*, **79**, 2715–2725, [https://doi.org/10.1175/1520-0477\(1998\)079<2715:IMOET>2.0.CO;2](https://doi.org/10.1175/1520-0477(1998)079<2715:IMOET>2.0.CO;2).
- Gill, A. E., 1980: Some simple solutions for heat-induced tropical circulation. *Quart. J. Roy. Meteor. Soc.*, **106**, 447–462, <https://doi.org/10.1002/qj.49710644905>.
- Ham, Y.-G., J.-S. Kug, J.-Y. Park, and F.-F. Jin, 2013: Sea surface temperature in the north tropical Atlantic as a trigger for El Niño/Southern Oscillation events. *Nat. Geosci.*, **6**, 112–116, <https://doi.org/10.1038/ngeo1686>.
- Hartmann, D. L., and F. Lo, 1998: Wave-driven zonal flow vacillation in the Southern Hemisphere. *J. Atmos. Sci.*, **55**, 1303–1315, [https://doi.org/10.1175/1520-0469\(1998\)055<1303:WDFVI>2.0.CO;2](https://doi.org/10.1175/1520-0469(1998)055<1303:WDFVI>2.0.CO;2).
- Hendon, H. H., and D. L. Hartmann, 1985: Variability in a nonlinear model of the atmosphere with zonally symmetric forcing. *J. Atmos. Sci.*, **42**, 2783–2797, [https://doi.org/10.1175/1520-0469\(1985\)042<2783:VIANMO>2.0.CO;2](https://doi.org/10.1175/1520-0469(1985)042<2783:VIANMO>2.0.CO;2).
- Horel, J. D., and J. M. Wallace, 1981: Planetary-scale atmospheric phenomena associated with the Southern Oscillation. *Mon. Wea. Rev.*, **109**, 813–829, [https://doi.org/10.1175/1520-0493\(1981\)109<0813:PSAPAW>2.0.CO;2](https://doi.org/10.1175/1520-0493(1981)109<0813:PSAPAW>2.0.CO;2).
- Hoskins, B. J., I. N. James, and G. H. White, 1983: The shape, propagation and mean-flow interaction of large-scale weather systems. *J. Atmos. Sci.*, **40**, 1595–1612, [https://doi.org/10.1175/1520-0469\(1983\)040<1595:TSPAMF>2.0.CO;2](https://doi.org/10.1175/1520-0469(1983)040<1595:TSPAMF>2.0.CO;2).
- Huang, R., R. Zhang, and B. Yan, 2001: Dynamical effect of the zonal wind anomalies over the tropical western Pacific on ENSO cycles. *Sci. China*, **44D**, 1089–1098, <https://doi.org/10.1007/BF02906865>.
- Jin, F.-F., 1997: An equatorial ocean recharge paradigm for ENSO. Part I: Conceptual model. *J. Atmos. Sci.*, **54**, 811–829, [https://doi.org/10.1175/1520-0469\(1997\)054<0811:AEORPF>2.0.CO;2](https://doi.org/10.1175/1520-0469(1997)054<0811:AEORPF>2.0.CO;2).
- Kanamitsu, M., W. Ebisuzaki, J. Woollen, S.-K. Yang, J. J. Hnilo, M. Fiorino, and G. L. Potter, 2002: NCEP–DOE AMIP-II Reanalysis (R-2). *Bull. Amer. Meteor. Soc.*, **83**, 1631–1643, <https://doi.org/10.1175/BAMS-83-11-1631>.
- Kao, H.-Y., and J.-Y. Yu, 2009: Contrasting eastern-Pacific and central-Pacific types of ENSO. *J. Climate*, **22**, 615–632, <https://doi.org/10.1175/2008JCLI2309.1>.
- Kug, J.-S., F.-F. Jin, and S.-I. An, 2009: Two types of El Niño events: Cold tongue El Niño and warm pool El Niño. *J. Climate*, **22**, 1499–1515, <https://doi.org/10.1175/2008JCLI2624.1>.
- Lai, A. W., M. Herzog, and H. F. Graf, 2015: Two key parameters for the El Niño continuum: Zonal wind anomalies and Western Pacific subsurface potential temperature. *Climate Dyn.*, **45**, 3461–3480, <https://doi.org/10.1007/s00382-015-2550-0>.

- Lau, N.-C., 1988: Variability of the observed midlatitude storm tracks in relation to low-frequency changes in the circulation pattern. *J. Atmos. Sci.*, **45**, 2718–2743, [https://doi.org/10.1175/1520-0469\(1988\)045<2718:VOTOMS>2.0.CO;2](https://doi.org/10.1175/1520-0469(1988)045<2718:VOTOMS>2.0.CO;2).
- , and E. O. Holopainen, 1984: Transient eddy forcing of the time-mean flow as identified by geopotential tendencies. *J. Atmos. Sci.*, **41**, 313–328, [https://doi.org/10.1175/1520-0469\(1984\)041<0313:TEFOTT>2.0.CO;2](https://doi.org/10.1175/1520-0469(1984)041<0313:TEFOTT>2.0.CO;2).
- , and M. J. Nath, 1996: The role of the “atmospheric bridge” in linking tropical Pacific ENSO events to extratropical SST anomalies. *J. Climate*, **9**, 2036–2057, [https://doi.org/10.1175/1520-0442\(1996\)009<2036:TROBTI>2.0.CO;2](https://doi.org/10.1175/1520-0442(1996)009<2036:TROBTI>2.0.CO;2).
- Lee, S.-S., J.-Y. Lee, B. Wang, K.-J. Ha, K.-Y. Heo, F.-F. Jin, D. M. Straus, and J. Shukla, 2012: Interdecadal changes in the storm track activity over the North Pacific and North Atlantic. *Climate Dyn.*, **39**, 313–327, <https://doi.org/10.1007/s00382-011-1188-9>.
- Lengaigne, M., E. Guilyardi, J. P. Boulanger, C. Menkes, P. Delecluse, P. Inness, J. Cole, and J. Slingo, 2004: Triggering of El Niño by westerly wind events in a coupled general circulation model. *Climate Dyn.*, **23**, 601–620, <https://doi.org/10.1007/s00382-004-0457-2>.
- Li, C., 1990: Interaction between anomalous winter monsoon in East Asia and El Niño events. *Adv. Atmos. Sci.*, **7**, 36–46, <https://doi.org/10.1007/BF02919166>.
- Lian, T., D. Chen, Y. Tang, and Q. Wu, 2014: Effects of westerly wind bursts on El Niño: A new perspective. *Geophys. Res. Lett.*, **41**, 3522–3527, <https://doi.org/10.1002/2014GL059989>.
- Limpasuvan, V., and D. L. Hartmann, 1999: Eddies and the annular modes of climate variability. *Geophys. Res. Lett.*, **26**, 3133–3136, <https://doi.org/10.1029/1999GL010478>.
- , and —, 2000: Wave-maintained annular modes of climate variability. *J. Climate*, **13**, 4414–4429, [https://doi.org/10.1175/1520-0442\(2000\)013<4414:WMAMOC>2.0.CO;2](https://doi.org/10.1175/1520-0442(2000)013<4414:WMAMOC>2.0.CO;2).
- Linkin, M. E., and S. Nigam, 2008: The North Pacific oscillation–west Pacific teleconnection pattern: Mature-phase structure and winter impacts. *J. Climate*, **21**, 1979–1997, <https://doi.org/10.1175/2007JCLI2048.1>.
- Lorenz, D. J., and D. L. Hartmann, 2001: Eddy–zonal flow feedback in the Southern Hemisphere. *J. Atmos. Sci.*, **58**, 3312–3327, [https://doi.org/10.1175/1520-0469\(2001\)058<3312:EZFIFT>2.0.CO;2](https://doi.org/10.1175/1520-0469(2001)058<3312:EZFIFT>2.0.CO;2).
- , and —, 2003: Eddy–zonal flow feedback in the Northern Hemisphere winter. *J. Climate*, **16**, 1212–1227, [https://doi.org/10.1175/1520-0442\(2003\)16<1212:EFFITN>2.0.CO;2](https://doi.org/10.1175/1520-0442(2003)16<1212:EFFITN>2.0.CO;2).
- Luo, J. J., S. Masson, S. Behera, and T. Yamagata, 2008: Extended ENSO predictions using a fully coupled ocean–atmosphere model. *J. Climate*, **21**, 84–93, <https://doi.org/10.1175/2007JCLI1412.1>.
- , R. Zhang, S. K. Behera, Y. Masumoto, F. Jin, R. Lukas, and T. Yamagata, 2010: Interaction between El Niño and extreme Indian Ocean dipole. *J. Climate*, **23**, 726–742, <https://doi.org/10.1175/2009JCLI13104.1>.
- McPhaden, M. J., 1999: Genesis and evolution of the 1997–98 El Niño. *Science*, **283**, 950–954, <https://doi.org/10.1126/science.283.5404.950>.
- , 2003: Tropical Pacific Ocean heat content variations and ENSO persistence barriers. *Geophys. Res. Lett.*, **30**, 1480, <https://doi.org/10.1029/2003GL016872>.
- , S. E. Zebiak, and M. H. Glantz, 2006: ENSO as an integrating concept in Earth science. *Science*, **314**, 1740–1745, <https://doi.org/10.1126/science.1132588>.
- Mo, K. C., 2010: Interdecadal modulation of the impact of ENSO on precipitation and temperature over the United States. *J. Climate*, **23**, 3639–3656, <https://doi.org/10.1175/2010JCLI3553.1>.
- Nakamura, T., Y. Tachibana, M. Honda, and S. Yamane, 2006: Influence of the Northern Hemisphere annular mode on ENSO by modulating westerly wind bursts. *Geophys. Res. Lett.*, **33**, L07709, <https://doi.org/10.1029/2005GL025432>.
- , —, and H. Shimoda, 2007: Importance of cold and dry surges in substantiating the NAM and ENSO relationship. *Geophys. Res. Lett.*, **34**, L22703, <https://doi.org/10.1029/2007GL031220>.
- North, G. R., T. L. Bell, R. F. Cahalan, and F. J. Moeng, 1982a: Sampling errors in the estimation of empirical orthogonal functions. *Mon. Wea. Rev.*, **110**, 699–706, [https://doi.org/10.1175/1520-0493\(1982\)110<0699:SEITEO>2.0.CO;2](https://doi.org/10.1175/1520-0493(1982)110<0699:SEITEO>2.0.CO;2).
- , F. J. Moeng, T. L. Bell, and R. F. Cahalan, 1982b: The latitude dependence of the variance of zonally averaged quantities. *Mon. Wea. Rev.*, **110**, 319–326, [https://doi.org/10.1175/1520-0493\(1982\)110<0319:TLDOTV>2.0.CO;2](https://doi.org/10.1175/1520-0493(1982)110<0319:TLDOTV>2.0.CO;2).
- Park, J.-Y., S.-W. Yeh, J.-S. Kug, and J. Yoon, 2013: Favorable connections between seasonal footprinting mechanism and El Niño. *Climate Dyn.*, **40**, 1169–1181, <https://doi.org/10.1007/s00382-012-1477-y>.
- Rogers, J. C., 1981: The North Pacific oscillation. *J. Climatol.*, **1**, 39–57, <https://doi.org/10.1002/joc.3370010106>.
- Schopf, P. S., and M. J. Suarez, 1988: Vacillations in a coupled ocean–atmosphere model. *J. Atmos. Sci.*, **45**, 549–566, [https://doi.org/10.1175/1520-0469\(1988\)045<0549:VIACOM>2.0.CO;2](https://doi.org/10.1175/1520-0469(1988)045<0549:VIACOM>2.0.CO;2).
- Smith, T. M., R. W. Reynolds, T. C. Peterson, and J. Lawrimore, 2008: Improvements to NOAA’s historical merged land–ocean surface temperature analysis (1880–2006). *J. Climate*, **21**, 2283–2296, <https://doi.org/10.1175/2007JCLI2100.1>.
- Song, L.-Y., and W.-S. Duan, 2015: Interannual relationship between the winter Aleutian low and rainfall in the following summer in South China. *Atmos. Oceanic Sci. Lett.*, **8**, 271–276, <https://doi.org/10.3878/AOSL20150021>.
- , S. F. Chen, W. Chen, and X. Chen, 2017: Distinct impacts of two types of La Niña events on Australian summer rainfall. *Int. J. Climatol.*, **37**, 2532–2544, <https://doi.org/10.1002/joc.4863>.
- Straus, D. M., and J. Shukla, 2000: Distinguishing between the SST-forced variability and internal variability in mid latitudes: Analysis of observations and GCM simulations. *Quart. J. Roy. Meteor. Soc.*, **126**, 2323–2350, <https://doi.org/10.1002/qj.49712656716>.
- Thompson, D. W., and J. M. Wallace, 1998: The Arctic Oscillation signature in the wintertime geopotential height and temperature fields. *Geophys. Res. Lett.*, **25**, 1297–1300, <https://doi.org/10.1029/98GL00950>.
- Trenberth, K. E., 1986: An assessment of the impact of transient eddies on the zonal flow during a blocking episode using localized Eliassen–Palm flux diagnostics. *J. Atmos. Sci.*, **43**, 2070–2087, [https://doi.org/10.1175/1520-0469\(1986\)043<2070:AAOTIO>2.0.CO;2](https://doi.org/10.1175/1520-0469(1986)043<2070:AAOTIO>2.0.CO;2).
- , 1997: The definition of El Niño. *Bull. Amer. Meteor. Soc.*, **78**, 2771–2777, [https://doi.org/10.1175/1520-0477\(1997\)078<2771:TDOENO>2.0.CO;2](https://doi.org/10.1175/1520-0477(1997)078<2771:TDOENO>2.0.CO;2).
- Vimont, D. J., 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>.
- , J. M. Wallace, and D. S. Battisti, 2003: The seasonal footprinting mechanism in the Pacific: Implications for ENSO. *J. Climate*, **16**, 2668–2675, [https://doi.org/10.1175/1520-0442\(2003\)016<2668:TSMFIT>2.0.CO;2](https://doi.org/10.1175/1520-0442(2003)016<2668:TSMFIT>2.0.CO;2).

- Walker, G. T., and E. Bliss, 1932: World weather V. *Mem. Roy. Meteor. Soc.*, **4**, 53–84.
- Wallace, J. M., and D. S. Gutzler, 1981: Teleconnections in the geopotential height field during the Northern Hemisphere winter. *Mon. Wea. Rev.*, **109**, 784–812, [https://doi.org/10.1175/1520-0493\(1981\)109<0784:TITGHF>2.0.CO;2](https://doi.org/10.1175/1520-0493(1981)109<0784:TITGHF>2.0.CO;2).
- Wang, B., R. Wu, and X. Fu, 2000: Pacific–East Asian teleconnection: How does ENSO affect East Asian climate? *J. Climate*, **13**, 1517–1536, [https://doi.org/10.1175/1520-0442\(2000\)013<1517:PEATHD>2.0.CO;2](https://doi.org/10.1175/1520-0442(2000)013<1517:PEATHD>2.0.CO;2).
- Wang, C., R. H. Weisberg, and J. Virmani, 1999: Western Pacific interannual variability associated with the El Niño–Southern Oscillation. *J. Geophys. Res.*, **104**, 5131–5149, <https://doi.org/10.1029/1998JC900090>.
- Wang, L., J.-Y. Yu, and H. Paek, 2017: Enhanced biennial variability in the Pacific due to Atlantic capacitor effect. *Nat. Commun.*, **8**, 14887, <https://doi.org/10.1038/ncomms14887>.
- Wang, S., M. L’Heureux, and H. H. Chia, 2012: ENSO prediction one year in advance using western North Pacific sea surface temperatures. *Geophys. Res. Lett.*, **39**, L05702, <https://doi.org/10.1029/2012GL050909>.
- Wang, X., C. Wang, W. Zhou, D. Wang, and J. Song, 2011: Teleconnected influence of North Atlantic sea surface temperature on the El Niño onset. *Climate Dyn.*, **37**, 663–676, <https://doi.org/10.1007/s00382-010-0833-z>.
- Webster, P. J., and S. Yang, 1992: Monsoon and ENSO: Selectively interactive systems. *Quart. J. Roy. Meteor. Soc.*, **118**, 877–926, <https://doi.org/10.1002/qj.49711850705>.
- Wu, R., Z.-Z. Hu, and B. P. Kirtman, 2003: Evolution of ENSO-related rainfall anomalies in East Asia. *J. Climate*, **16**, 3742–3758, [https://doi.org/10.1175/1520-0442\(2003\)016<3742:EOERAI>2.0.CO;2](https://doi.org/10.1175/1520-0442(2003)016<3742:EOERAI>2.0.CO;2).
- Yu, B., and F. Zwiers, 2007: The impact of combined ENSO and PDO on the PNA climate: A 1,000-year climate modeling study. *Climate Dyn.*, **29**, 837–851, <https://doi.org/10.1007/s00382-007-0267-4>.
- 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>.
- , and —, 2013: Identifying the types of major El Niño events since 1870. *Int. J. Climatol.*, **33**, 2105–2112, <https://doi.org/10.1002/joc.3575>.
- , 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>.
- 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).
- Zhu, J., A. Kumar, B. Huang, M. Balmaseda, Z.-Z. Hu, L. Marx, and J. L. Kinter III, 2016: The role of off-equatorial surface temperature anomalies in the 2014 El Niño prediction. *Sci. Rep.*, **6**, 19677, <https://doi.org/10.1038/srep19677>.
- Zhu, Y. L., and H. J. Wang, 2010: The relationship between the Aleutian Low and the Australian summer monsoon at interannual time scales. *Adv. Atmos. Sci.*, **27**, 177–184, <https://doi.org/10.1007/s00376-009-8144-1>.