Combined Impacts of PDO and Two Types of La Niña on Climate Anomalies in Europe

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ABSTRACT

Combined impacts of the Pacific decadal oscillation (PDO) and two types of La Niña on climate anomalies in Europe are studied. Particularly, the conjunction of the negative PDO phase and two different types of La Niña events favors strong and significant North Atlantic Oscillation (NAO) pattern anomalies with opposite polarity. For the central Pacific (CP) La Niña, a clear positive NAO signal can be detected, which is accompanied by positive surface air temperature (SAT) anomaly and a dipolar structure of precipitation anomalies in Europe. In addition, a typical negative Pacific–North America (PNA) teleconnection pattern forms, including a high pressure anomaly over the southeastern United States, which may contribute to the development and maintenance of the NAO anomaly by strengthening the baroclinicity and the local eddy–mean flow interaction. However, for the eastern Pacific (EP) La Niña, a zonal wave train in the high latitudes can be observed, which is quite different from the typical PNA structure. Here, an anomalous anticyclone over southern Greenland supports a negative NAO pattern through the local eddy–mean flow interaction and the associated vorticity advection. Hence, reversed SAT and precipitation anomalies occur over Europe. Further analyses indicate that the wave trains emanating from the North Pacific and the synoptic eddy–mean flow interaction play essential roles in forming the anomalous NAO phases. The different wave trains for the CP and EP La Niña events may be attributed to the differences in the location and intensity of anomalous convection induced by different types of SST anomaly as well as by the corresponding background westerly wind anomalies in the upper troposphere.

1. Introduction

El Niño–Southern Oscillation (ENSO), with a quasi-periodic fluctuation between warm (El Niño) and cold (La Niña) sea surface temperature (SST) anomalies over the equatorial central-eastern Pacific, is the most important interannual coupled ocean–atmosphere signal in the tropical Pacific. It not only directly influences the climates over the tropical Pacific, but also indirectly induces temperature and precipitation anomalies over the extratropics through teleconnections such as the Pacific–North America (PNA) teleconnection (Wallace and Gutzler 1981; Barnston and Livezey 1987).

In recent years, central Pacific (CP) El Niño, with maximum SST warming over the equatorial central Pacific accompanied by prominent changes in the location of the tropical anomalous precipitation and convective activity, has received extensive attention (Larkin and Harrison 2005; Ashok et al. 2007; Kao and Yu 2009; Kug et al. 2009; Yeh et al. 2009). This “new” type of El Niño occurs more frequently after the 1990s (Lee and McPhaden 2010; Pascolini-Campbell et al. 2015) and generates different climate responses in the Pacific rim compared to the conventional type, named the eastern Pacific (EP) El Niño. This includes temperature and precipitation anomalies over the Asia–Australia region in different seasons (Wang and Hendon 2007; Weng et al. 2007, 2009; Chen et al. 2009; Taschetto and England 2009; Feng et al. 2010, 2011; Feng and Li 2011; Zhang et al. 2011; Yuan and Yang 2012; Karori et al. 2013; Zhang et al. 2013, 2014; Jia et al. 2016), the PNA teleconnection related to North American climate anomalies during boreal winter and summer (Weng et al. 2007, 2009; Yu and Zou 2013), and the frequency of both Atlantic hurricanes and northwestern Pacific tropical cyclones (Kim et al. 2009; Chen and Tam 2010; Wang et al.
2013). Graf and Zanchettin (2012) also suggested that CP El Niño can induce significant cooling over the North Atlantic/European (NA/E) sector associated with a negative phase of the North Atlantic Oscillation (NAO) in winter through a “subtropical bridge” mechanism, while EP El Niño results in weak warming with insignificant impact on the NAO pattern. Through a sensitivity study with an atmospheric general circulation model, Frauen et al. (2014) further confirmed the different responses of sea level pressure (SLP) and tropical rainfall anomalies to two types of ENSO.

These previous studies mainly investigated the different impacts of CP and EP El Niño. A few investigations also discussed the cold phase of ENSO and whether La Niña events can also be classified into different types. Some studies argued that since the tropical SST and precipitation anomalies patterns are seemingly similar for all La Niñas there is no possibility of separating different types (Kug and Ham 2011; Ren and Jin 2011). On the contrary, Cai and Cowan (2009), Yuan and Yan (2013), and Shinoda et al. (2011) suggest that CP La Niña can be distinguished from EP La Niña and their impact on climate is different in the tropical Pacific and its neighborhood. On the basis of the normalized Niño-3 and Niño-4 indices and the SST anomaly distribution during the mature phase, Yuan and Yan (2013) divided La Niña into CP and EP types and revealed that the tropical atmosphere responds to CP La Niña quite differently than to EP La Niña, with the former leading to less precipitation over the equatorial central Pacific while the latter is associated with more rainfall over the eastern Indian Ocean and western Pacific. CP La Niña seasons show a more westward location of the large-scale sinking motion in the troposphere with stronger low-level divergent winds and high-level convergent wind anomalies over the central-eastern Pacific than EP La Niña. Cai and Cowan (2009) also reported increased autumn rainfall over northwestern Australia extending to the northern Murray-Darling basin during CP La Niña, rather than over eastern Australia during EP La Niña. Approximately opposite climate anomalies are also discovered over the NA/E sector for the different types of La Niña. Zhang et al. (2015) identified two types of La Niña based on the winter Niño-3.4 index and the location of the strongest negative SST anomaly. They indicated that the response to CP La Niña exhibits a positive NAO pattern with a strengthened North Atlantic jet, leading to a warmer and wetter winter over western Europe. During EP La Niña the anomalies have opposite sign. Although some numerical experiments were performed supporting the observed differences between CP and EP La Niña, the fundamental dynamical mechanism of La Niña–NAO teleconnection is still ambiguous. It was suggested that the PNA-like wave train resulting from convective heating anomalies over the equatorial Pacific plays an important role in connecting the tropical SST anomalies in the Pacific and NAO-like atmospheric anomalies (Honda et al. 2001; Honda and Nakamura 2001; Pinto et al. 2011). For example, a positive geopotential height anomaly in the upper troposphere over the southeastern United States associated with negative NAO not only results in a northward shift of the North Atlantic storm track (Honda et al. 2001; Honda and Nakamura 2001), but also induces a southwest–northeast-oriented enhancement of baroclinicity in the upstream of Newfoundland extending toward the North Atlantic, providing favorable conditions for the growth of eddies at the entrance of the North Atlantic storm track (Pinto et al. 2011). The increased baroclinicity is generated by the modulation of cold air advection from the north and warm air advection from the south connected with the PNA centers over western Canada and southeastern United States. Then, through the persistent positive feedback from transient eddies along the storm track, a positive NAO-like pattern is established, to keep growing until it gradually reaches the mature phase by late winter (Li and Lau 2012a,b). However, during EP La Niña, a negative NAO signal accompanied by a negative value of PNA is observed in late winter, which is inconsistent with the teleconnection between PNA and NAO as pointed out by previous studies (Honda et al. 2001; Honda and Nakamura 2001; Pinto et al. 2011). Here, a hypothesis that different Rossby wave trains may be the cause for the opposite climate responses over the NA/E sector to the two types of La Niña is put forward and will be investigated in the study.

Earlier studies also showed that the climate signals of ENSO over the NA/E sector are quite variable, which is possibly caused by factors like internal variability of the extratropical circulation (Kumar and Hoerling 1998), tropical volcanic eruptions (Brönnimann et al. 2006), the SST variability in tropical Atlantic (Mathieu et al. 2004; Sung et al. 2013; Ham et al. 2014), the multidecadal changes in the ocean mean state (López-Parages et al. 2015, 2016), different regimes of stratospheric polar vortex circulation (Graf et al. 2014), and the complexity of ENSO itself (Greatbatch et al. 2004). Among those various influencing factors, the Pacific decadal oscillation (PDO), defined as the leading EOF mode of SST anomalies in the North Pacific Ocean, poleward of 20°N (Zhang et al. 1997; Mantua et al. 1997), is a key factor possibly modulating the ENSO-related circulation, temperature, and rainfall anomalies. For example, the relation of ENSO to
both East Asian winter monsoon (EAWM) and East Asian summer monsoon (EASM) were shown to be modulated by the PDO (Chan and Zhou 2005; Wang et al. 2008; Chen et al. 2013; Feng et al. 2014). The ENSO–EAWM relationship is robust only during the negative PDO phase, owing to the reinforced anomalous Philippine anticyclone (Wang et al. 2008). The anomalous summer monsoon precipitation in China exhibits a tripolar (dipolar) distribution during the decaying phase of El Niño when the PDO is in its positive (negative) phase (Feng et al. 2014). Yoon and Yeh (2010) found that the relationship between El Niño and the northeast Asian summer monsoon (NEASM) is intensified (weakened) when El Niño appearing during the positive (negative) PDO. Yu and Zwiers (2007) proposed that the in-phase combination of ENSO and PDO would enhance the PNA-like wave structure and climate impacts on North America. However, the climate response over the NA/E sector to La Niña combined with the PDO is still unclear. Most of the above discussed previous work underlined the modulating role of PDO, but it will be interesting to examine the impact of various combinations of PDO and two types of La Niña. Hu and Huang (2009) pointed out that PDO besides the interdecadal time scale has also large variability at interannual and interseasonal time scales (e.g., at the ENSO time scale). Therefore, we mainly focus on the combined influence of PDO and the two types of La Niña on NAO-like climate anomalies and the associated dynamical processes.

The structure of this manuscript is organized as follows: The datasets, analysis methods, and the classification for two types of La Niña and their separation into different groups according to the phases of PDO are described in section 2. The combined climatic impact of PDO and the two types of La Niña over the NA/E sector is demonstrated in section 3, paying special attention to the possible physical processes and mechanisms. Discussion and summary of the results are provided in section 4.

2. Data and methods

a. Data

The monthly mean Hadley Centre Global Sea Ice and Sea Surface Temperature (HadISST1) datasets (Rayner et al. 2003) from 1854 to the present with 1° longitude/latitude resolution used in this study are derived from the Met Office Hadley Centre (http://www.metoffice.gov.uk/hadobs/hadisst/data/download.html). Atmospheric datasets are obtained from the National Centers for Environmental Prediction (NCEP)–National Center for Atmospheric Research (NCAR) reanalysis (Kalmay et al. 1996), including the monthly mean geopotential height and horizontal winds at 850 and 250 hPa, sea level pressure (SLP), and the daily mean geopotential height and horizontal winds at 250 hPa, which has a 2.5° × 2.5° horizontal resolution from 1948 to the present (http://www.esrl.noaa.gov/psd/gridded/data.ncep.reanalysis.html). University of Delaware air temperature and precipitation (v4.01) data (Matsuura and Willmott 2009) from 1901 to 2014 are employed to characterize the monthly mean surface temperature and land rainfall variability over Europe (http://www.esrl.noaa.gov/psd/gridded/data.UDel_AirT_Precip.html#detail). The Precipitation Reconstruction Dataset (PREC; Chen et al. 2002) from NOAA is used to describe the tropical precipitation anomalies (http://www.esrl.noaa.gov/psd/data/gridded/data.prec.html).

Multiple indices are also made use of in this study. The Niño-3.4 index based on HadISST1 datasets (http://www.esrl.noaa.gov/psd/gcos_wgsp/Timeseries/Data/nino34.long.anom.data) is derived from the Working Group on Surface Pressure (WGSP) of the Physical Sciences Division (PSD) in the Earth System Research Laboratory (ESRL). It is defined by the SST anomalies averaged in the region 5°N–5°S, 170°E–120°W. The station-based NAO index from 1865 to 2015 proposed by Hurrell (1995), defined as the difference of the normalized monthly mean sea level pressure between Lisbon, Portugal, and Stykkisholmur/Reykjavik, Iceland, is used to describe the variation of atmospheric circulation over the North Atlantic and western Europe (https://climatedataguide.ucar.edu/sites/default/files/nao_station_monthly.txt). The PDO index from 1900 to the present is used to determine the phase of Pacific decadal oscillation in winter (http://research.jisao.washington.edu/pdo/PDO.latest). As suggested by Hu and Huang (2009), the short time scale variations of PDO are not ignorable. Significant spectral power and relatively large variance result from short time scales and it is reasonable to use the original PDO index instead of only considering the interdecadal phase of low-pass-filtered PDO.

The overlapping period of all datasets and indices extends from January 1950 to December 2012. Anomalies of all variables are calculated as the deviation from the means of 1950–2012. We also examined the linearly detrended datasets, including the Niño-3.4 index, the NAO index, the PDO index, and all monthly mean variables, and did not detect relevant differences that would change our conclusions. In addition, the NOAA-CIRES Twenty-first Century Reanalysis (V2) dataset (Compo et al. 2011) from 1871 to 2012 is used to depict the impacts of two types of La Niña since 1900 and confirm the
Composite analysis is the primary method used in this study to investigate the different impacts of two types of La Niña associated with different phases of PDO. We adopt Cramer’s test to examine the difference between the means of subgroups and the mean of the whole group with the entire period of $tk$. Cramer’s test statistic is given by

$$t_k = \sqrt{n(N-2)/(N-n-n^2)} \times \tau_k,$$

where

$$\tau_k = (\bar{x}_k - \bar{x})/\sigma, \quad \bar{x}_k = (1/n) \sum_{i=k+1}^{k+n} x_i, \quad \bar{x}_k = (1/N) \sum_{i=1}^{N} x_i,$$

where $\bar{x}$ and $\sigma$ are the mean and standard deviation of the whole group with the entire period of $N$ years, respectively, and $\bar{x}_k$ is the mean of subgroup $k$ with the subperiod of $n$ years to be compared with $\bar{x}$. The test statistic $t_k$ accords with the Student’s $t$ distribution with $(N-2)$ degrees of freedom.

b. Extended Eliassen–Palm flux

The extended Eliassen–Palm (EP) flux, proposed by Hoskins et al. (1983) and modified by different authors thenceforth (Plumb 1985; Hendon and Hartmann 1985; Trenberth 1986), is a valuable diagnostic tool to demonstrate the interactive processes between synoptic eddies and the low-frequency flow. Here, the horizontal flux components of the extended EP flux from Trenberth (1986) is adopted:

$$E_u = \frac{1}{2}(v'^2 - u'^2)i, \quad -\bar{u'}\bar{v'}$$

where $u'$ and $v'$ represent synoptic-scale zonal and meridional winds, respectively, the overbars denote time average, and $\phi$ is latitude. To obtain the synoptic-scale disturbances, a Butterworth bandpass filter is applied to retain fluctuations with 2–8-day period (Murakami 1979).

In the upper troposphere, the divergence of the local extended EP flux associated with enhanced synoptic-scale eddy activity is accompanied by cyclonic forcing of eddy vorticity fluxes to the north of the divergence region and anticyclonic forcing to the south, which induces an eastward acceleration of the mean flow and a negative (positive) geopotential height tendency immediately to the north (south) (Lau 1988). The convergence of local extended EP flux acts opposite, leading to westward acceleration of the mean flow and positive (negative) geopotential height tendency immediately to the north (south).

c. Horizontal wave activity flux

In this study, we utilize the horizontal wave activity flux proposed by Takaya and Nakamura (1997, 2001) to portray the propagation features of quasi-stationary Rossby waves and the associated teleconnection patterns. Under the Wentzel–Kramers–Brillouin (WKB) approximation, this flux is parallel to the local group velocity of a stationary Rossby wave train and is independent of wave phase. The formulation of the horizontal wave activity flux in the log-pressure coordinate is given as follows:

$$W = \frac{1}{2|\bar{u}|} \left[ \frac{\bar{\psi}^2}{\bar{\psi}_{xx}^2} + \frac{\bar{\psi}_{yy}^2}{\bar{\psi}_{xy}^2} \right],$$

where $\psi$, $u$, and $v$ are the perturbed geostrophic streamfunction, the zonal wind velocity, and the meridional wind velocity, respectively. The overbar represents the basic states or climatologic means. The subscripts $x$ and $y$ indicate the partial derivatives in the zonal and meridional directions, respectively.

d. Classification of two types of La Niña

In some recent studies (Ashok et al. 2007; Kao and Yu 2009; Kug et al. 2009; Ren and Jin 2011), it was suggested that El Niño can be classified into CP and EP types on the basis of normalized $D(-1)F(0)$ (i.e., December–February, with December from the preceding year) Niño-3 and EMI indices or Niño-3 and Niño-4 indices. Here, $-1 \times D$ represents the year before (during) the mature phase of ENSO events. However, a few CP El Niños were misclassified when relying solely on indices instead of analyzing the corresponding SST anomaly patterns (Pascolini-Campbell et al. 2015). Dommenget et al. (2013) indicated that the CP events can be regarded as a nonlinear manifestation of the canonical (EP) events. Thus, it is as difficult to separate La Niña into two types based on present indices (Kao and Yu 2009; Kug and Ham 2011; Ren and Jin 2011) as it is for El Niños (Pascolini-Campbell et al. 2015). Lai et al. (2015) also pointed out that different types of El Niño can be seen as members of a continuum with few CP and EP El Niños as the end members of the distribution and a larger number of hybrid El Niños filling the distribution. Although not discussed in their paper, it is well possible that the same holds for La Niña and this is a ready explanation for the difficulties in classifying different types as shown in Fig. 2 of Frauen et al. (2014).
In view of the difficulty in clearly distinguishing two types of La Niña, we separate the sample by using just the Niño-3.4 index and then analyze the spatial distribution and evolution of SST anomalies as well as of atmospheric circulation anomaly patterns over the tropical Pacific. First, 18 La Niña events are selected by requiring the 3-month running mean Niño-3.4 index from January 1950 to December 2012 to be...
below $-0.5^\circ$C and to persist for at least four months. Here, we use four months instead of five months suggested by NOAA to retain some borderline cases. Considering the timing of the mature phase of La Niña and since its significant impact on extratropical circulation mainly occurs in winter, 21 La Niña winters are defined (1950/51, 1954/55, 1955/56, 1964/65, 1970/71, 1971/72, 1973/74, 1974/75, 1975/76, 1983/84, 1984/85, 1988/89, 1995/96, 1998/99, 1999/2000, 2000/01, 2005/06, 2007/08, 2008/09, 2009/10, 2010/11) from the 18 individual La Niña events. La Niñas obviously form clusters and sometimes last for more than one year. Then we define winters with large negative SST anomaly primarily located in the west (east) of 150°W, the boundary between the Niño-3 and Niño-4 areas, during the developing and mature phases of the events as CP (EP) La Niña winter following Zhang et al. (2015). Based on this definition, nine CP La Niñas (1950/51, 1973/74, 1975/76, 1988/89, 1998/99, 2000/01, 2008/09, 2010/11, 2011/12) and six EP La Niñas (1954/55, 1955/56, 1964/65, 1984/85, 1995/96, 2005/06) are identified, which is quite similar to the samples identified in previous studies (Yuan and Yan 2013; Zhang et al. 2015). In consideration of the salient effects of volcanic activity on the climate lasting at least 2 years, we exclude the 1983/84 La Niña winter following the strong volcanic eruption of El Chichon in 1982 (Graf et al. 2014). Besides that, when examining each individual case, two cases (1971/72 and 1974/75) exhibit different SST and atmospheric anomalies from others, whose significant signals are mainly located in the North Pacific instead of tropical Pacific (see Fig. A1 in the appendix). The large SST cooling is situated near the east coast of Mexico and extends southwestward to the tropics with relatively weak negative SST anomalies covering the equator, which is accompanied by large SST warming over the central North Pacific to its northwest (Figs. A1a,b). This structure of SST anomalies is similar to a negative PDO-like pattern rather than La Niña. It is also noteworthy that the corresponding anomalous centers of velocity potential lie in the north to equator (Figs. A1c,d). Since tropical large SST cooling and the associated atmospheric responses play a dominant role in the combined impacts of PDO and La Niña on the European climate, we further distinguish whether the equatorial SST cooling is a La Niña event or only a part of negative PDO phase by dividing the $D(-1)$ JF(0) Niño-3.4 index into the PDO-related part and the residual part. Considering that PDO (ENSO) is the principal mode at the decadal (interannual) time scale, the atmospheric responses to changes in PDO can be approximated by using (1) $D(-1)$ JF(0) Niño-3.4 as a proxy for ENSO, and (2) $D(-1)$ JF(0) Niño-3.4 as a residual component of PDO, which is considered to be the part of PDO unrelated to ENSO.

<table>
<thead>
<tr>
<th>Year</th>
<th>PDO</th>
<th>CP La Niña</th>
<th>EP La Niña</th>
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<tr>
<td>1938/39</td>
<td></td>
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<td>1995/96</td>
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<td>2000/01</td>
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<td>2005/06</td>
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scale, the PDO-related part is defined as the value regressed on the 7-yr running mean PDO index while the residual part is calculated by subtracting the PDO-regressed part from the original Niño-3.4 index. Following this procedure the 1971/72 and 1974/75 winters are removed from our list because their residual Niño-3.4 indices are no longer lower than −0.5°C and the anomaly patterns of SST and atmosphere are supposedly the result of the negative PDO phase without evident tropical diabatic heating anomaly (not shown).

Figure 1 shows the composites of D(−1)JF(0) SST anomalies for CP La Niña, EP La Niña, and the other three years (1970/71, 1999/2000, 2007/08) defined as mixed La Niña by Zhang et al. (2015) with large cooling of SST stretching across both EP and CP areas during the mature phase. Both CP and mixed La Niña exhibit an east–west dipole pattern in SST anomaly, with a cooling area over the tropical central-eastern Pacific and a warming area over the Philippine sea extending north-eastward (Figs. 1a,c); however, there still exists a little structural difference in the negative equatorial SST anomaly. The large SST cooling for CP La Niña is situated to the west of 150°W (Fig. 1a), whereas the mixed La Niña exhibits two centers of negative SST anomaly straddling 150°W (Fig. 1c). Except for the differences observed over the eastern Pacific, the geographical position of SST anomaly patterns for mixed La Niña is similar to that for CP La Niña, with a spatial correlation coefficient of 0.95 exceeding the 99% confidence level. The responses of the tropical and extratropical atmosphere to mixed La Niña (including anomalous velocity potential, divergent wind, Walker circulation, SLP, surface horizontal wind, and precipitation) also show resembling spatial features to CP La Niña (not shown). Therefore, on account of the similar SST and atmospheric circulation anomalies, the three mixed La Niña winters are merged with the CP La Niña class. As for EP La Niña, significant large SST cooling is mainly located to the east of 150°W with insignificant SST warming covering the northwest Pacific to the east of the Philippine Sea (Fig. 1b). Contrasting with CP La Niña, the dipole SST anomaly pattern shifts toward the equatorial eastern Pacific, accompanied by significant negative SST anomaly over the South China Sea. Along with the differences in SST anomaly pattern, the corresponding tropical atmospheric responses move eastward by about 15°–20° longitude (not shown). It is worth noting that the amplitude of negative SST anomaly and the associated tropical atmospheric responses are weaker than those for CP La Niña. However, we will mainly focus on discussing the difference in spatial patterns for different types of La Niña in the rest of this paper, since the intensity of these anomalies could be highly influenced by the number of years in the composite maps.

To further substantiate the robustness of the results, we select another 11 La Niña winters applying the same method from 1900 to 1950 and separate them into two types based on the seasonal evolution of tropical SST anomaly pattern, the location of large SST cooling in winter and the related atmospheric responses.

3. Results

a. Combined impacts of PDO and the two types of La Niña on NAO and the associated climate anomalies

The results of classification of La Niña winters are listed in Table 1 together with the phase of PDO. CP and EP La Niña winters are further separated into two groups, respectively, according to the different phases of the winter PDO. Here, CP and EP La Niña conjunct with the positive (negative) PDO phase are denoted as CP+/PDO+ (CP−/PDO−) and EP−/PDO+ (EP+/PDO−), respectively. Notably, the occurrence of CP La Niña winters is strongly conjunct with the negative PDO phase, except the 2000/01 La Niña winter, while the relation between EP La Niña and PDO with 5 EP−/PDO+ winters and 5 EP+/PDO− winters does not have an obvious bias.

Generally, an ENSO event, cold or warm, reaches its mature phase during the late autumn and winter, while the associated significant atmospheric responses in the form of NAO covering the NA/E sector mainly occurs delayed in the mid to late winter (Huang et al. 1998;
Gouirand and Moron 2003; Knippertz et al. 2003; Brönnimann et al. 2007), which has large impacts on the regional climate variability (Shabbar et al. 2001; Huang et al. 2006). Figure 2 shows the monthly evolution of the NAO index from June(−1) to May(0) since 1950 for the CP−/PDO− (red), EP−/PDO− (blue), and EP−/PDO+ (black), respectively. When the PDO is in its negative phase, the NAO signal is stronger than in PDO positive phase, but with reversed sign in late winter for the different types of La Niña. For CP La Niña, statistically significant (filled dots) positive NAO begins in January(0), reaches its peak in February(0), and maintains to March(0) (see red solid line in Fig. 2). On the contrary, a nearly symmetric negative NAO value emerging from January(0) to March(0) with strongest and significant signal in February(0) (see blue solid line in Fig. 2) can be detected for EP La Niña. When the PDO is in its positive phase, EP La Niña exhibits a weaker and not statistically significant negative NAO signal in the wintertime other than February(0) with a positive value but close to zero (see dashed black line in Fig. 2). 2000/01 is the only CP La Niña winter accompanied with positive PDO phase. Its maximum negative NAO mainly occurs in December(−1) and March(0) rather than January(0) or February(0) (not shown). The composites for extended time series, as shown in Fig. A3, display approximately a similar evolution of the NAO index except a few differences, especially during EP La Niña. The negative NAO signal persists from December(−1) to March(0) with a significant peak in January(0) rather than February(0) for the EP−/PDO− winter (see blue solid line in Fig. A3), while a stronger negative NAO signal with longer persistence...
occurs in the extended wintertime for the EP
/ PDO
winter (see dashed black line in Fig. A3). We also show the JFM(0) NAO index since 1900 for CP (red) and EP (blue) La Niñas versus the PDO value in Fig. 3. Here, the individual cases after (before) 1950 are represented by a dot (circle). The NAO index is positive (negative) in JFM(0) for each CP (EP) La Niña winter conjunct with the negative PDO phase. When the PDO value is positive, four out of five EP La Niña winters have a negative, but weaker, NAO. Thus, the NAO signal for each individual case is consistent with the results of the composite analysis, especially during the combination of negative PDO and La Niña.

Since the most notable anomaly of NAO is primarily detected during late winter, we shall concentrate on the JFM(0) climate responses over the NA/E sector to the two types of La Niña conjunct with different phases of PDO. Considering the quality and homogeneity of the datasets, we mainly focus on the period extending from 1950 to the present and then utilize the longer dataset since 1900 to verify the conclusions, which are presented in the appendix. Since there is only one CP /PDO+ winter, only the other three conditions are discussed below.

Figure 4 shows SLP and surface wind (left panel) and 250-hPa geopotential height (right panel) anomalies in JFM(0) during the CP /PDO-, EP /PDO-, and EP /PDO+ winters. When a La Niña event combines with the negative phase of PDO, the NAO-like atmospheric response is evident with two significant anomaly centers and approximately reversed structures for CP and EP types. For the CP /PDO- winter, a significant high SLP anomaly to the south of 55°N and low pressure to the north are accompanied by surface westerly wind anomalies in the mid-to high latitudes extending zonally from the western North Atlantic to western Europe in JFM(0) (Fig. 4a). This resembles a positive NAO-like pattern. In comparison, a roughly reverse NAO-like response in SLP and surface wind anomalies can be observed during the EP /PDO- winter (Fig. 4c). The negative NAO signal of EP La Niña can also be detected when combining with the positive phase of PDO. However, only the positive SLP anomaly is statistically significant and it is not in the climatological position of NAO, but is rather shifted to the east toward central and northern Europe (Fig. 4e). The 250-hPa geopotential height anomaly composites indicate that the anomalous pattern in the upper troposphere is similar to that in the lower troposphere over the NA/E sector (see Figs. 4b,d,f). This is consistent with Ting (1996), suggesting that the tropical heating anomaly can trigger an extratropical barotropic response in the atmospheric circulation. Such inverse anomaly patterns also emerge in the composites of zonal wind anomalies at 250 hPa between CP and EP La Niña, especially conjunct with the negative PDO phase, as depicted in Figs. 5a and 5b. For the CP /PDO- winter, the zonal wind anomalies have a tripolar structure with westerly wind anomalies in the midlatitudes.
and easterly anomalies in the high latitudes as well as in the sub-tropics (Fig. 5a), which is in accordance with the positive NAO-like atmospheric response. As a result, the North Atlantic jet stream significantly moves northward and extends farther eastward reaching northwestern Europe. The North Africa jet is also strengthened at its core and becomes narrow. For the EP⁻/PDO⁻/winter, an analogous but reversed tripolar structure north of approximately 45°N in zonal wind anomalies at 250 hPa is found (Fig. 5b), resulting in a weak North Atlantic jet and a northward shifted North Africa jet. In contrast to the EP⁻/PDO⁻ winter, the tripolar zonal wind anomalies move eastward during the EP⁻/PDO+ winter (Fig. 5c), influencing the shift of the jet stream over northern Africa.

It is known that the winter surface air temperature and precipitation anomalies over Europe may be partly affected by the NAO (e.g., Jones et al. 2003). Hence, the
climate response to CP La Niña is quite different from EP La Niña because of the approximately opposing NAO-like pattern when combining with the negative phase of PDO. Figure 6 shows the composites of anomalous surface temperature and precipitation in JFM(0). The results for the CP−/PDO−, EP−/PDO−, and EP+/PDO+ are shown in the upper, middle, and bottom panels of Fig. 6, respectively. For the CP−/PDO− winter, a significant warm SAT anomaly up to +1°C covers the central, western, and northern parts of Europe (Fig. 6a). This is attributed to the midlatitudinal westerly wind anomaly related to the positive NAO phase in the lower troposphere, which extends eastward to central and northern Europe and transports relatively warm and moist air from the North Atlantic to these regions (Fig. 4a). The anomalous precipitation displays a dipolar structure with excessive (deficient) precipitation over northern (southern) Europe (Fig. 5a). Because of the northward movement of the North Atlantic jet, the jet exit region covers northern Europe inducing there
anomalous rising motion, which is favorable for wet conditions in Scandinavia. This is supported by the increasing warm and moist air advection by the strengthened surface westerly winds in the midlatitudes. The dry conditions around the northern parts of the Mediterranean Sea may be ascribed to the narrowed entrance region of the North Africa jet suppressing the local rising motion to the north. In contrast, for the EP−/PDO− winter, the midlatitudinal easterly wind anomaly weakens the background westerlies and reduces the intrusion of warm and moist air from the North Atlantic to Eurasia, whereas the southwestward wind anomaly at the southeastern flank of the anomalously strong anticyclonic anomaly over the central northern North Atlantic tends to transport cold and dry air from high latitudes to central and northern Europe (Fig. 4c). This results in colder JFM(0) over most regions of Europe (Fig. 6c). Simultaneously, the precipitation is enhanced (reduced) in southern (northern) Europe (Fig. 6d), which is associated with the changed position of the North Africa (Atlantic) jet (Fig. 5b). It is noteworthy that the SAT and precipitation anomalies of the EP−/PDO− winter (see Figs. 6e, f) are similar to that of the EP−/PDO− winter, but much stronger and more significant. This is because the negative NAO-like atmospheric circulation anomalies are shifted toward to the east and the center of the anomalous anticyclone dominates over northern Europe, strengthening the local easterly wind anomaly in the lower and upper troposphere.

More individual La Niña winters are selected and separated since 1900 in order to confirm the reliability of the results (see the appendix). When combining with the negative phase of PDO, the two types of La Niña for the extended time series also exhibit opposite
NAO-like atmospheric responses both at the surface (Fig. A4a,c) and in the upper troposphere (Fig. A4b,d), and lead to reversed SAT and precipitation anomalies in Europe, which are consistent with the structures obtained from Figs. 4a–d and 6a–d, respectively, although a subtle difference exists in the intensity of anomalous activity centers. The combination of negative PDO and CP La Niña shows a stronger positive NAO-like pattern (Fig. A4a,b), inducing an enhanced warmer European winter (Fig. A6a) and wetter (drier) conditions in the northern (southern) part (Fig. A6b). The combination of negative PDO and EP La Niña shows a slightly weaker negative NAO-like pattern (Fig. A4c,d), which is accompanied by the weakening of SAT cooling in western Europe (Fig. A6b). However, no distinct differences can be detected in European dipole rainfall anomalies (Fig. A6d). For the EP+/PDO+ winter, the eastward shifted negative NAO signal (Fig. A4e,f) and the related climate responses (Figs. A6e,f) also resemble the patterns in Figs. 4e,f and Figs. 6e,f, respectively, but are obviously somewhat weaker.

b. Possible mechanisms

According to the composite analysis above, the two types of La Niña with different SST anomaly patterns display approximately opposite anomalous atmospheric and climate responses over the NA/E sector especially conjunct with the negative PDO phase. In the following we explore the mechanism bridging between the tropical SST anomaly and the extratropical circulation anomalies. Here, we primarily focus on the possible physical mechanism for the formation of NAO anomalies. Figure 7 demonstrates the anomalies of JFM(0) averaged 250-hPa streamfunction and horizontal wave activity flux, the 250-hPa extended EP flux.
and its divergence, and the 250-hPa storm track activity for the CP−/PDO− winter. Figures 8 and 9 are similar to Fig. 7, but for the EP−/PDO− and EP−/PDO+ winters, respectively.

When combining with the negative phase of PDO, CP and EP La Niña exhibit distinct differences in the wave train from the North Pacific to the North Atlantic. For the CP−/PDO− winters, a strong and significant anticyclonic anomaly is observed over the North Pacific, indicating the weakening of the Aleutian low. The anticyclonic perturbation first propagates northeastward to western Canada and then propagates southward to the southeastern United States, forming a negative PNA-like wave train corresponding to previous studies (Honda et al. 2001; Honda and Nakamura 2001; Pinto et al. 2011) with alternating anticyclonic, cyclonic, and anticyclonic anomaly action centers (Fig. 7a).

Supported by the occurrence of the positive geopotential height anomaly over the southeastern United States, the storm track represented by the movement of...
FIG. 11. JF(0) 50-hPa geopotential height anomaly (contour; interval: 50 m) during (a) strong (1950/51, 1973/74, 1975/76, 1988/89, 1999/2000, 2010/11), (b) weak (1970/71, 2008/09, 2011/12), and (c) normal (1998/99, 2007/08) polar vortex for the CP–PDO winters since 1950. (b),(d),(f) As in (a),(c),(e), but for 500 hPa. Shadings for contour indicate the 90% confidence level.
synoptic-scale transient eddies is strengthened and moves slightly toward the polar latitudes (Fig. 7c) together with a northward migration of the North Atlantic jet. This is linked to a larger Eady growth rate (not shown) and, hence, stronger baroclinicity in the vicinity of Newfoundland, providing a favorable baroclinic environment for the growth of synoptic-scale transient waves reinforcing the local interaction between transient eddies and the mean flow. Simultaneously, an EP flux divergence anomaly around $50^\circ$N extends from the southeastern United States eastward to the northwest of Great Britain (Fig. 7b). As mentioned in section 2c, the divergence of EP flux is expected to enhance the forcing of cyclonic vorticity to its north and anticyclonic vorticity to its south, and a westerly anomaly between these two cells (Lau 1988; Chen et al. 2014). An anomalous anticyclone, partly originating from the eddy–mean flow interaction, appears to the south of the EP flux divergence in the region $30^\circ$–$55^\circ$N, $70^\circ$W–$20^\circ$E, and an anomalous cyclone appears at the northern flank of the EP flux anomaly near southern Greenland. The accelerated westerly wind related to the EP flux divergence in turn intensifies the local storm track activity, forming a continuous positive eddy–mean flow feedback to support circulation anomalies associated with the positive NAO as described above. The EP flux convergence located in the subtropical North Atlantic is also propitious for the enhancement of an anticyclonic perturbation to its north, forming the southern part of the positive NAO-like pattern. For the EP–/PDO$-$ winters, a relatively weak anticyclonic perturbation over the North Pacific slightly shifts toward the polar latitudes and propagates northeastward through central North America into southern Greenland (Fig. 8a), composing a negative PNA wave train that is quite different from the typical structure. It is worth noting that the zonal wave train propagating in the high latitudes is stronger and more evident in D($-1$)JF(0) (not shown). The easterly anomaly at the southern flank of the anomalous anticyclone over southern Greenland suggests the suppression of storm track activity in the midlatitude North Atlantic (Fig. 8c), giving rise to the emergence of the local EP flux convergence around $50^\circ$N (Fig. 8b). The convergence of EP flux leads to an anomalous anticyclone (cyclone) to its north (south) and decelerates the eastward wind, which further attenuates the storm track activity and in turn increases the EP flux convergence. The positive feedback mechanism between transient eddies and the mean flow makes a contribution to the establishment and development of a
negative NAO-like circulation anomaly. Considering that only three EP−/PDO− winters are observed, thus providing a very small sample, we also examine each individual winter and show the existence of the analogous wave train structure as well as the local interaction between synoptic eddy and low-frequency flow for each individual case. Besides the important role of the Rossby wave train and the transient eddy–mean flow interaction, nonrotational flow may also affect the atmospheric circulation over the NA/E sector for the combination of La Niña and negative PDO by changing the Walker and Hadley circulation. This is shown in Fig. 10 depicting the differences of divergent wind and velocity potential anomaly between low-level (850 hPa) and high-level (250 hPa) in JFM(0). CP−/PDO− winter displays an anomalous two-cell Walker circulation with a divergence anomaly over the central-eastern Pacific and a convergence anomaly over the Maritime Continent and northern South America (Fig. 10a). The significant convergence anomaly, indicating an ascending motion anomaly, located over northern South America, alters the local Hadley circulation and leads to a significant but weak divergence and descending motion anomaly to its north (Fig. 10a), which may slightly contribute to the maintenance of the anomalous anticyclone (Fig. 7a). In comparison, EP−/PDO− winter displays a single cell Walker circulation with a convergence anomaly over the central-western Pacific and a divergence anomaly over the area from the eastern Pacific to the tropical Atlantic (Fig. 10b). The significant divergence anomaly may induce a weak convergence (ascending motion) anomaly to its northeast over the subtropical Africa (Fig. 10b), which also favors the negative NAO-like pattern (Fig. 8a). However, in the EP+/PDO+ winter, the northeastward propagation of the wave train is missing and the North Pacific anticyclonic anomaly is rather weak and insignificant (Fig. 9a). In addition, no obvious velocity potential anomaly can be detected covering the tropical and extratropical Atlantic (Fig. 10c). It seems that the formation of atmospheric anomalies over Europe has no connection with the SST anomalies over the tropical Pacific, but may be affected by other factors.

Perlwitz and Graf (1995) pointed out that the stratospheric polar vortex plays an important role in modulating Northern Hemispheric climate anomalies. On the one hand, barotropic effects in polar latitudes induce tropospheric pressure anomalies over the Arctic that are similar to the stratospheric signal (Ambaum and Hoskins 2002; Castanheira et al. 2009; Graf et al. 2014). On the other hand, at the edge of the polar vortex, midlatitudinal zonal-mean zonal wind anomalies penetrate into the troposphere and interact with topography, resulting in
NAO-like atmospheric anomalies (Castanheira et al. 2009; Graf et al. 2014). The changed shear of upper tropospheric winds and the latitudinal position of the jet can also alter the anomalies of planetary wave phase and amplitude by affecting the Eady growth rate in the storm track region and the intensity of the tropospheric eddy feedback, respectively (Walter and Graf 2005; Scaife et al. 2012; Garfinkel et al. 2013; Graf et al. 2014). Here, we also examine the JF(0) (January–February) geopotential height anomalies in the stratosphere (50hPa) and calculate a Polar Vortex Index (PVI: Graf et al. 2014), describing the strength of the stratospheric polar vortex over the northern polar cap (north of 65°N). Winters characterized by strong and weak polar vortex correspond to those years when PVI is, respectively, above 0.5σ and below −0.5σ. Two out of three EP−/PDO+ winters (1984/85, 2005/06) are accompanied by a significant, weaker polar vortex (not shown). The stratospheric high pressure anomaly over the Arctic elongates toward Europe, partly illustrating the occurrence of European atmospheric anomalies in the troposphere through the barotropic effects and the transient eddy–mean flow interaction. It is noteworthy that different stratospheric signals can be detected for two types of La Niña conjunct with the negative PDO phase, which may influence tropospheric teleconnection patterns over the NA/E sector (Walter and Graf 2005, 2006). Six CP−/PDO+ winters (1950/51, 1973/74, 1975/76, 1988/89, 1999/2000, 2010/11) are accompanied by a strong polar vortex (Fig. 11a), resulting in a meridional dipole structure of geopotential height anomalies over the central North Atlantic (Fig. 11b); three CP−/PDO− winters (1970/71, 1998/99, 2008/09, 2011/12) are accompanied by a weak polar vortex (Fig. 11c), triggering a dipole pattern situated in the eastern North Atlantic (Fig. 11d); and the others (1998/99, 2007/08) are accompanied by a normal polar vortex (Fig. 11e) with anomalous centers shifting toward western North Atlantic (Fig. 11f) (Walter and Graf 2005, 2006). For each EP−/PDO− winter, different stratospheric signals will alter the atmospheric anomalies in the troposphere, but the negative NAO-like pattern is still robust (not shown).

We also employed the Twentieth Century Reanalysis (V2) dataset to calculate the horizontal wave activity flux and received negative PNA-like structures (not shown) in JFM(0) comparable to Figs. 7a and 8a, with a conventional PNA-like wave train during the CP−/PDO+ winter and a zonally propagating wave train in the high latitudes during the EP−/PDO− winter. For the EP−/PDO+ winter, no wave train propagating from the North Pacific to the North Atlantic can be detected (not shown).

Thus, when combining with the negative phase of PDO, CP and EP La Niña excite different negative PNA-like wave trains propagating from the North Pacific to the North Atlantic, which is likely attributed to the differences in precipitation anomalies (Figs. 12b,d) as a response to different types of SST anomalies and the corresponding background wind, especially the subtropical jet stream over the Northern Hemispheric oceans. Then, consistent with prior suggestions (Limpasuvan and Hartmann 1999; Chen et al. 2014), through the interaction between synoptic eddies and the mean flow and the associated vorticity transport, the inverse NAO-like circulation anomalies for the two types of La Niña gradually establish over the North Atlantic and develop and maintain during late winter. Through altering the Walker and Hadley
circulation, the nonrotational flow also makes some small contribution to the maintenance of the NAO-like anomaly pattern.

4. Discussion and summary

Previous studies have demonstrated opposite responses of NAO to different types of La Niña (Zhang et al. 2015). They attributed the contrasting atmospheric anomalies to the different SST cooling patterns based on modeling experiments. In our (reanalysis) data-based study, we further consider the combined influences of PDO and two types of La Niña on NAO and highlight the possible physical processes for the occurrence of these different NAO anomalies especially during the combination of negative PDO and La Niña.

For the CP–/PDO– winter, a positive NAO-like signal with anomalous high (low) pressure to the south (north) of 55°N dominates the JFM(0) climate anomalies of the NA/E sector, which is accompanied by SAT warming covering the whole of Europe, except the eastern part, and a dipolar structure of precipitation anomalies with excessive (deficient) rainfall over northern (southern) Europe. To explore the possible mechanisms causing the NAO-like response, we examine the horizontal wave activity flux, the anomalous streamfunction, extended EP flux, and anomalous storm track activity in JFM(0) using the reanalysis data for the period 1950–2012. A significant anticyclonic perturbation located in the central North Pacific first propagates northeastward into western Canada and then turns southeastward, forming a negative PNA-like wave train with a high pressure anomaly over the southeastern United States. The positive geopotential height anomaly not only strengthens the baroclinicity at the entrance of the storm track in the midlatitude North Atlantic, but also promotes the amplification of EP flux divergence with advection of cyclonic vorticity to its north and anticyclonic vorticity to its south. Then, by the effects of local interaction between transient eddies and the low-frequency flow, the anomalous anticyclone constantly develops farther downstream over the North Atlantic. To its north, an anomalous cyclonic anomaly gradually establishes over southern Greenland. Both anomaly centers together constitute a positive NAO-like atmospheric circulation anomaly pattern. For the EP–/PDO– winter, a weaker anticyclonic perturbation in the North Pacific shifts northward and propagates farther northeastward into southern Greenland, composing a wave train that is quite different from the conventional negative PNA structure. The anomalous anticyclone over southern Greenland contributes to the development and maintenance of a negative NAO signal through local synoptic eddy–mean flow interaction and the associated vorticity transport, giving rise to SAT cooling over Europe and less (more) rainfall in the northern (southern) part. It should be noted that we examined each individual case for the EP–/PDO– winter and found analogous zonal wave train structure as well as local eddy–mean flow interaction in the extratropics for all three events (not shown).

Therefore, the wave trains emanating from the anticyclonic perturbation over the North Pacific and the synoptic eddy–mean flow interactive processes play an essential role in forming and maintaining the NAO-like atmospheric responses for the two types of La Niña when combining with the negative phase of PDO. CP and EP La Niña exhibit different structures of
tropospheric Rossby wave trains as shown in Figs. 12a and 12c, leading to opposite NAO signals. For CP La Niña, significant tropical negative geopotential height anomalies extend from the central Pacific to America with two symmetrical centers around 140°W straddling the equator (Fig. 12a). To its north, an anomalous anticyclonic perturbation develops over the eastern North Pacific and then propagates into the southeastern United States in the form of a negative PNA-like wave train (see Figs. 7a and 12a). For EP La Niña, significant negative geopotential height anomalies can be observed covering the whole tropics except the eastern Pacific (Fig. 12c). To the north of low pressure anomalies, an anomalous anticyclone, shifted northwestward compared to CP La Niña, stretches across the North Pacific and then zonally propagates eastward into Greenland (see Figs. 8a and 12c). It is worth noting that the zonal wave train and its related geopotential height anomaly centers, especially the ones located in the North Pacific, are much stronger and more prominent in DJF(0) (not shown). Graf and Zanchettin (2012) indicated that two types of El Niño are characterized by the different location of anomalous convective heating along the equatorial Pacific relative to the subtropical jet (STJ), determining the amplitude of the atmospheric disturbance and the propagation path. This can possibly explain the different Rossby wave trains and climate responses. We may speculate that the same holds for the two types of La Niña due to differences in the location and intensity of anomalous precipitation (Figs. 12b,d) as a response to different types of SST anomalies as well as the corresponding background upper tropospheric...
westerly winds over the Northern Hemispheric oceans. In addition, two-cell and single cell anomalous Walker circulation occurs during the CP\textsuperscript{2}/PDO\textsuperscript{2} and EP\textsuperscript{2}/PDO\textsuperscript{2} winter, respectively, due to different SST anomaly patterns, which may contribute to the maintenance of NAO signals through the alteration of nonrotational flow and the local Hadley circulation. However, because of the small sample especially for the EP\textsuperscript{2}/PDO\textsuperscript{1} winter, these results and speculations need to be further verified by numerical experiments. López-Parages et al (2015, 2016) operated a set of partially coupled experiments with different SST anomaly patterns superimposed over distinct tropical Pacific and Atlantic SST mean states by using a global atmospheric general circulation model. The results demonstrated that the teleconnection between two types of ENSO and NA/E climate at interannual timecales is possibly modulated by the multidecadal changes in the SST. In future studies, numerical analysis will be performed to investigate the combined role of negative PDO and two types of La Niña.

For the EP\textsuperscript{2}/PDO\textsuperscript{1} winter, a negative NAO signal can also be observed with a significant anticyclonic anomaly dominating over Europe, resulting in a colder winter and a drier (wetter) condition in the western and northern (southern) part. It seems, however, that there is no connection with the SST anomaly over the tropical Pacific since no evident wave train originates from the North Pacific and propagates into the NA/E sector. The formation of the anomalous anticyclone over Europe is still unclear and may be influenced by other factors.

The stratospheric polar vortex is also a key factor in winter to influence the tropospheric atmospheric anomalies over the Northern Hemisphere through barotropic effects and transient eddy–mean flow interaction (Perlwitz and Graf 1995; Graf et al. 2014). Two out of three EP\textsuperscript{2}/PDO\textsuperscript{1} winters experience a weak polar vortex, giving rise to an anticyclonic anomaly dominating the European climate variability. 1995/96 is the only EP\textsuperscript{2}/PDO\textsuperscript{1} winter accompanied by a strong polar stratospheric vortex. However, in contrast to expectations from previous studies that would suggest a positive NAO for strong polar vortex winters, a negative NAO signal is detected that may be affected by internal variability or other external forcings. For the CP\textsuperscript{2}/PDO\textsuperscript{2} winters, the position of positive NAO-like response will be modulated by the strength of the polar vortex with a meridional dipole structure over the central North Atlantic when the polar vortex is stronger, over the eastern North Atlantic when the polar vortex is weaker, and over the western North Atlantic when the polar vortex is normal. For the EP\textsuperscript{2}/PDO\textsuperscript{1} winter, different states of the polar vortex have little demonstrable impact on the negative NAO-like pattern in the troposphere. Although the strength of the polar vortex may modulate the tropospheric response over the NA/E sector to the two types of La Niña conjunct with the negative PDO phase, the NAO signal is stable and robust, suggesting
the dominating role of anomalous convection over the tropical Pacific. It is noteworthy that low Arctic sea ice in summer (Cohen et al. 2013; Jaiser et al. 2013), as well as extensive Siberian snow cover in October (Cohen et al. 2007, 2013), will trigger a weaker stratospheric polar vortex in the following winter months through enhancing the planetary wave activity and their interaction with baroclinic processes (Jaiser et al. 2013). Then, the stratospheric signal propagates down into the troposphere, leading to wave anomalies projecting on a negative NAO-like pattern. In addition, during periods of strong stratospheric polar vortex, a basinwide tripole correlation pattern exists between SST anomaly over North Atlantic and tropospheric variability (Graf and Walter 2005), indicating that a tripole SST anomaly pattern contributes to the formation of NAO-like response by the interaction between the heating-forced anomalous flow, the position of the region of maximum baroclinicity and Eady growth rate, and the Atlantic storm track (Peng et al. 2003).
To verify the robustness of the conclusions, more individual La Niña winters were selected and analyzed using long datasets since 1900. Then, similar atmospheric and climate anomalies can be detected over the NA/E sector, except some small difference in the intensity, including SLP, surface wind, 250-hPa geopotential height, 250-hPa zonal wind, SAT, and precipitation. When combining with the negative phase of PDO, CP La Niña exhibits a much stronger positive NAO-like response, while EP La Niña exhibits a slightly weaker negative NAO-like response. For the EP+/PDO+ winter, the anomalous anticyclone controlling the whole of Europe is somewhat weaker, giving rise to attenuated SAT and precipitation anomalies. Thus, the results based on the shorter and more complete and reliable data are supported with the longer data.

In this study, we suggest that, conjunct with the negative PDO phase, CP La Niña is favorable for the emergence of positive NAO-like atmospheric circulation anomalies, while EP La Niña exerts NAO anomalies of opposite sign. Our results may contribute to improving the seasonal forecast skill in late winter of SAT and precipitation anomalies over Europe. In addition, the modulating effects of the background state on the planetary wave propagation should be further considered in the seasonal forecasts. Graf and Zanchettin (2012) suggested the indispensable role of the STJ and its strength, position, and extension because only a strong zonal jet can act as a zonal waveguide. The relative geographical position between the STJ and the meridional path of the Rossby waves is also important since waves can only be trapped when passing through the jet. Thus, it will be of great interest to investigate the modulation of the climate anomaly patterns due to modifications in the background state. This can be achieved by a series of sensitivity experiments with background zonal winds and diabatic heating or SST anomaly patterns prescribed or pre-selected as external forcing.

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APPENDIX

Extended Time Series Results

Figs. A1–A7 are corresponding figures that utilize the longer dataset since 1900 to verify the conclusions.

REFERENCES


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