The Characteristic Variability of Boreal Wintertime Atmospheric Circulation in El Niño Events

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ABSTRACT

The influence of the Pacific sea surface temperature (SST) on the characteristic variability of boreal wintertime atmospheric circulation among El Niño events is investigated by carrying out singular value decomposition (SVD) analysis and general circulation model (GCM) experiments. From the SVD analysis of SST and 300-hPa streamfunction in the El Niño winter, the first mode associated with the strength of the El Niño mode itself has a positive trend in its time series and the second mode associated with the North Pacific SST variation includes substantial event-to-event variability. The second mode is thus important for explaining the inter–El Niño variability in atmospheric circulation. Four El Niños among 15 El Niño events from 1950 to 1998 have a mainly positively contribution from the first SVD mode, and the other four events are influenced by the second mode as well as the first mode. The atmospheric responses related to the latter four events show the weakening of the Aleutian low and an eastward-shifted pattern of circulation in contrast with the former ones.

Results of ensemble GCM experiments imply that the North Pacific SST leads to differences in the atmospheric circulation among El Niño years. When the SST anomaly exists only in the North Pacific, the atmospheric responses to SST forcing are very weak. However, in the presence of a negative SST anomaly in the North Pacific with a positive SST anomaly in the equatorial eastern Pacific, the atmospheric responses are much stronger than the case with the absence of SST forcing in the North Pacific. When a positive SST anomaly exists in the North Pacific, the responses are weaker. Therefore, the intensity of the atmospheric responses depends partly on the sign of the SST anomaly in the North Pacific.

In addition, the potential predictability using ensemble experiments is calculated to investigate the relative importance of SST forcing with respect to the internal dynamics in atmospheric circulation. The predictability of atmospheric responses is poor for the SST forcing in the North Pacific only, but is very good for the negative SST forcing in the North Pacific and the positive SST forcing in the equatorial eastern Pacific.

1. Introduction

The influence of tropical sea surface temperature (SST) anomalies during El Niño on the wintertime atmospheric circulation has been studied by many authors during the past decade (Grimm and Silva Dias 1995; Held and Kang 1987; Hou 1998; Kang 1996; Li and Nathan 1994, 1997; Ting and Sardesmukh 1993; Trenberth et al. 1998). These studies have shown that the midlatitude circulation anomalies in winter vary each El Niño. Kumar and Hoerling (1997) showed that inter–El Niño variabilities exist in the wintertime midlatitude circulation anomalies and that these variabilities are primarily due to the internal dynamics, although the differences in SST forcing for various El Niño events cannot be ignored. Although variations in midlatitude circulation during El Niño have been studied intensively so far, there remain several unsolved problems regarding inter–El Niño variability in the wintertime midlatitude circulation. Among them are the following issues. 1) Are there particular modes that cause most of the inter–El Niño variability in the midlatitude circulation? 2) What is the role of extratropical Pacific SST anomalies in modulating the atmospheric circulation in an El Niño year?

It is known that the extratropical circulation anomalies are strongly controlled by the tropical heating anomalies, although the midlatitude SST and internal transients also affect the circulation in the extratropics. Held and Kang (1987) indicated that the anomalous subtropical convergence plays a key role in generating the extratropical wave train in the barotropic model. The result of Grimm and Silva Dias (1995) supports the suggestion of Held and Kang (1987) that emphasizes the need for correct specification of the subtropical divergence/convergence anomalies associated with the tropical heat source. Hou (1998) showed that the low-frequency temporal variability in the Hadley circulation is a source of the wave train modulation in the winter extratropics. According to Ting and Sardesmukh (1993),
the extratropical response to tropical forcing was found to be very sensitive to the basic surrounding state and to the longitudinal position of the tropical forcing with respect to the climatological waves. Kang (1996) also suggested that interannual and interdecadal variations of the global-mean surface air temperature are associated with corresponding variations of tropical Pacific SST.

Not only the tropical SST anomalies but also extratropical SST anomalies are shown to be an important source of extratropical circulation anomalies (Namias 1969). Recent studies show that there is significant cooling of SST in the central North Pacific (Nakamura et al. 1997), the Aleutian low tended to be more intense, the Pacific storm track was likely to be shifted to the south (Trenberth and Hurrel 1994), and the atmospheric intraseasonal fluctuation over the North Pacific tended to be weaker than before (Nakamura 1996). It is suggested that such variations in the North Pacific could be caused by the atmosphere–ocean interaction in the midlatitudes as well as tropical forcing (Lau 1997; Nakamura et al. 1997; Bond and Harrison 2000).

The variation of midlatitude SST anomalies both related and unrelated to tropical SST anomalies and its influence on the midlatitude circulation are of interest to many researchers. Observational analyses suggest that SST anomalies in the midlatitudes are initiated by atmospheric fluctuations (Palmer and Sun 1985; Deser and Timlin 1997). That is, the atmospheric circulation plays a crucial role as an “atmospheric bridge” linking SST changes in different parts of the world’s oceans (Lau and Nath 1994). However, the variation of midlatitude SST does not have a simple relation with the tropical SST, but has a characteristic variability. According to the study of Nakamura et al. (1997), the decadal SST variability within the subtropical region is strongly influenced by the Tropics, whereas the variability within the midlatitudes cannot be explained solely by the tropical forcing and may be maintained by local atmosphere–ocean interactions. On the other hand, the GCM experiments of Lau (1997), which attempted to explain the feedback effect of midlatitude SST forcing on the extratropical atmospheric circulation, showed that the tropical Pacific SST fluctuation associated with ENSO episodes produces a strong extratropical response in the model atmosphere, whereas the atmospheric signal associated with midlatitude Pacific SST anomalies is less robust. The atmospheric response to the tropical SST forcing, however, is stronger in the presence of the midlatitude SST anomalies. Although the feedback processes of SST anomalies in the midlatitudes on the atmospheric circulation are as yet unclear, many studies supported that the midlatitude SST anomalies do force significant changes in the atmosphere circulation (Palmer and Sun 1985; Brankovic et al. 1994; Latif and Barnett 1996).

Another factor related to inter–El Niño variation of the atmospheric circulation is the effect of internal dynamics associated with transients (Held et al. 1989; Hoerling and Ting 1994; Peng et al. 1997; Sheng et al. 1998; Peng and Whitaker 1999). Peng et al. (1997) explained that the atmospheric response to SST anomalies in the midlatitudes depends strongly on the GCM climatology, that is, the background meridional flow. Mean flow changes over the midlatitude ocean are related to the variation in high-frequency transients (Branstator 1995). The anomalous transient eddy forcing enhances the influence of the atmospheric response compared to the anomalous heating (Held et al. 1989; Peng and Whitaker 1999). Hoerling and Ting (1994) also suggested that the extratropical vorticity transients, which are influenced by the tropical forcing, are the primary mechanisms in maintaining the stationary wave anomalies over the Pacific–North American (PNA) region observed during ENSO winters. That is, the vorticity flux divergence anomaly in the PNA region is found to be well correlated with the height field, particularly in the upper troposphere. The sensible heat flux divergence, on the other hand, is negatively correlated with the temperature anomaly in the PNA region, so that the transient eddies tend to destroy the temperature anomaly (Sheng et al. 1998).

The present paper suggests possible causes for inter–El Niño variability in the atmospheric circulation and their resulting impact, especially the effect of the SST anomaly in the midlatitudes. To solve the above-mentioned problems, we analyzed the observational data using several statistical methods and carried out model experiments. The datasets and model experiments used in our study are described in section 2. The observational analysis is presented in section 3 and the model analysis results are shown in section 4. The summary and discussion are given in section 5.

2. Data

The observed data used in this study are obtained from the monthly mean reanalysis dataset of the National Centers for Environmental Prediction (NCEP) and the National Oceanic and Atmospheric Administration (NOAA) SST dataset for the 48-winter period from 1950/51 to 1997/98. The resolution of the NCEP reanalysis data is 2.5° in latitude and 2.5° in longitude along with the 12 pressure levels from 1000 to 100 hPa. The observed SST data are those reconstructed recently at NCEP. They have been obtained by applying a new interpolation method, based on spatial patterns from empirical orthogonal functions (EOF) of SST anomalies for the period of 1982–92, to the monthly SST of the Comprehensive Ocean–Atmosphere Data Sets (COADS) for the period of 1950–92. The reconstructed SST fields have lower root-mean-square differences than SST fields derived from the traditional NCEP analysis of optimum interpolation. The horizontal resolution of SST data is 2° in latitude and 2° in longitude. Each

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The winter season consists of the 3-month period from December to February.

The definition of El Niño in this paper follows that of Trenberth (1997). He defined El Niño as the year when 5-month running means of SST anomalies in the Niño-3.4 region (5°N–5°S, 120°–170°W) exceed 0.4°C for 6 months or more. The El Niño winters after the 1950s used in the present study are listed in Table 1.

Figure 1 shows the horizontal distribution of mean SST anomaly and mean 300-hPa streamfunction anomaly in El Niño years for the whole periods mentioned above.

3. Observational features

a. SVD analysis

To reveal the effect of SST forcing on the midlatitude circulation and to find the ocean–atmosphere coupled mode that is responsible for the variability of the wintertime atmospheric circulation among El Niño events, the singular value decomposition (SVD) analysis is performed. The results of SVD analysis between the SST anomaly and streamfunction anomaly at 300 hPa in the Northern Hemisphere winter of El Niño years are shown in Fig. 2. The SVD of the SST anomaly is performed in the Pacific region (about 30°S–60°N, 100°E–80°W), and that of the streamfunction anomaly in the region (about 30°S–90°N, 60°E–0°W). Since the data for only El Niño years were used in this analysis, Fig. 2 does not include the mean field of El Niño anomalies of Fig. 1.

The first SVD mode is associated with the intensity of the typical El Niño pattern itself in Fig. 1. The SST anomaly distribution of the first mode has a seesaw pattern between the tropical Pacific and the central extratropical Pacific (Fig. 2a). The 300-hPa streamfunction field shows a combination of the PNA and the Tropical Northern Hemisphere (TNH) pattern (Fig. 2b). A pronounced negative phase of the TNH pattern is often observed during December and January when Pacific warm episode (ENSO) conditions are present (Barnston et al. 1991). If the decadal component is considered as shown in Zhang et al. (1997), the conventional PNA pattern will be more evident.

The SST pattern of the second mode, on the other hand, has the maximum variability in the central North Pacific, and this midlatitude SST variation is not accompanied with strong tropical SST anomalies (Fig. 2c). It is noted that this SST pattern is very similar to the second SVD mode from SVD analysis between zonal wind at 200 hPa and SST by Yang et al. (2002). The 300-hPa streamfunction anomaly of the second mode shows a great circle path from the equatorial date line to North America through the North Pacific (Fig. 2d). Figures 2c and 2d imply that the strong positive SST anomaly in the North Pacific is associated with the weakening of the Aleutian low. Meanwhile, the streamfunction pattern of the first SVD mode (Fig. 2b) shows the strengthening of the Aleutian low in the case of strong positive SST anomaly in the equatorial eastern Pacific.

From SVD analysis for all years as well as El Niño years using the SSTs in both the entire Pacific and only the North Pacific, we have found that the patterns from all-year SVD analyses are similar to those for El Niño years only. Therefore, the second mode seems to be a fundamental one. The SVD analysis of SST and 300-hPa zonal wind indicates that the second SVD mode is related to the variations of the strength and location of the Asian jet stream (not shown); in the case of the positive SST anomaly in the North Pacific, the Asian jet is shifted northward and weakened (Yang et al. 2002).

Figure 3 presents the time series of the amplitude of the first and the second SVDs normalized by each standard deviation. The time series of the first mode shows a positive trend, implying that recent El Niños have been
Figure 2. Horizontal distribution of the first and second SVD modes of the 300-hPa streamfunction anomaly and SST anomaly in the Pacific region in El Niño winters. Contour interval is 0.02 and the areas of negative values are shaded.

relatively strong. Meanwhile, the second mode lacks any systematic trends and includes substantial event-to-event variability. Thus, it is expected that the second mode may play an important role in producing the inter-

El Niño variability of the atmospheric circulation, although the first mode leads to the significant variation of the strength of ENSO events. In order to analyze the inter-El Niño variability of atmospheric circulation and the role of the second SVD mode, El Niño events have been classified according to the amplitude of SVD time series. Four winters (1977/78, 1982/83, 1991/92, and 1997/98) among El Niño events, which have positive maximum peaks in the first mode time series and negative or small values in the second one, are called “SIM1” years. The first SVD mode is dominant in these years. On the other hand, four winters (1951/52, 1963/64, 1965/66, and 1972/73), which have positive maximum peaks in the second mode time series, are called “SIM2” years. The SST and the atmospheric circulation in these years are influenced by the second mode as well as the first one. We will describe the characteristics of SIM1 and SIM2 in the next section.

b. Characteristics in SIM1 and SIM2

To investigate the characteristics of SIM1 and SIM2, composites of SST, latent heat flux, and 300-hPa stream-function anomalies have been made. Figure 4 shows the composites of the observed SST anomaly of SIM1 and SIM2. Since SIM1 years are strong El Niño years, the

![Diagram](https://via.placeholder.com/150)
pattern of the SST anomaly in SIM1 years is very similar to that of the SST of the first SVD mode that has strong positive anomalies in the equatorial Pacific and relatively weak negative anomalies in the North Pacific. In SIM2 years, when influenced by the second SVD mode in addition to the first mode, relatively weak and positive SST anomalies are located in the equatorial Pacific and positive anomalies in the North Pacific. Previous studies (Lau 1997; Lau and Nath 1994) indicated that there was a negative correlation between SST anomalies in the equatorial Pacific and those in the North Pacific. The first mode in our SVD analysis also shows a negative relationship between SST anomalies in the two regions. On the other hand, for the decadal timescale, the variation of SST in the midlatitude region of the Pacific Ocean is not related to the SST variation in the equatorial Pacific (Nakamura et al. 1997). Therefore, a large portion of the SST variation in SIM2 may exist over the decadal timescale because of little relationship between SST variation in two Pacific regions for the second SVD mode. The relationship between SST anomalies in these two Pacific regions is not yet completely revealed.

Figure 5 depicts the observed latent heat flux anomalies at the earth surface in SIM1 and SIM2 years. The positive heat flux anomalies exist in the central North Pacific where the negative SST anomalies are located in SIM1. This implies that the upward (downward) latent heat flux drives the SST to decrease (increase). In the case of SIM2, the downward latent heat flux is associated with the positive SST anomalies in the central and western North Pacific. Our result is fully consistent with those from previous studies (Lau 1997; Lau and Nath 1994). The pattern of the observed sensible heat flux is similar to that of the latent heat flux (not shown). It indicates that the SST does not directly influence the atmosphere but the atmosphere drives the variation of SST in the North Pacific. However, it is possible that the variation of SST indirectly influences the atmospheric circulation (Barnett et al. 1999; Bond and Harrison 2000; Yang et al. 2002). We will test the indirect effect of SST in the North Pacific on the atmospheric
circulation by conducting GCM experiments, as described in section 4.

The observed streamfunction anomaly pattern at 300 hPa in SIM1 and SIM2 is shown in Fig. 6. The streamfunction anomalies of SIM1 resemble the typical El Niño pattern. The anomaly distributions of SIM2 show that the negative center over the North Pacific and the positive center over the tropical Pacific are located more eastward than those of SIM1. They also indicate that other positive anomalies exist in the North Pacific. The amplitudes of the anomalies in SIM2 are smaller than those in SIM1. Since SIM2 patterns have larger components of the second mode than SIM1 patterns, it can be hypothesized that abnormal features in some El Niño years, such as the eastward shift of the positive and negative centers and the weakening of the Aleutian low, are caused by the second SVD mode. The abnormal pattern in these El Niño years is also seen in the zonal wind anomaly at 300 hPa (not shown). In SIM2 years the east Asian jet stream tends to be shifted northward and weakened, whereas these changes in the east Asian jet stream do not appear in SIM1 years (Yang et al. 2002).

4. Model experiments

The analysis of the observational data indicates that the second SVD mode is one of the factors causing the different atmospheric circulation among El Niño events. Since the first SVD mode represents only the variability of ENSO intensity, the effect of the second SVD mode on inter–El Niño variability of atmospheric circulation has been tested in this section. The SST of the second SVD mode has a large variability in the North Pacific (Fig. 2c), so using a GCM will test the effect of SST in the North Pacific on the atmospheric circulation in El Niño years.

\[ \text{Fig. 6. Observed streamfunction anomalies at 300 hPa of (a) SIM1 and (b) SIM2. Contour interval is } 2 \times 10^{-5} \text{ m}^2 \text{ s}^{-1} \text{ and zero lines are removed.} \]

\[ \text{a. Model and experimental design} \]

The Seoul National University (SNU) AGCM with triangular truncation at the two-dimensional wave-number 31 (T31), which has been developed in the Climatological Dynamics Laboratory of Seoul National University is used in our computations. This spectral model is based on the three-dimensional primitive equation with 20 vertical levels. The SNU AGCM consists of several physical processes and a detailed explanation of the model is provided by Kim (1999). Many researchers have adequately simulated the Asian monsoon (Shen et al. 1998), quasi-biennial oscillation (QBO) (Takahashi et al. 1997), Arctic Oscillation (Yamazaki and Shinya 1999), and several other phenomena such as convective activities (Baik and Takahashi 1995; Wang et al. 2000) using this AGCM.

Five experiments, TPNP, TPNN, TP, NP, and NN have been performed, using the SST anomalies shown in Fig. 7 as lower boundary conditions. These boundary conditions are fixed during the integration. Table 2 summarizes the above five experiments. The SST pattern of TPNP, which is close to the sum of the first and the second SVD modes of SST, represents the SST anomaly of SIM2. The SST pattern of TPNN is different from that of TPNP only in the sign of the SST anomaly in the North Pacific. The maximum value of the SST anomaly is about 3°C in the tropical Pacific and about 2°C in the North Pacific. Experiments TP, NP, and NN are performed to investigate the independent effect of each SST anomaly in the equatorial Pacific and the extratropical Pacific. We integrate the model for 180 days for each experiment and the results are averaged over the last 120 days. In this experiment, the calendar date is fixed at 16 January during the integration. Anomalies in model experiments are defined as the values subtracted by the mean of the control run, which is referred to the experiment without the SST anomaly forcing. Each experiment consists of 10 ensembles whose initial atmospheric conditions are chosen from the run of SNU AGCM. The SST, sea ice, and ozone data provided by the Atmospheric Model Intercomparison Project (AMIP) were used as boundary conditions in our run. Ten initial conditions for ensemble runs are given by SNU AGCM run results for 16 January of each year from 1987 to 1996. Any attempts to define atmospheric responses to tropical SST anomalies must recognize the
inherently chaotic nature of the atmospheric circulation. Therefore, distinguishing the SST-related signal from the noise generated by midlatitude internal dynamics requires ensemble averaging (Straus and Shukla 1997; Smith 1995; Barnett et al. 1994).

Figure 8 shows the simulated streamfunction anomaly at 300 hPa. The streamfunction anomaly patterns of TPNP, TPNN, and TP are very similar to one another except for their amplitudes. The amplitude of the streamfunction anomaly over the North Pacific in TPNN is much larger than that in TP, whereas the amplitude in TPNP is significantly smaller than that in TP. This implies that the magnitude of atmospheric responses in El Niño events depends partly on the sign of the SST anomaly in the North Pacific. The negative SST anomaly in the North Pacific enhances the atmospheric circulation anomalies forced by the warm tropical SST anomaly. To assess whether differences among these runs are significant, the \( t \) test has been carried out. The difference between TPNP and TPNN results and that between TPNN and TP results are significant with a
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Fig. 8. The streamfunction anomaly distributions at 300 hPa simulated in (a) TPNP, (b) TPNN, (c) TP, (d) NP, and (e) NN runs. Contour interval is $2 \times 10^6$ m$^2$ s$^{-1}$.

The differences of streamfunction intensity among three experiments (TPNP, TPNN, TP) are associated with the intensity differences of the diabatic heating rate anomaly that is shown in Fig. 9. One can see great matches between Fig. 8 and Fig. 9. The vertical profiles of diabatic heating anomalies indicate that the influence of SSTA is not confined to the planetary boundary layer. Extrema in the mid troposphere suggest that these diabatic heating anomalies are due to the latent heat. The SST anomalies (SSTAs) may cause changes in precipitation through their effect on static stability and vertical velocity in the storm track area, especially when the SSTA is positive in the North Pacific (not shown). The streamfunction amplitudes in NP and NN runs of Fig. 8 are much smaller than those in the other three experiments; especially, in NP where the amplitudes are much smaller. This result implies that the midlatitude atmospheric response to the North Pacific SST is more associated with other factors such as internal dynamics rather than SST itself. As shown in Figs. 8d and 8e, the atmospheric responses to SST in the North Pacific are smaller in NP run than in NN run, especially over the Atlantic Ocean. This implies that atmospheric responses are more influenced by internal dynamics in the NP run than the NN run because the differences among the 10 ensembles in the NN run are larger than those in the NP run. It will be mentioned in section 4c that the standard deviation of 10 ensembles reflects the effect of internal dynamics. Thus, the atmospheric responses to the negative SST anomaly are stronger than the responses to the positive SST anomaly in the North Pacific.

It is worthwhile to note that the amplitude of the difference of the streamfunction anomaly between TPNN and TP (Fig. 10b) is larger and more significant statistically than that between TPNP and TP (Fig. 10a). That is, the cooling of SST in the North Pacific has more impact on the atmospheric circulation than the warming of SST. There seems to be a strong nonlinear relation between the atmospheric responses and the SST anomaly in the North Pacific. On the other hand, the
amplitude of the difference between TPNP and TPNN (Fig. 10c) is larger than that of the difference between NP and NN (Fig. 10d). This also indicates that the SST anomaly in the North Pacific has more influence on the atmospheric response when it is accompanied by the tropical SST anomaly. Figure 10e presents the sum of NP and NN runs. It shows the nonlinearity of atmospheric responses to SST forcing in the North Pacific. In particular, the atmospheric response to North Pacific SST forcing is highly nonlinear in the midlatitudes of the Northern Hemisphere. According to Hoerling et al. (1997), circulation anomalies in the upper troposphere exhibit appreciable nonlinearity with respect to the SST forcing in the Tropics. During the northern winter, there is a phase shift in teleconnection patterns over the PNA region between El Niño and La Niña composites because two teleconnection patterns have different locations of tropical sources. They showed that this nonlinearity was due to the nonlinear interaction between the SST anomaly and the divergence anomaly as a source in the Tropics. However, the nonlinear process due to the midlatitude SST is different from that due to the tropical SST. From Figs. 10d and 10e, we imagine that this nonlinear process results from the nonlinear interaction between the SST and the diabatic heating.

We found from model experiments that the SST anomaly in the North Pacific associated with the second SVD mode could change the strength of midlatitude
response in El Niño events. However, the SST forcing in the North Pacific alone is not sufficient to explain the observed variability of the atmospheric circulation among El Niño events. Therefore, there may exist other factors responsible for the inter–El Niño variability in atmospheric circulation, for example, the transient forcing by the SST anomaly.

c. Internal dynamics

In the previous section, we showed that the SST anomaly in the North Pacific plays an important role in the inter–El Niño variability of atmospheric circulation and the effect of internal dynamics could be a factor causing this variability. Following the work in the previous section, we have investigated not only the relative importance of SST forcing and internal dynamics to atmospheric responses but also the sensitivity of the potential predictability of atmospheric circulation to SST forcing using the results of ensemble experiments. The sensitivity of atmospheric responses to atmospheric initial conditions can be used to quantify the interannual variability of atmospheric responses due to internal dynamics. The similarity among ensemble members can be used as a tool to measure the potential predictability due to the variation of SST. Based on these concepts, we introduce variance technique to measure the potential predictability (Rowell 1998).

The atmospheric response forced by the SST forcing consists of two independent components: one is due to SST itself and the other is due to internal dynamics. The SST-forced response can be represented by the magnitude of the 10-ensemble mean of the streamfunction anomaly ($|\psi'|$). The response forced by internal dy-
FIG. 11. Same as Fig. 8 except for the potential predictability of streamfunction at 300 hPa. Contour interval is 0.6 in (a)–(c) and 0.3 in (d)–(e). The area greater than 1.8 of potential predictability is shaded in (a)–(c), and that greater than 0.9 shaded in (d)–(e).

namics can be estimated by the standard deviation of 10 ensembles ($\sigma_{\text{int}}$). As explained by Rowell (1998), the potential predictability is measured as the ratio of SST-forced variability to the total variability. However, the SST-forced variability and the total variability cannot be defined in our study because we performed the January perpetual run. Instead, the potential predictability ($\rho$) is measured as the ratio of SST-forced response to the internal-dynamics-forced response: $\rho = |\psi'|/\sigma_{\text{int}}$. The larger $\rho$ means better predictability of the atmospheric circulation, which implies that the SST forcing has more influence on the atmospheric response than internal dynamics. According to Rowell (1998), the potential predictability varies with the location of response. In particular, the precipitation is more predictable in the Tropics than in the midlatitudes. In other words, the atmospheric circulation is more deterministic in the Tropics and more stochastic at the higher latitudes.

Figure 11 shows the potential predictability of the streamfunction at 300 hPa. The potential predictability is sensitive to both the position and the sign of the SST anomaly. For the tropical SST forcing, the predictabilities of TPNP, TPNN, and TP are commonly large in the PNA region. The atmospheric responses of NP and NN are less predictable than those of TPNP, TPNN, and TP due to the internal dynamics originating from initial conditions. That is, the SST forcing is important for the atmospheric responses in the Tropics, while the internal dynamics is significant for those in the midlatitudes. It is noted that the potential predictability also depends on the sign of the SST anomaly. When the sign of the SST anomaly in the North Pacific is the same as that in the Tropics (TPNP), the predictability is reduced compared to the TPNN. This difference is meaningful with a confidence level of 95%. Therefore, we may conclude that both the position and the sign of the SST anomaly in the North Pacific could be associated with the relative importance of external forcing and internal dynamics in the inter–El Niño variability of atmospheric circulation.

5. Summary and discussion

In this paper, we show the SST variation in the North Pacific as one of the factors responsible for the characteristic variability of the atmospheric circulation among El Niño events. In order to find the cause of inter–El Niño variability in the atmospheric circulation, we analyzed the observational data by employing a sta-
The atmospheric responses to the extratropical SST GCM experiments show that the extratropical SST in the El Niño years when positive peaks in the time series are less predictable than the responses to the tropical SST anomaly due to the effect of internal dynamics. In addition, when the signs of the SST anomaly in the North Pacific and/or the tropical Pacific SST control the intensity of atmospheric responses in the El Niño event. The effects of the extratropical SST anomaly on the atmospheric response are stronger when the tropical SST anomaly exists than when it does not exist.

The atmospheric responses to the extratropical SST anomaly are less predictable than the responses to the tropical SST anomaly due to the effect of internal dynamics. In addition, when the signs of the SST anomaly in the North Pacific and the Tropics are different from (the same as) each other, the effect of internal dynamics on the atmospheric circulation is less (more) pronounced.

These model results are similar to those of Lau (1997). The GOGA and Tropical Ocean Global Atmosphere (TOGA) mixed layer (ML) runs, which included the variation of midlatitude SST, showed that the midlatitude SST did not change significantly the pattern of the atmospheric circulation anomaly in the TOGA run but tended to enhance the anomaly. Our works improve Lau's (1997) research by considering the sign of North Pacific SST and the influence of both the presence and absence of tropical SST.

The SST has been warming since the 1950s especially in the equatorial eastern Pacific compared with the equatorial western Pacific, whereas a cooling trend is seen in the North Pacific (Fig. 12), which was also indicated by Lau and Weng (1999). The results of our experiments imply that the coupling effect of cooling in the North Pacific and warming in the equatorial Pacific has led to strengthening of the atmospheric response in recent El Niño years.

There are many factors to make the difference between the ST300 anomaly pattern of SIM1 and SIM2 in Fig. 6. The midlatitude SST forcing is one of them and we have tested the effect of midlatitude forcing to the atmospheric circulation in this paper. Although our analysis presents the role of North Pacific SST as one of the factors contributing to the variability of the atmospheric circulation among El Niño events, there remain some questions to be answered. The SST forcing alone is not sufficient to explain the observed inter–El Niño variability of the atmospheric circulation except for the weakening of the Aleutian low. According to Ferranti et al. (1994) and Saravanan (1998), the forced atmospheric responses to midlatitude SST anomalies roughly account for only one-third of the statistically derived coupled anomalies. Therefore, we considered transient forcing as another factor and tested the effect by using a barotropic model (not included in this paper). From this test, we found that the transient vorticity forcing could lead to the similar pattern in Fig. 6b. Thus, the atmospheric variation due to the transient forcing as mentioned in the introduction should be considered in order to understand more comprehensively the differences in the atmospheric circulation among El Niño events.

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