PDO-Related Wintertime Atmospheric Anomalies over the Midlatitude North Pacific: Local versus Remote SST Forcing

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ABSTRACT

Observed wintertime atmospheric anomalies over the central North Pacific associated with the Pacific decadal oscillation (PDO) are characterized by a cold/trough (warm/ridge) structure, that is, an anomalous equivalent barotropic low (high) over a negative (positive) sea surface temperature (SST) anomaly. While the midlatitude atmosphere has its own strong internal variabilities, to what degree local SST anomalies can affect the midlatitude atmospheric variability remains unclear. To identify such an impact, three atmospheric general circulation model experiments each having a 63-yr-long simulation are conducted. The control run forced by observed global SST reproduces well the observed PDO-related cold/trough (warm/ridge) structure. However, the removal of the midlatitude North Pacific SST variabilities in the first sensitivity run reduces the atmospheric response by roughly one-third. In the second sensitivity run in which large-scale North Pacific SST variabilities are mostly kept, but their frontal-scale meridional gradients are sharply smoothed, simulated PDO-related cold/trough (warm/ridge) anomalies are also reduced by nearly one-third. Dynamical diagnoses exhibit that such a reduction is primarily due to the weakened transient eddy activities that are induced by weakened meridional SST gradient anomalies, in which the transient eddy vorticity forcing plays a crucial role. Therefore, it is suggested that midlatitude North Pacific SST anomalies make a considerable (approximately one-third) contribution to the observed PDO-related cold/trough (warm/ridge) anomalies in which the frontal-scale meridional SST gradient (oceanic front) is a key player, although most of those atmospheric anomalies are determined by the SST variabilities outside of the midlatitude North Pacific.

1. Introduction

During the past decades, a number of observational studies have revealed that the midlatitude ocean–atmosphere system has significantly coherent variabilities on decadal-to-interdecadal time scales in both oceanic and atmospheric spatiotemporal structures (Cayan 1992a,b; Deser and Blackmon 1993; Kushnir 1994; Hurrell 1995; Dickson et al. 1996). In around 1976/77, a striking decadal change happened in the North Pacific ocean–atmosphere system with sea surface temperature (SST) cooling in the western-central North Pacific and Aleutian low strengthening and southeastward shifting. Such a decadal change is called Pacific decadal oscillation (PDO) (Mantua et al. 1997;

Minobe 1997; Enfield and Mestas-Núñez 1999; Zhu and Yang 2003; Deser et al. 2004). The covarying of ocean and atmosphere associated with PDO indicates the existence of interaction between ocean and atmosphere in the midlatitudes.

Since the internal variability of midlatitude atmosphere is extremely strong, there has been a long emphasis on the atmospheric forcing on the ocean in the midlatitude air–sea interaction (Hasselmann 1976; Battisti et al. 1995; Frankignoul et al. 1997), as the ocean has a relatively weak feedback on the atmosphere. However, complex coupled ocean–atmosphere model simulations have shown that the decadal variability of midlatitude SST can be strengthened by air–sea coupling process (Delworth and Greatbatch 2000). The midlatitude air–sea coupling has been considered one of causes of the decadal variability in a number of recent studies (Mantua and Hare 2002; Liu and Wu 2004; Zhong et al. 2008; Newman et al. 2016; Liu and Di Lorenzo 2018). The atmospheric forcing of the ocean is well examined, but to what extent the basin-scale

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midlatitude SST variabilities could affect the large-scale atmospheric circulation is still an open question.

Previous studies showed that the atmospheric response to midlatitude SST anomalies can be completely different in different models. Some demonstrated that the atmospheric response is barotropic, with atmospheric high (low) pressure over warm (cold) SST anomalies as in observations (Palmer and Sun 1985; Peng et al. 1995, 1997; Sun et al. 2018) or atmospheric low pressure over warm SST anomalies (Pitcher et al. 1988; Kushnir and Lau 1992; Peng and Whitaker 1999), while others showed that the atmospheric response is baroclinic in linear models (Hoskins and Karoly 1981; Kushnir and Held 1996). In terms of Peng and Whitaker (1999), the structure of the atmospheric response to diabatic heating is baroclinic, and with the help of transient eddies, the response turns to be barotropic. Therefore, midlatitude air–sea interaction is more complicated than expected when atmospheric transient eddy activity is involved.

The midlatitude atmosphere has strong baroclinicity, and thus synoptic transient eddies accompanying the storm track develop actively (Ren et al. 2010; Chu et al. 2013; Liu et al. 2014). With high-resolution satellite data, recent studies revealed a consistency between the atmospheric storm track and oceanic front (Nakamura and Shimpo 2004; Minobe et al. 2008; Nakamura et al. 2008; Wang et al. 2017). A variety of model simulations suggested that the oceanic front zone is a key region. The effect of a large SST gradient in oceanic front zone can propagate from the lower to the upper troposphere, bringing an indirect influence on atmospheric circulation (Feliks et al. 2004, 2007; Sampe et al. 2010; Yao et al. 2016, 2017; Chen et al. 2019; Wang et al. 2019). However, previous sensitivity simulations altered the large-scale SST greatly when the role of oceanic front was identified; that is, the impacts of both the SST itself and the SST meridional gradient are mixed. Isolating the role of oceanic front anomalies from large-scale SST variabilities is also an open question.

Furthermore, based on observational analysis, Fang and Yang (2016) proposed a hypothesis in which there is a positive feedback between midlatitude ocean–atmosphere system. During the PDO warm phase, the strengthened Aleutian low drives negative SST anomalies through increasing upward surface heat flux and southward Ekman advection. In the southern flank of the cooling SST, the meridional SST gradient is strengthened, leading to an enhanced atmospheric baroclinicity above, which favors the generation of more atmospheric transient eddies. Through the transient eddy vorticity forcing, an equivalent barotropic low geopotential height anomaly is formed, which in

turn enhances the Aleutian low. The roles of meridional SST gradient and atmospheric transient eddy vorticity forcing in the impact of midlatitude SST anomalies on the atmosphere are emphasized.

Therefore, previous studies provide a clue that oceanic thermal conditions can affect the atmosphere in two possible ways: direct thermal forcing by diabatic heating and indirect thermal and dynamical forcing by atmospheric transient eddies. The diabatic heating in most of the midlatitudes is quite weak as compared to that in the tropics, and confined in the lower troposphere because of stable atmospheric stratification. On the other hand, by the steady heating flux on the air–sea interface, the oceanic frontal zone produces a large low-level atmospheric baroclinicity, leading to atmospheric synoptic transient eddy activities. The atmospheric transient eddies can redistribute heat and momentum in the middle to upper troposphere, effectively influencing the large-scale atmospheric circulations. However, the dynamical processes and relative role of the midlatitude ocean’s impact in the formation of midlatitude atmospheric anomalies are still unclear.

This study aims to understand the role of midlatitude oceanic thermal condition in the formation of PDO-related winter atmospheric anomaly structure over North Pacific, by using an atmospheric model with prescribed SST. Specifically, the study identifies the relative contributions of diabatic heating forcing and transient eddies’ thermal and dynamical forcing by quantitative analyses. The rest of the paper is organized as follows. Section 2 describes the data and model experiments. Section 3 represents the structure of the anomalous atmospheric circulation over the midlatitude North Pacific during the PDO warm phase in both observation and Global Ocean and Global Atmosphere (GOGA) experiment. The effects of SST anomalies and meridional SST gradient anomalies in the midlatitude North Pacific are examined in sections 4 and 5, respectively. Dynamical processes and relative role of SST anomalies in the midlatitude North Pacific in the formation and maintenance of the atmospheric anomalies are investigated in section 6. The final section is devoted to conclusions and discussion.

2. Data and model experiments

In this study, the observed monthly-mean atmospheric variables, including geopotential height, wind velocity, and air temperature at 12 standard pressure levels from 1000 to 100 hPa, are taken from NCEP–NCAR monthly reanalysis data (Kalnay et al. 1996). The observed global SST data are taken from Atmospheric Model Intercomparison Project II (AMIP II) (Gates 1992; Kanamitsu et al. 2002).

FIG. 1. (a) Regressed wintertime (DJF) global SST anomalies (K) upon the standardized PDO index in observation (also used in the GOGA run). (b) Time series (blue bar) of the North Pacific SST anomalies used in the GOGA_smth run projected onto the regressed SST anomalies in the red box region of (a), in comparison with the time series of PDO index (red line), for 1948–2010. The dots in (a) indicate the regions exceeding 90% confidence level with the nonparameter random-phase test.

The PDO index is represented by the leading principal component of the wintertime midlatitude North Pacific SST anomalies (20° – 70° N) (Mantua et al. 1997; Zhang et al. 1997). Through linear regression upon the PDO index, the spatial patterns of the wintertime oceanic and atmospheric anomalies during the PDO warm phase are obtained, respectively.

The atmospheric GCM used in this study is GFDL AM2.1 model developed by Geophysical Fluid Dynamics Laboratory (GFDL) with a finite-volume dynamical core. The latitude–longitude horizontal grid is the staggered Arakawa B grid with a resolution of 2° latitude $\times$ 2.5° longitude. In the vertical, the model has 24 levels with the lowest model level about 30 m above the surface (Anderson et al. 2004). To distinguish the impact of the midlatitude North Pacific SST anomalies on the atmosphere, three experiments with the atmospheric GCM are conducted, a control run (GOGA in short) and two sensitivity runs, which are named xNPOGA (global ocean with North Pacific Ocean absent and global atmosphere) and GOGA_smth (global ocean and global atmosphere with smoothed North Pacific), respectively. In the GOGA run, long-term observed global SSTs taken from AMIP II are prescribed in the GCM as the boundary forcing. In the first sensitivity run (i.e., the xNPOGA run), the SSTs used to force the atmosphere are the same as in the GOGA run, except for those in the midlatitude North Pacific (shown with the red box in Fig. 1a) where only the climatological SSTs are prescribed. This fixation leads to a small SST discontinuity around 20° N, and has little influence on atmospheric responses. In the second sensitivity run (i.e., the GOGA_smth run), the SSTs used to force the atmosphere are also the same as in the GOGA run, except for those in the midlatitude North Pacific where the large-scale SST variabilities are kept but their meridional gradients are greatly smoothed by

applying a 3-point smoother 1000 times to the North Pacific SST anomalies north of 20° N. The 3-point smoother can effectively reduce small-scale or frontal-scale SST gradients. Unlike in the xNPOGA run, this smoothing does not bring large SST discontinuity around 20° N. To evaluate the differences of the SST variabilities associated with PDO before and after smoothing, the winter SST anomalies used in the GOGA_smth run are projected onto the spatial pattern of the observed (also GOGA's) PDO-related SST anomalies in the midlatitude North Pacific (shown with the red box in Fig. 1a). As shown in Fig. 1b, the decadal variabilities of the smoothed SST (blue bar) are basically unchanged (Figs. 7a,c), as the correlation coefficient between the time series of PDO index and the projected time series for the North Pacific SST anomalies used in the GOGA_smth is 0.989. This indicates that the GOGA_smth run keeps the large-scale SST variabilities but considerably removes the oceanic front variabilities in the North Pacific.

All the three runs are integrated from 1 September 1947 to 31 December 2010, wherein the first 4 months (September–December 1947) of model integrations are removed as the spinup time, and only the outputs of remaining 63 years (January 1948–December 2010, the same period for observation) are used for analysis. The wintertime atmospheric anomalies are only investigated in this study, and winter here is defined as the 3-month average of December–February (DJF). The synoptic eddies are extracted from the daily model outputs through the 2–8-day bandpass Lanczos filtering (Duchon 1979). The atmospheric baroclinicity is represented by Eady growth rate σ_{BI} at 850 hPa, which can be calculated by the formula $\sigma_{BI} = 0.31f|\partial\mathbf{V}/\partial z|/N$ (Vallis 2006).

The statistical regression method is used to identify the PDO-related atmospheric anomalies, which are measured by the regressions of atmospheric anomalies

upon the standardized PDO index. As the Student's t test depends on an accurate estimation of degrees of freedom, we choose to use a nonparameter method to test the significance of regression. This method was developed by Ebisuzaki (1997) and is usually referred to as the random-phase test (Wu et al. 2016). The basic calculating procedure for this test is as follows. Assume that A and B are two time series, and $r(A, B)$ is their regression. The statistical robustness of regression can be simply tested with the following two steps. First, the time series A is reconstructed N times randomly, but all of the reconstructed time series with random temporal phases have the same power spectrum with A , through a discrete Fourier analysis. Second, we obtain N reconstructed regressions by regressing B upon every reconstructed time series independently. If the magnitude of $r(A, B)$ exceeds a percentage (say, 90%) of the reconstructed regressions, then we say that $r(A, B)$ passes confidence level at this percentage. In this study, N is set to be 5000 to ensure the robustness of significance test, and the confidence level is set to be 90%.

3. Observed and simulated PDO-related atmospheric anomalies

The PDO is the principal signature of SST variabilities in the North Pacific (Fig. 1) and a significant cold-to-warm phase PDO transition occurred in the winter of 1976/77 (Nitta and Yamada 1989; Miller et al. 1994; Francis and Hare 1997; Mantua et al. 1997), as shown in Fig. 1b (red line). During the PDO warm phase, an El Niño-like SST warming locates in the central-eastern tropical Pacific, while the SST anomalies exhibit a cooling in the central North Pacific and a warming along the west coast of North American continent (Fig. 1a). With linear regressions upon the standardized PDO index, spatial patterns of the observed and GOGA-simulated winter-time atmospheric anomalies associated with PDO are shown in Figs. 2 and 3.

Corresponding to the PDO warm phase, a similar basin-scale cooling is observed in the lower-level (850 hPa) air temperature anomalies over the midlatitude North Pacific, and an anomalous low 850-hPa geopotential height is found north of the air temperature cooling center, implying an enhanced Aleutian low, as shown in Fig. 2a. At 250 hPa (Fig. 2b), the air temperature is anomalously warm, but the geopotential height remains anomalously low, with amplitude much larger than that at 850 hPa. Therefore, the negative geopotential height anomalies are characterized by an equivalent barotropic vertical structure that is clearly confirmed from the latitude–altitude cross section along the 140°E–120°W averaged longitude shown in Fig. 3a. The geopotential height is consistently

lower than normal throughout the whole troposphere with its minimum center at 300 hPa around 45°N. Following the hydrostatic relation, the air temperature anomalies (also Fig. 3a) are colder than normal in the lower troposphere but warmer in the upper troposphere, consistent with their horizontal distributions in the lower and upper troposphere (Figs. 2a,b). Such a vertical structure of decadal atmospheric anomalies over a cooling SST in the midlatitude North Pacific is called the equivalent barotropic cold/trough structure (Fang and Yang 2016). Resultantly, increased (decreased) westerly anomalies appear in the southern (northern) flank of the negative geopotential height anomalies, that is, south (north) of 45°N (Figs. 2e,f), and their vertical cross section also exhibits an equivalent barotropic structure (Fig. 3b). The westerly jet is thus greatly enhanced and slightly southward shifted.

Basically, the GOGA run with prescribed observed long-term global SST qualitatively reproduces well the observed PDO-related atmospheric anomalies over the midlatitude North Pacific. Simulated spatial patterns (either horizontal or vertical distributions) of the anomalous geopotential height (Figs. 2c,d and 3c) and zonal wind (Figs. 2g,h and 3d) regressed upon the PDO index are quite similar to the observed, even though their locations are slightly different and their amplitudes are slightly weaker. Specifically, the GOGA run successfully simulated those features of atmospheric responses which are characterized by the equivalent barotropic cold/trough structure (Fig. 3c) and by the enhanced westerly jet (Fig. 3d). The consistency between the observed and GOGA-simulated PDO-related atmospheric anomalies over the North Pacific allows us in the following sections to identify the relative role of local versus remote SST variabilities in those atmospheric anomalies.

4. Role of the midlatitude North Pacific SST variabilities

As reviewed by Newman et al. (2016), the PDO-related midlatitude atmospheric anomalies are also closely related to the tropical SST anomalies. To what extent and in what way the midlatitude SST variabilities can affect the atmospheric circulation is still unclear. The role of the midlatitude SST anomalies in the North Pacific atmospheric anomalies can be identified from the first sensitivity run (xNPOGA) in which the midlatitude North Pacific SST variabilities are removed. Generally, the typical spatial patterns of PDO-related atmospheric anomalies as seen from the GOGA run are reproduced in the xNPOGA run. As shown in Fig. 4a, during the PDO warm phase, at 850 hPa, the geopotential height exhibits an anomalous low over the northeastern North Pacific, while the air temperature features an anomalous

FIG. 2. Regressions of wintertime (a),(c) 850-hPa geopotential height (shaded; gpm) and air temperature (contour; K), (b),(d) 250-hPa geopotential height (shaded; gpm) and air temperature (contour; K), (e),(g) 850-hPa zonal wind (shaded; m s^{-1}), and (f),(h) 250-hPa zonal wind (shaded; m s^{-1}) anomalies upon the standardized PDO index, for observation in (a), (b), (e), and (f) and the GOGA run in (c), (d), (g), and (h). Observed climatological zonal wind speeds (contour; m s^{-1}) are shown at 850 and 250 hPa in (e) and (f), respectively. The dots indicate the regions exceeding 90% confidence level with the nonparameter random-phase test.

FIG. 3. Latitude–altitude sections averaged between 140°E and 120°W of regressions of wintertime (a),(c) geopotential height (shaded; gpm) and air temperature (contour; K), and (b),(d) zonal wind (shaded; m s^{-1}) on the standardized PDO index, for (top) observation and (bottom) the GOGA run. Observed and GOGA-simulated climatological zonal wind speeds (contour; m s^{-1}) are shown in (b) and (d), respectively. The dots indicate the regions exceeding 90% confidence level with the nonparameter random-phase test.

cooling over the western-to-middle North Pacific; at 250 hPa (Fig. 4b), the anomalous geopotential low over the northeastern North Pacific is retained, while the cooling temperature anomalies turn to be anomalous warming. Accordingly, the enhanced westerly winds appear in the exit region of the westerly jet in both the lower and upper troposphere (Figs. 4e,f). The geopotential height anomalies also display an equivalent barotropic vertical structure (Fig. 5a) with maximal amplitude at 300 hPa, as in the GOGA run, despite their smaller amplitudes and slight northward shifting. The westerly jet is also accelerated throughout the whole layer as in the GOGA run but with a weaker strength

(Fig. 5b). These results indicate that the removal of midlatitude North Pacific SST variabilities only weakens the amplitudes of the PDO-related atmospheric anomalies over North Pacific, without largely altering their spatial patterns.

To what degree the amplitude of atmospheric response is reduced by the removal of the midlatitude North Pacific SST variabilities can be further identified by calculating the regressions of differences between the GOGA and xNPOGA runs upon the PDO standardized index. As shown in Figs. 4c and 4d, the cooling SST anomalies in the midlatitude North Pacific tend to induce atmospheric anomalies with a cold/trough structure in the

FIG. 4. Regressions of xNPOGA-simulated wintertime (a) 850-hPa geopotential height (shaded; gpm) and air temperature (contour; K), (b) 250-hPa geopotential height (shaded; gpm) and air temperature (contour; K), (e) 850-hPa zonal wind (shaded; m s^{-1}), and (f) 250-hPa zonal wind (shaded; m s^{-1}) anomalies upon the standardized PDO index. (c), (d), (g), (h) As in (a), (b), (e), and (f), respectively, but for the regressions of differences between GOGA and xNPOGA (GOGA – xNPOGA). The xNPOGA-simulated climatological zonal wind speeds (contour; m s^{-1}) at 850 and 250 hPa are shown in (e), (g) and in (f), (h), respectively. The dots indicate the regions exceeding 90% confidence level with the nonparameter random-phase test.

FIG. 5. Latitude–altitude sections averaged between 140°E–120°W of regressions of xNPOGA-simulated wintertime (a) geopotential height (shaded; gpm) and air temperature (contour; K) and (b) zonal wind (shaded; m s^{-1}) on the standardized PDO index. (c),(d) As in (a) and (b), respectively, but for the regressions of differences between GOGA and xNPOGA (GOGA–xNPOGA). The xNPOGA-simulated climatological zonal wind speeds (contours; m s^{-1}) are shown in (b) and (d). The dots indicate the regions exceeding 90% confidence level with the nonparameter random-phase test.

western basin, consistent with Wu et al. (2017). Previous observational analysis also demonstrated that such a cold (warm) SST anomaly tends to induce an equivalent barotropic low (high), but the anomalous trough (ridge) is more deepened than our simulation and slightly extends northward and downstream (Wen et al. 2010; Révelard et al. 2018), which may be attributed to the air–sea coupling feedback. The anomalous geopotential low lies west of 180° at both 850 and 250 hPa, with a negative air temperature anomaly locating at lower levels but a positive temperature anomaly locating at upper levels (Figs. 4c,d). Accordingly, an enhanced westerly wind appears in the

southern flank of the jet, with weakened wind in the northern flank (Figs. 4g,h). The cold/trough structure and the whole-layer enhanced westerly wind induced by the midlatitude North Pacific SST variabilities can be seen again in the altitude–latitude sections (Figs. 5c,d), as in the GOGA run (Figs. 3c,d), but the amplitude is greatly reduced. A simplified modeling study presented similar evidence that PDO-like cold SST anomalies in the midlatitude North Pacific can trigger a significant dipole mode of westerly jet response (Tao et al. 2019).

To give a quantitative estimate of the relative roles of local North Pacific SST forcing versus remote SST

FIG. 6. The ratios of $(\text{GOGA} - \text{xNPOGA})/\text{GOGA}$ (red dashed line) and $(\text{GOGA} - \text{GOGA_smth})/\text{GOGA}$ (blue dashed line) for PDO-regressed geopotential height anomalies averaged over ranges from 1000 to 100 hPa in the vertical direction, and from 30° to 40°N for $(\text{GOGA} - \text{xNPOGA})/\text{GOGA}$ and from 40° to 50°N for $(\text{GOGA} - \text{GOGA_smth})/\text{GOGA}$ in the meridional direction, plotted as a function of the years since 1948 for regression. The gray line is a reference line that indicates 33% on the y axis.

forcing in the formation of the PDO-related cold/trough structure, we define and calculate a ratio of the North Pacific SST anomaly-induced PDO-regressed geopotential height anomalies over those induced by the global SST anomalies, that is, $(\text{GOGA} - \text{xNPOGA})/\text{GOGA}$ for PDO-regressed geopotential height anomalies. In terms of the vertical distributions of regressed geopotential height anomalies (Figs. 3c and 5c), the ratio is averaged over ranges from 1000 to 100 hPa in vertical direction and from 30° to 40°N in the meridional direction, covering most of the regions with negative anomalies exceeding the 90% significance level. Figure 6 shows the ratio, which is plotted as a function of the years since 1948 for regression. It can be seen that the ratio of $(\text{GOGA} - \text{xNPOGA})/\text{GOGA}$ (red dashed line) becomes steady, being roughly 33%, as the years for regression exceed around 47 years since 1948. In other words, the midlatitude North Pacific SST variabilities contribute roughly by one-third to the PDO-related atmospheric anomalies over North Pacific.

How the local midlatitude SST anomalies can affect the overlying atmosphere is still an open question. The typical cold/trough structure in which an equivalent barotropic geopotential low is lying over a cooling SST surface in the midlatitudes is different from that in the tropics where the atmosphere is thermally driven and the atmospheric response to SST anomalies is usually characterized by a baroclinic structure (Matsuno 1966; Gill 1980). Different from the thermally driven mechanism in the tropics, the cause and maintenance of the unique equivalent barotropic cold/trough structure in the midlatitudes cannot be just interpreted by the diabatic heating. A previous study by Fang and Yang (2016) found the role of the atmospheric transient eddy in unstable midlatitude air–sea interaction in which the oceanic front zone defined by large meridional

SST gradient is considered as the key region. The SST gradient can exert influence on the low-level atmospheric baroclinicity, and then on the transient eddy activities on the upper troposphere (Nakamura and Shimpo 2004; Nakamura et al. 2008, 2004; Wang et al. 2017). In the following section, we further examine the role of the SST meridional gradient in the formation of the PDO-related atmospheric anomalies in the midlatitudes.

5. Role of the meridional gradient of midlatitude North Pacific SST variabilities

As shown in Figs. 7a and 7b, along with the anomalous basin-scale cooling in the PDO warm phase, the meridional SST gradient is strengthening (weakening) in the southern (northern) flank of the cooling SST anomaly. To isolate the role of the SST gradient in the formation of the atmospheric anomalies over the midlatitude North Pacific, the second sensitivity (i.e., GOGA_smth) run was conducted in which the midlatitude North Pacific SST variabilities (anomalies) used to force the GCM are meridionally smoothed. Compared to the original PDO-related SST anomalies used in the GOGA run (Figs. 7a,b), the smoothing does not affect too much the large-scale structure of the SST anomalies and their gradients, but their amplitudes are reduced to a different degree (Figs. 7c,d). As shown in Figs. 7e and 7f, the cooling SST anomalies are reduced by less than one-third, while the anomalous SST gradients are weakened by more than half. Furthermore, as mentioned at section 2, the PDO-related large-scale variabilities of the smoothed SST are basically unchanged, as compared to the original (Fig. 1b). Thus, the 3-point smoothing does not affect too much large-scale SST anomalies, but considerably reduces frontal-scale meridional SST gradient (i.e., oceanic front) variabilities. Thus, the GOGA_smth run can be considered as the GOGA run but with extremely weak frontal-scale meridional SST gradient (oceanic front) anomalies in the midlatitude North Pacific.

It is interesting to find that the PDO-regressed atmospheric anomalies in the GOGA_smth run (Fig. 8) are generally consistent with those in the xNPOGA run (Fig. 4). During the PDO warm phase, the anomalous geopotential low arises again (Figs. 8a,b), similar to that in the xNPOGA run, but locates more southward (Figs. 4a,b). Correspondingly, the anomalous air temperature is negative in the lower troposphere while positive in the upper troposphere (Figs. 8a,b). The westerly winds are strengthened in the center and downstream of climatological maximum (Figs. 8e,f), although these anomalies have a slight southwestward

FIG. 7. Regressions of wintertime (a),(c) SST (K) and (b),(d) SST gradient $[K (110 \text{ km})^{-1}]$ anomalies upon the standardized PDO index used in the (a),(b) GOGA and (c),(d) GOGA_smth runs. (e),(f) The regressions of differences between GOGA and GOGA_smth (GOGA – GOGA_smth), respectively, in which the contours are exactly the same as the shades in (a) and (b), respectively. The dots indicate the regions exceeding 90% confidence level with the nonparameter random-phase test.

shift as compared to those in the xNPOGA run (Figs. 4e,f). Furthermore, as shown in Figs. 9a and 9b, both the anomalous geopotential low and the strengthened zonal wind demonstrate again equivalent barotropic vertical structures that are quite similar to those in the xNPOGA run, not only in their distributions but their amplitudes. The similarity of atmospheric responses between the GOGA_smth and xNPOGA runs indicates that the respective removal of the North Pacific SST variabilities and their frontal-scale meridional gradients exerts an equivalent effect on the atmosphere, especially in the upper troposphere.

Similarly, by looking at the regression of difference between the GOGA and GOGA_smth runs upon the standardized PDO index, the effect of the frontal-scale

SST gradient variabilities can be seen more clearly. As shown in Figs. 8c and 8d, the meridional SST gradient anomalies in North Pacific tend to considerably induce an anomalous geopotential low over the northeastern North Pacific. Besides, they also strengthen the western part of the GOGA's cold/trough structure (Figs. 2c,d), and the anomalous geopotential low in the western basin is similar to that induced by the North Pacific SST anomalies (Figs. 4c,d). In the western basin, the westerly wind is enhanced (decreased) in the southern (northern) flank of the jet, while it is enhanced east of 180° , downstream of the jet (Figs. 8g,h). The latitude–altitude distributions exhibit that the SST gradient anomalies-induced geopotential height and zonal wind anomalies are equivalent barotropic (Figs. 9c,d). Compared to

FIG. 8. As in Fig. 4, but for the GOGA_smth run and for the regressions of differences between GOGA and GOGA_smth ($GOGA - GOGA_{smth}$).

those induced by the North Pacific SST anomalies (Figs. 5c,d), the atmospheric anomalies by the SST gradient anomalies (Figs. 9c,d) have the comparable amplitudes despite their northward shifts in location.

A quantitative relative contribution of the SST gradient anomalies to the total atmospheric response is similarly estimated, by calculating the ratio of

$(GOGA - GOGA_{smth})/GOGA$ for PDO-regressed geopotential height anomalies averaged over the ranges from 1000 to 100 hPa in vertical direction and from 40° to 50°N in meridional direction, which cover most of the regions with negative geopotential height anomalies exceeding 90% significance level. As shown in Fig. 6, the relative contribution to the geopotential

FIG. 9. As in Fig. 5, but for the GOGA_smth run and for the regressions of differences between GOGA and GOGA_smth ($GOGA - GOGA_{smth}$).

height anomalies by the frontal-scale meridional SST gradient anomalies (blue dashed line) merges with that by the North Pacific SST anomalies (red dashed line), as the years for regression exceed 60 years since 1948, although the meridional ranges for averaging geopotential height anomalies have a slight difference. Therefore, during the PDO warm phase, the frontal-scale SST gradient anomalies also contribute roughly by one-third to the anomalous cold/trough structure as well as the enhanced westerly wind in the GOGA run, comparable to what the North Pacific SST anomalies do. One possible deduction from the results is that the SST anomalies in the midlatitude North Pacific affect the atmosphere mainly through the anomalous meridional SST gradients. In the following section, we further examine possible mechanisms

and processes by which the midlatitude SST anomalies affect the atmosphere.

6. Ways the midlatitude North Pacific SST variabilities affect the atmosphere

a. Forcing sources of midlatitude seasonal mean atmospheric state

As mentioned in section 4, the midlatitude atmosphere is characterized by abundant synoptic transient eddy activities due to atmospheric baroclinicity. The midlatitude SST anomaly may affect the atmosphere by altering low-level atmospheric baroclinicity. In the GOGA run, as shown in Figs. 10a and 10b, the climatological meridional air temperature gradient is large over the Kuroshio–Oyashio Extension (KOE) regions,

FIG. 10. Regressions of wintertime (a) 850-hPa meridional temperature gradient [shaded; K (110 km)⁻¹] and (b) 850-hPa Eady growth rate (shaded; day⁻¹) anomalies in the GOGA run upon the standardized PDO index. Their respective climatologies (contours) are shown in (a) and (b). (c),(d) The corresponding regressions of differences between GOGA and xNPOGA (GOGA – xNPOGA, and (e),(f) the differences between GOGA and GOGA_smth (GOGA – GOGA_smth). The contours in (c),(e) and in (d),(f) are exactly the same as the shading in (a) and (b), respectively. The dots indicate the regions exceeding 90% confidence level with the nonparameter random-phase test.

coinciding with large climatological atmospheric baroclinicity represented by Eady growth rate at 850 hPa. During the PDO warm phase, both the air temperature gradient and atmospheric baroclinicity are enhanced downstream of their climatological maximums, with two branches west of 180° (Figs. 10a,b), corresponding to the enhanced SST gradient anomalies (Fig. 7b). For both GOGA minus xNPOGA and GOGA minus GOGA_smth, the SST and its frontal-scale meridional gradient anomalies tend to strengthen the atmospheric baroclinicity between 20° and 40°N over the North Pacific (Figs. 10c–f), when the PDO is in its positive phase. The former mainly strengthens the southern

branch while the latter mainly strengthens the northern branch, which is consistent with recent observational results that the basin-scale SST gradient anomalies in the central-to-eastern North Pacific could have an impact on the atmosphere comparable to those in the KOE region (Wang et al. 2017; Révelard et al. 2018). Gan and Wu (2013) utilized lagged maximum covariance analysis of observed wintertime storm tracks and SSTs and proposed that preceding cold SST anomalies in the western-central North Pacific are associated with the equatorward shift of atmospheric baroclinicity and storm track, which is consistent with our simulation results. Therefore, a strengthened SST gradient caused

by the SST anomaly generates a strengthened atmospheric temperature gradient as well as an enhanced atmospheric baroclinicity, favoring the generation of more synoptic baroclinic Rossby waves, namely transient eddies.

The synoptic transient eddy can redistribute heat and momentum efficiently in the upper troposphere, which is essential in formation and maintenance of the midlatitude eddy-driven jet (Williams 1979; Panetta and Held 1988; Panetta 1993). Following Fang and Yang (2016), the role of the synoptic transient eddy in the midlatitude seasonal-mean atmospheric state can be determined by the quasigeostrophic potential vorticity (QGPV) equation:

$$\left(\frac{\partial}{\partial t} + \bar{\mathbf{V}}_h \cdot \nabla\right) \left[\frac{1}{f} \nabla^2 \bar{\Phi} + f + \frac{\partial}{\partial p} \left(\frac{f}{\sigma_1} \frac{\partial \bar{\Phi}}{\partial p} \right) \right] = -f \frac{\partial}{\partial p} \left(\frac{\alpha}{\sigma_1} \frac{\bar{Q}_d}{\bar{T}} \right) - f \frac{\partial}{\partial p} \left(\frac{\alpha}{\sigma_1} \frac{\bar{Q}_{\text{eddy}}}{\bar{T}} \right) + \bar{F}_{\text{eddy}}, \quad (1)$$

where the overbar denotes the seasonal mean, Φ is the geopotential height, α is the reciprocal of density, σ_1 is the static stability parameter ($\sigma_1 = -\alpha \partial \ln \theta / \partial p$), and T is the temperature. Also, $\bar{Q}_d$ is the seasonal-mean diabatic heating, calculated with four components of model outputs (vertical diffusion, latent heat of condensation, longwave and shortwave radiation heating/cooling); $\bar{Q}_{\text{eddy}}$ and $\bar{F}_{\text{eddy}}$ are two transient eddy forcing terms, namely the seasonal-mean transient eddy heating and vorticity forcing term, respectively. They are basically determined by the convergence of transient eddy heat and vorticity fluxes, respectively, which can be expressed as $\bar{Q}_{\text{eddy}} = -\nabla \cdot \mathbf{V}'_h T'$ and $\bar{F}_{\text{eddy}} = -\nabla \cdot \mathbf{V}'_h \varsigma'$, where the prime denotes the deviations from seasonal mean, ς is the relative vorticity, R is the gas constant, and C_p is the specific heat at constant pressure. Since the low-level atmospheric baroclinicity is closely related to high-frequency transient eddy activity (Simmons and Hoskins 1978; Hoskins and James 2014), the role of synoptic-scale (2–8-day filtered) transient eddies is only discussed in this study.

On the right-hand side of Eq. (1), there are three forcing terms that can generate atmospheric potential vorticity (PV): diabatic heating (F1), transient eddy heating forcing (F2), and transient eddy vorticity forcing (F3). The midlatitude seasonal-mean atmospheric state is driven by both thermal and dynamical forcing. Thus, the midlatitude SST anomalies can affect the atmosphere through two ways: direct thermal forcing by diabatic forcing and indirect thermal and dynamical forcing by atmospheric transient eddy activities. These forcing terms that are associated with PDO are

calculated with daily model output data and presented in Figs. 11 and 12. The vertical structure of diabatic heating and transient eddy forcing terms can be identified by their zonal averages over the basin-scale midlatitude North Pacific. For the GOGA run, climatologically (shown with contours in Figs. 11a–c), the diabatic heating is generally confined to the lower troposphere over the midlatitude North Pacific (Fig. 11a), while the transient eddy heating has positive centers in the mid- to upper troposphere north of 32°N and negative centers in the mid- to lower troposphere south of 45°N, forming a baroclinic structure between 32° and 45°N (Fig. 11b). Interestingly, the transient eddy vorticity forcing is characterized by an equivalent barotropic meridional dipole structure in climatology, with larger positive (negative) centers north (south) of 35°N in the upper troposphere (Fig. 11c).

During the PDO warm phase, diabatic heating anomalies are basically in phase with its climatology, mainly confined to the lower troposphere, although some can penetrate into high-level atmosphere (Fig. 11a). However, through transient eddy activities, the transient eddy forcing can influence the mid- to upper troposphere. The transient eddy heating is enhanced and shifts southward in the whole troposphere at around 40°N, especially in the mid- to lower troposphere (Fig. 11b). The transient eddy vorticity forcing anomaly demonstrates a vertical structure similar to its climatology, but shifts southward (Fig. 11c). Overall, the transient eddy forcing is intensified and shifts southward, corresponding to the atmospheric baroclinicity anomalies (Fig. 10b). To compare with the anomalous transient eddy vorticity forcing term F3 (Fig. 12c), we further calculate the former two forcing terms (F1 and F2) that are proportional to the vertical gradient of diabatic heating anomalies and transient eddy heating anomalies, respectively. Different from the equivalent barotropic structure of F3 (Fig. 12c), both F1 and F2 display a baroclinic structure with a negative (positive) anomaly above (below) its maximal heating anomaly in the vertical direction, respectively, partly canceling each other out at around 40°N (Figs. 12a,b).

The effects of the midlatitude North Pacific SST anomalies and their SST gradients on the diabatic heating, transient eddy heating, and transient eddy vorticity forcing are also shown. Basically, the anomalies of those forcing terms induced by the SST anomalies and by the SST gradient anomalies are similar, and all in phase with the anomalies from the GOGA run, but their amplitudes are relatively weak (Figs. 11 and 12). The transient eddy transport anomalies induced by the SST anomalies slightly shift southward (Figs. 11e,f), but are shifted northward by the SST gradient anomalies

FIG. 11. Latitude–altitude sections averaged between 140°E and 120°W of regressions of wintertime (a) diabatic heating (shaded; K day^{-1}), (b) transient eddy heating (shaded; K day^{-1}), and (c) transient eddy vorticity forcing (shaded; 10^{-11} s^{-2}) anomalies in the GOGA run upon the standardized PDO index. Their respective climatologies (contours) are shown in (a)–(c). (d)–(f) The corresponding regressions of differences between GOGA and xNPOGA (GOGA – xNPOGA), and (g)–(i) the differences between GOGA and GOGA_smth (GOGA – GOGA_smth). The contours in (d)–(f) and in (g)–(i) are exactly the same as the shades in (a)–(c), respectively. The dots indicate the regions exceeding 90% confidence level with the nonparameter random-phase test.

(Figs. 11f,i), corresponding to the shifts of the atmospheric low-level temperature gradient and baroclinicity, respectively (Figs. 10c–f). As in the GOGA run, F1 and F2 induced by the SST anomalies and by the SST gradient anomalies are both baroclinic, and partly cancel out each other (Figs. 12d,e,g,h), and only

F3 is equivalent barotropic in the vertical direction (Figs. 12f,i).

b. Relative contributions of different forcing sources

To quantitatively analyze the relative contributions of anomalous diabatic heating, transient eddy heating, and

FIG. 12. As in Fig. 11, but for the forcing terms (10^{-11} s^{-2}) induced by (a),(d),(g) diabatic heating anomalies (F1), (b),(e),(h) transient eddy heating anomalies (F2), and (c),(f),(i) transient eddy vorticity forcing anomalies (F3).

transient eddy vorticity forcing to the winter mean atmospheric anomalies, the tendency of geopotential height anomalies induced by those forcing anomalies is determined by the following relation (Fang and Yang 2016):

$$\left[\frac{1}{f} \nabla^2 + f \frac{\partial}{\partial p} \left(\frac{1}{\bar{\sigma}_1} \frac{\partial}{\partial p} \right) \right] \left(\frac{\partial \Delta \Phi}{\partial t} \right) \propto -f \frac{\partial}{\partial p} \left(\frac{\alpha}{\bar{\sigma}_1} \frac{\Delta \bar{Q}_d}{\bar{T}} \right) - f \frac{\partial}{\partial p} \left(\frac{\alpha}{\bar{\sigma}_1} \frac{\Delta \bar{Q}_{\text{eddy}}}{\bar{T}} \right) + \Delta \bar{F}_{\text{eddy}}, \quad (2)$$

where Δ denotes the anomaly. Given the forcing terms, the tendency of geopotential height anomalies can be numerically solved with the successive overrelaxation (SOR) method. The settings of boundary conditions are important and may disturb the solution as the SOR method is applied. The effect of horizontal boundary condition is small, which can be set as

$$\left(\frac{\partial \Phi}{\partial t} \right)_{2.5^\circ \text{S}} = 0, \quad \frac{\partial}{\partial y} \left(\frac{\partial \Phi}{\partial t} \right)_{90^\circ \text{N}} = 0, \quad (3)$$

in the y direction. A cycling boundary is applied in the x direction. Following Lau and Holopainen (1984), we set the vertical boundary conditions for Q_d and Q_{eddy} as

$$\left[\frac{\partial}{\partial p} \left(\frac{\partial \Phi}{\partial t} \right) \right]_{1000 \text{ hPa}/100 \text{ hPa}} = -\frac{R}{p} Q_{1000 \text{ hPa}/100 \text{ hPa}}, \quad (4)$$

at 1000 and 100 hPa, respectively. For F_{eddy} , they are set as

$$\left[\frac{\partial}{\partial p} \left(\frac{\partial \Phi}{\partial t} \right) \right]_{1000 \text{ hPa}/100 \text{ hPa}} = 0. \quad (5)$$

The vertical boundary conditions may have some effect on the geopotential tendencies in the stratosphere induced by Q_d and Q_{eddy} , which are supposed to vanish there. The geopotential height tendency induced by the diabatic heating, transient eddy heating, and transient eddy vorticity forcing, respectively, is calculated and shown in Fig. 13.

In the GOGA run, the anomalous diabatic heating tends to induce geopotential tendency anomaly with a tripole structure in vertical direction at around 40°N (Fig. 13a), while the geopotential tendency induced by the anomalous transient eddy heating features a dipole in the vertical (Fig. 13b). Therefore, the geopotential responses to both thermal forcing terms are baroclinic. Nevertheless, the geopotential tendency induced by the anomalous transient eddy vorticity forcing is equivalent barotropic, characterized by a negative geopotential height tendency anomaly at around 40°N (Fig. 13c), corresponding to the anomalous geopotential low in Fig. 3c. Therefore, the transient eddy vorticity forcing is more important than other two thermal forcing terms in the formation of the cold/trough structure, as found by Fang and Yang (2016). As for the GOGA minus xNPOGA and GOGA minus GOGA_smth cases, all of the geopotential tendencies induced by diabatic heating forcing and transient eddy heating forcing are also baroclinic (Figs. 13d,e,g,h). Only the geopotential tendency induced by transient vorticity forcing is equivalent barotropic in both GOGA – xNPOGA and GOGA – GOGA_smth, and has a southward shift in the former case but a northward shift in the latter case, corresponding to the anomalous geopotential height in Figs. 5c and 9c, respectively.

In terms of the geostrophic relationship, the zonal wind tendency can be derived from the geopotential height tendency. The zonal wind is increased in the southern flank of the geopotential low, while decreased in its northern flank. Again, the zonal wind tendency induced only by the transient eddy vorticity forcing is equivalent barotropic (Fig. 14c), while that induced by either the diabatic heating or the transient eddy heating

is baroclinic in the GOGA run (Figs. 14a,b). As for the GOGA minus xNPOGA and GOGA minus GOGA_smth cases, the zonal wind tendency induced by those terms shows similar structure but with weaker amplitude, and southward shift in the GOGA minus xNPOGA case and northward shift in the GOGA minus GOGA_smth case (Figs. 14d–i). The equivalent barotropic positive zonal wind tendency induced by the transient eddy vorticity forcing in the GOGA, GOGA minus xNPOGA, and GOGA minus GOGA_smth cases corresponds to the enhanced westerly wind, as shown in Figs. 3d, 5d, and 9d, respectively.

Therefore, the above results indicate that the transient eddy vorticity forcing is a key dynamical process for the midlatitude North Pacific SST and its gradient variabilities to affect the atmosphere. Moreover, the contrast among different runs indicates that large-scale SST variabilities in the midlatitude North Pacific mainly affect the diabatic heating, while the similarity among different runs implies again that the frontal-scale SST meridional gradient is the key for the midlatitude North Pacific SST anomalies to affect the atmosphere in which the transient eddy vorticity forcing is a key player.

7. Conclusions and discussion

The PDO is the dominant decadal-to-interdecadal climate variability in the North Pacific air–sea system. During the PDO warm phase, the anomalous winter-time air–sea system of the midlatitude North Pacific is characterized by a cold/trough structure in observation; that is, an anomalous equivalent barotropic geopotential low lies upon the negative SST anomaly. Forced by long-term observed global SST in a control run, the typical wintertime cold/trough structure over the midlatitude North Pacific is captured by an atmospheric GCM (GFDL AM2.1), although simulated amplitudes of the anomalous trough and associated strengthened westerly wind are weaker than the observed, which may be attributable to the lack of air–sea coupling. To identify the impact of the midlatitude oceanic thermal condition on the atmosphere, the SST variabilities in the midlatitude North Pacific are removed by simply setting the SST there as climatology in the first sensitivity run. The atmospheric response also shows an equivalent barotropic cold/trough structure in the vertical direction. As for the anomalous amplitude, the lack of the midlatitude North Pacific SST variabilities tends to reduce the atmospheric response by roughly one-third. Therefore, the midlatitude North Pacific SST anomalies make a considerable (approximately one-third) contribution to the PDO-related cold/trough anomalies, although most of those atmospheric anomalies are determined by the SST variabilities

FIG. 13. As in Fig. 11, but for the geopotential tendencies ($10^{-4} \text{ m}^2 \text{ s}^{-3}$) induced by (a),(d),(g) diabatic heating anomalies (F1), (b),(e),(h) transient eddy heating anomalies (F2), and (c),(f),(i) transient eddy vorticity forcing anomalies (F3).

outside of the midlatitude North Pacific, especially by the tropical Pacific SST variabilities.

Different from the thermally driven mechanism induced by deep convection in the tropics, the diabatic heating is mainly confined to the low-level troposphere in the midlatitudes. The typical cold/trough structure cannot be explained by only thermal processes. During the PDO warm phase, the meridional SST gradient is anomalously strengthened in the southern flank of the

cooling SST anomaly. In the second sensitivity run, the frontal-scale meridional SST gradient variabilities in the North Pacific are sharply smoothed, while the large-scale SST variabilities there are kept. In this case, the simulated PDO-related cold/trough anomalies are also reduced by nearly one-third. Therefore, the frontal-scale meridional SST gradient (oceanic front) is essential in the process of the midlatitude SST anomaly's impact on the atmosphere above.

FIG. 14. As in Fig. 13, but for the zonal wind tendencies (10^{-6} m s^{-2}).

Further dynamical diagnoses based on a QGPV equation exhibit that although all the diabatic heating, transient eddy heating, and transient eddy vorticity forcings support the low-level low in the control run, the transient eddy vorticity forcing is the dominant forcing term in the formation and maintenance of the cold/trough anomalies. Compared to the baroclinic structure induced by diabatic heating and transient eddy heating, the geopotential low and strengthened westerly wind induced only by transient eddy vorticity forcing are equivalent barotropic.

Furthermore, the removal of either the midlatitude SST variabilities or the frontal-scale meridional SST gradient variabilities primarily decreases the low-level air temperature gradient as well as the atmospheric baroclinicity. Then the transient eddy activities are weakened, giving rise to a decrease in the transient eddy transport. Such a decrease induces a considerable weakening of the cold/trough structure by decreasing transient eddy vorticity forcing. Therefore, the midlatitude North Pacific SST anomalies play an essential rather than a trivial role

in the winter atmospheric anomalies associated with PDO, in which the frontal-scale meridional SST gradient (oceanic front) and synoptic transient eddy dynamical feedback are key players.

As the transient eddy activities are important in the midlatitude air–sea interaction, finer-resolution models can be better in resolving storm track. Thus, a comparison among different models with diverse resolution is needed. Furthermore, the seasonal-mean relationship between the SST and the atmosphere identified in this study is simultaneous, in terms of AMIP-like experiments. The lead–lag relationships, especially when the SST anomalies lead the atmospheric anomalies, need to be further examined with coupled ocean–atmosphere model experiments.

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