Four Decadal Ocean-Atmosphere Modes in the North Pacific Revealed by Various Analysis Methods

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Various statistical methods (empirical orthogonal function (EOF), rotated EOF, singular value decomposition (SVD), principal oscillation pattern (POP), complex EOF (CEOF) and joint CEOF) were applied to low-pass filtered (>7 years) sea surface temperature (SST), subsurface temperature and 500 hPa geopotential height in order to reveal standing and propagating features of decadal variations in the North Pacific. Four decadal ocean-atmosphere covariant modes were found in this study. The first mode is the well-known ENSO-like mode associated with the “Pacific-North American” atmospheric pattern, showing SST variations reversed between the tropics and the extratropics. In the western tropical Pacific, subsurface temperature variations were found to be out of phase with the SST variations. The other three modes are related to the oceanic general circulation composed of the subtropical gyre, the Alaskan gyre and the subpolar gyre, respectively. The 1988/89 event in the northern North Pacific was found to be closely associated with the subtropical gyre mode, and the atmospheric pattern associated with this mode is the Arctic Oscillation. An upper ocean heat budget analysis suggests that the surface net heat flux and mean gyre advection are important to the Alaskan gyre mode. For the subpolar gyre mode, the mean gyre advection, local Ekman pumping and surface net heat flux play important roles. Possible air-sea interactions in the North Pacific are also discussed. The oceanic signals for these decadal modes occupy a thick layer in the North Pacific, so that accumulated heat content may in turn support long-term climate variations.

1. Introduction

Many efforts have been made to investigate both the mechanisms and features of decadal variability in the North Pacific. One well-known hypothesis is based on the idea that the decadal variability of the tropical ocean-atmosphere system could bring about changes in the midlatitude Pacific through atmospheric teleconnection, just like the interannual El Niño-Southern Oscillation (ENSO) variability. This hypothesis appears to be supported by one robust example, the so-called 1976/77 climate regime shift in the North Pacific (e.g., Nitta and Yamada, 1989; Trenberth, 1990; Graham et al., 1994).

Subduction and shallow overturning cells may provide the subtropical-to-tropical feedback (e.g., Gu and Philander, 1997; Zhang et al., 1998; Schneider et al., 1999; Nonaka et al., 2000; Luo and Yamagata, 2001). This combination of oceanic pathways and atmospheric teleconnection (Lau, 1997) may complete a coupled scenario, called the tropics-extratropics interaction mode (see Latif (1998) for a review).

Some data analyses, however, suggested that a decadal mode independent of the tropical variations may exist in the northern North Pacific (Deser and Blackmon, 1995; Nakamura et al., 1997). Namely, winter sea surface temperature (SST) variations along the subarctic front zone associated with the “Pacific-North American” (PNA) atmospheric pattern are independent of SST variations in the eastern tropical Pacific. In a pioneering paper, Kort (1970) suggested a possible role of the oceanic subtropical gyre in large-scale air-sea interactions in the North Pacific. Latif and Barnett (1994) supported the idea that oceanic heat content anomalies circulate around the sub-
tropical gyre and advanced a hypothesis of their own. This phenomenon has also been confirmed by recent observations (Zhang and Levitus, 1997; Tourre et al., 1999). Using a coupled model, Latif and Barnett (1994, 1996) proposed that the atmospheric pattern coupled with this gyre mode is the PNA pattern; this appears to be consistent with the above data analysis results.

There is some other evidence which suggests that a decadal event independent of tropical influences was excited in the northern North Pacific around 1988/89. Namely, positive SST anomalies have been observed at midlatitudes since 1988/89, and the corresponding atmospheric pattern is a combination of anticyclonic anomalies over the subarctic North Pacific and cyclonic anomalies over the arctic regions (Tachibana et al., 1996; Walsh et al., 1996; Koide and Kodera, 1997). This occurrence of the atmospheric dipole pattern between the high and mid-latitudes has been referred to as the Arctic Oscillation (AO) (Thompson and Wallace, 2000). Whether the atmospheric pattern related to the midlatitude SST changes could be PNA or AO, as debated in previous studies, raises a difficult problem of determining the atmospheric response to midlatitude oceanic forcing due to strong internal variability in the midlatitude atmosphere. In fact, this issue has been addressed for almost two decades (e.g., Hoskins and Karoly, 1981; Palmer and Sun, 1985; Peng et al., 1997).

All of these studies have certainly deepened our understanding of the decadal variability in the North Pacific. However, the atmospheric pattern associated with the midlatitude independent mode is still in dispute. Furthermore, regional intense oceanic processes such as the Kuroshio and the Oyashio may also have important roles in the regional decadal phenomena (e.g., Latif and Barnett, 1996; Miller et al., 1998; Xie et al., 2000; Qiu, 2000; Seager et al., 2001; Luo and Yamagata, 2002). Most previous work has usually focused on the standing features; the lack of propagating evidence might be one origin of difficulty in seeking mechanisms to explain the decadal variations. Therefore, paying special attention to the regional phenomena, we have explored decadal ocean-atmosphere modes in the North Pacific. We have investigated both traveling and standing features in detail, and we suggest here possible physical mechanisms for those ocean-atmosphere covariant phenomena. Various hypotheses concerning the decadal variability in midlatitudes are also discussed.

The structure of this paper is as follows. Data and analysis methods are presented in Section 2. Section 3 deals with the standing mode analysis, and dynamics of regional modes are discussed in Section 4. In Section 5 we describe the propagation features. Summary and discussions are given in Section 6.

2. Data and Methods

Monthly SST data for the period 1950–1997 used in this study were taken from the Global Sea-Ice and SST (GISST, version 2) archive at the Hadley Center (Parker et al., 1995) with a high spatial resolution of 1° by 1°. As an atmospheric variable, we adopted monthly 500 hPa geopotential height (Z\textsubscript{500}), derived from the National Centers for Environmental Prediction (NCEP)/National Center for Atmospheric Research (NCAR) reanalysis data for the period 1958–1997. The data are gridded on a 2.5° × 2.5° mesh. We also used the monthly upper ocean temperature data from sea surface to the depth of 400 m for the period 1955–1998. This gridded data set was obtained from the Scripps Institution of Oceanography and has a resolution of 5° in longitude and 2° in latitude (White, 1995). For the upper ocean heat budget analysis we adopted the observed monthly surface heat flux and wind stress data (1° × 1°) produced by da Silva et al. (1994) for the period 1955–1993. For the period 1994–97 we used the corresponding NCEP/NCAR reanalysis data with possible jumps in 1993/94 removed. The present analysis used data mainly over the Pacific basin. We analyzed the monthly anomalies after removing the climatological seasonal cycle for the period 1961–1990.

We applied complex Morlet wavelet analysis to generate low-pass filtered (longer than 7 years) data (Torrence and Compo, 1998). The window was selected to remove the interannual ENSO signals and to focus on decadal variations (e.g., Nakamura et al., 1997). The term “decadal” in present study denotes the variability with a period longer than 7 years. To reduce influences of starting and ending points of the time series we adopted a padding method by mirroring the time series on two sides of the finite-length data to make the transition smooth. This yields good low-pass filtered results near the ends. Figure 1 shows an example; the low-pass filtered time series fits the observed one well, even at both ends (Fig. 1(a)). A coherence analysis confirms the in-phase covariability between the two time series for periods longer than 7 years (Fig. 1(b)). The results suggest that the influence of the end points is negligible.

Conventional empirical orthogonal function (EOF), rotated EOF (REOF) (see Richman (1986) for a review) and singular value decomposition (SVD) (Bretherton et al., 1992) methods were applied to explore standing decadal modes in the North Pacific. The complex EOF (CEOF) (Horel, 1984), principal oscillation pattern (POP) (see a summary review by Von Storch et al. (1995)) and joint CEOF methods were adopted to investigate traveling features for each mode. Note that the SST, subsurface temperature and Z\textsubscript{500} anomaly fields were normalized prior to the statistical analyses. Since we also study regional phenomena in the North Pacific, various methods...
have been adopted to build a reliable as well as comprehensive knowledge base about the decadal variations on the basis of careful comparison among results from the different methods. This is because each analysis method has its own bias. For example, the EOF, SVD and CEOF methods suffer from a strong orthogonality constraint, which may be inconsistent with real physical processes. The REOF and POP methods can efficiently weaken the orthogonality constraint, revealing local features and identifying independent harmonic modes, respectively. The joint CEOF modes must be orthogonal to each other, as required by the CEOF method. There is, however, no requirement for the atmospheric or oceanic variable alone to retain orthogonality (Newman and Sardeshmukh, 1995). It turned out that major results from the different methods are quite consistent, despite some minor intermethod differences. We therefore focus here on common features confirmed by all methods. To compare our results with previous work, we first reveal the standing features mainly obtained from the SVD method. Joint CEOF results are adopted to reveal the corresponding propagating features.

3. Standing Mode Analysis

3.1 Standing decadal modes in the North Pacific

The SST patterns and their expansion coefficients of five leading SVD modes between SST and $Z_{500}$ are well reproduced by the EOF and REOF methods. Among them, the expansion coefficients of the first and second SVD modes show an equivalent time scale, with the former lagging the latter by about 6 years. This means that the two modes actually capture the same phenomenon at two different evolution stages, which has been independently confirmed by the results from the CEOF, POP and joint CEOF. We have plotted the first, third, fourth and fifth SVD modes for SST and $Z_{500}$ in Fig. 2, which may represent four different physical processes. Shown are heterogeneous regression maps produced by regressing one field on the expansion coefficients of the other field. Homogeneous regression maps are similar to the heterogeneous maps. Time series of the SST and $Z_{500}$ fields for each SVD mode are highly correlated with correlation coefficients above 0.9. The fractions of the squared covariance explained by the four modes are 69%, 10%, 4% and 2%, respectively. Note that the fourth and fifth modes, despite their small contributions, are quite important in particular regions, as is demonstrated below. Their corresponding REOF and joint CEOF modes explain about 10% of the total variance, as shown in Sections 3 and 5.

The first leading SVD mode shows negative SST anomalies in the central North Pacific and positive anomalies along the northern and eastern Pacific boundaries during the positive phase (Fig. 2(a)). The related atmospheric change corresponds to the well-known PNA pattern, showing negative anomalies in the central North Pacific and positive ones along the western America. This mode represents a typical ENSO-like pattern (cf. Nitta and Yamada, 1989; Trenberth, 1990). The time series of its expansion coefficient clearly captures the prominent 1976/77 climate regime shift and is similar to that of SST anomalies in the central North Pacific (Fig. 3(a)). It is also consistent with the time series of SST anomalies in the eastern equatorial Pacific (not shown).

The third SVD mode, which is similar to the midlatitude decadal mode independent of the tropical influences, is characterized by zonally elongated positive SST anomalies centered near 30°–40°N (Fig. 2(b)). It is associated with zonally elongated atmospheric anticyclonic anomalies with the center displaced a little north of the SST anomalies. This ocean-atmosphere pattern is

![Figure 1](image-url)
similar to that reported by Koide and Kodera (1997) and Tachibana et al. (1996) in regard to the 1988/89 warming event in the northern North Pacific. We note in Fig. 3(b) that a rapid phase transition of this mode actually occurs around 1988/89. Yasunaka and Hanawa (2002) took this abrupt transition to be a regime shift in association with their own interpretations. The time series of its expansion coefficient is quite similar to the SST variations at midlatitudes (Fig. 3(b)) and also similar to that of surface vorticity field over the central Arctic Ocean given by Walsh et al. (1996) (see their figure 6). The result is consistent with the fact that the atmospheric pattern associated with the midlatitude SST change in late 1980s is not PNA but AO. This can be seen more clearly in the $Z_{500}$ field when we expand the analysis region to include the whole Northern Hemisphere (Fig. 4).

In addition to the above two decadal modes, which have been discussed by many investigators, two more local decadal modes have been found in the North Pacific. Figure 2(c) shows the fourth SVD mode. Negative SST anomalies are located in the central North Pacific, which are sandwiched by positive anomalies located to the west and to the southeast. The corresponding atmospheric signal is cyclonic with its center shifted a little north of the negative SST anomalies. The time series of the expansion coefficient is similar to that of the SST anomalies in the core region of this mode, showing a period of ~14 years (Fig. 3(c)).
The fifth SVD mode, although its contribution to the variance over the whole domain is weak, shows an interesting local feature: positive SST anomalies occur in the Kuroshio/Oyashio Extension (KOE) region, above which atmospheric anticyclonic anomalies are located (Fig. 2(d)). This ocean-atmosphere pattern is similar to the “warm SST-ridge” response of a model atmosphere forced by a similar SST anomaly pattern (Ferranti et al., 1994; Peng et al., 1997). Negative SST anomalies are seen in the north and northeast of the positive SST anomalies; this suggests a possible role of the subpolar gyre circulation. Figure 3(d) shows a downward linear trend in the time series of the expansion coefficient of this mode as well as decadal variations with a period of about 10 years. The time series corresponds well to the SST changes in the KOE region. The linear decreasing trend was also reported by Deser and Blackmon (1995). Miller et al. (1998) showed the similar temperature changes at a depth of 400 m in the Extension region.

Note that although the fractions of the total squared covariance between SST and $Z_{500}$ fields accounted for by the last two modes are rather small, they in fact explain 10.7% and 22.4% of the total SST variance averaged over the regional domains of $(30^\circ-55^\circ N, 170^\circ E-130^\circ W)$ and $(35^\circ-45^\circ N, 140^\circ E-180^\circ)$, respectively. The two domains are regions where strong signals appear for the fourth and fifth SVD modes, respectively. In such regions of active cyclogenesis, high-resolution AGCM experiments showed that the atmosphere is very sensitive to the localized SST forcing even if its contribution to the total variance is small (Ferranti et al., 1994; Peng et al., 1997).

Based on individual expansion coefficients of the four SVD modes, composite/difference maps of the upper ocean temperature averaged from the surface to the depth of 400 m are plotted in Fig. 5, which shows that thermal anomaly patterns are quite similar to the corresponding SST patterns of the SVD modes in Fig. 2. This fact indicates that decadal signals occupy a thick layer and maintain an in-phase variation with depth. White and Cayan (1998) also discovered that decadal temperature variations are approximately in phase at different depths with a larger amplitude in deeper layers. The large heat content in the thick oceanic layer may provide a long memory for each decadal process.
3.2 Robustness of the four standing modes

Several studies have claimed that the SVD may result in unrealistic relations of two fields even though the calculated correlations between them may be high (Newman and Sardeshmukh, 1995; Cherry, 1997; Hu, 1997). In this subsection, we check whether the decadal modes derived from the SVD method reflect real physical structures.

We first compared the SVDs with separate EOFs of the SST and $Z_{500}$ fields. Strong evidence of air-sea coupling is provided if the coefficient time series of one SVD mode and corresponding separate EOF modes are highly correlated, and the SVD patterns are similar to those of the separate EOF modes (cf. Newman and Sardeshmukh, 1995; Cherry, 1997). In the present case it was found that the patterns and time series of the corresponding EOFs resemble those of the SVDs (not shown). This suggests that the four SVDs do represent ocean-atmosphere covariant signals.

Figure 6(a) shows the percentage of variance explained by ten leading EOFs of the SST and $Z_{500}$ fields. Except for the first mode, they seem to be degenerate, considering the small number of degrees of freedom due to the low-pass filtering. It is known that the REOF method is effective in isolating real physical structures, even when the unrotated eigenvectors are completely degenerate (Cheng et al., 1995). Furthermore, REOF modes are statistically more robust and much less vulnerable to sampling errors (Richman, 1986). This method is also useful in extracting two independent modes when they have similar spatial patterns (Kawamura, 1994). Considering these advantages, we rotated the 10 leading EOFs of SST field with the Varimax method (Fig. 7). As commonly happens, the ranks of the corresponding REOFs are different from those of EOFs. The first, sixth, fourth and fifth REOFs correspond to the first, third, fourth and fifth SVDs, respectively. They account for 28.8%, 7.6%, 10.4% and 9.5% of the total variance, respectively. The corresponding SST patterns of the REOFs are similar to those of the SVDs. Their time series are also consistent with those of the SVDs and SST variations in the core regions for each mode (see Fig. 3, short-dashed lines). This suggests that the four decadal modes obtained by the SVD method are real physical structures in the North Pacific.

In an attempt to directly test the statistical significance of the SVDs, we adopted the Monte Carlo approach based on the squared covariance (SC) associated with each mode (e.g., Wallace et al., 1992; Peng and Fyfe, 1996). The $Z_{500}$ data are first randomly scrambled in the time domain without replacement, so that the chronological order of the SST field relative to the $Z_{500}$ field is destroyed. The time unit for the scrambling is 2 years; this unit length is a trade-off between the (limited) time span of the data set and the (limited) effective temporal degrees of freedom. Thus, 24 SST maps are paired with 24 $Z_{500}$ maps chosen randomly. The SVD is then applied to the scrambled data. Results from repeating the above procedure 100 times suggest that the four SVDs are robust (Fig. 6(b)). However, the significance test might be biased by the scrambling procedure because it could artificially increase the temporal degrees of freedom (Peng and Fyfe, 1996). The bootstrap method is useful in estimating standard errors and confidence intervals (Efron and Tibshirani, 1993).
The bootstrap procedure is similar to the Monte Carlo test, except that the $Z_{500}$ field is now randomly scrambled with replacement. Namely, the field at one time unit might replace others randomly and appear several times. The standard errors of the SCs explained by each SVD mode are estimated from 100 bootstrap samples produced by the above procedure. The ratios of the observed SCs to their standard errors for the four SVDs are 8, 13, 7 and 9, respectively, which suggests that the four SVDs are significant according to the null hypothesis. Dashed lines with error bars in Fig. 6(b) show the 95% confidence intervals for the observed SCs of each SVD mode derived from the bootstrap results. Few SCs from the Monte Carlo samples appear within the 95% confidence intervals, except for the second SVD mode. Those decadal modes suggest quite interesting dynamics, as is demonstrated in the following section.

4. Dynamics of Regional Modes

The dynamics for the ENSO-like mode and the 1988/89 event in the northern North Pacific will be reported elsewhere (Luo and Yamagata, 2001, 2002). We just mention here that air-sea interactions in the South Pacific and oceanic subduction process are important for the ENSO-like mode. For the 1988/89 event, the mean subtropical gyre advection and surface heat flux forcing related to air-sea interactions at midlatitudes are important.

In this section we focus on the dynamics of the two lower modes on the basis of an upper ocean heat budget analysis. Note that midlatitude atmospheric signals on decadal time scales are nearly barotropic; results based on the sea level pressure and surface winds are consistent with those of $Z_{500}$ (not shown, see also Chen et al. (1992)). The rate of change of upper ocean heat content anomaly is written as follows (McCulloch and Leach, 1998).

$$\frac{\partial H'}{\partial t} = Q_{\text{net}}' + \rho c \int_{-D}^{0} \left( -\bar{v}'_g \cdot \nabla T' + v'_g \cdot \nabla \right) + \left( w \frac{\partial T'}{\partial z} \right) + A_h \nabla^2 T' \, dz$$

where $H' = \rho c \int_{-D}^{0} T' \, dz$ is the heat content anomaly integrated from the depth $D$ to the sea surface. We set 400 m for the depth $D$ since decadal signals occupy a thick layer (see Fig. 5). $Q_{\text{net}}'$ is the surface net heat flux, $\rho$ and $c$ are the reference density and specific heat of sea water, $\bar{v}_g$ is the mean geostrophic velocity estimated from the temperature and salinity climatologies provided by the World Ocean Atlas 1998 with a reference level of 1500 dbar, $w$ is the vertical velocity component, $A_h$ is the horizontal eddy diffusion coefficient, $V_e = \tau \times k/\rho f$ is the Ekman transport, $\tau$ is the surface wind stress, and $T_e$ is the mean temperature in the Ekman layer. We simply assume that the Ekman layer depth is constant at a value of 40 m. The dashed round bracket ( )' denotes the monthly anomaly. The anomalous geostrophic advection, $-v'_g \cdot \nabla \bar{T}$, cannot be estimated since there are no observational data. We also neglect the diffusion and vertical advection terms for simplicity. To estimate each term in Eq. (1), we have interpolated all variables onto a $1^\circ \times 1^\circ$ grid. Results are then low-pass filtered (> 7 years) with the wavelet transform and divided by $\rho c D$ to give a unit of temperature.
Figure 8 shows the upper 400 m heat budget (see Eq. (1)) averaged in the region of ($180^\circ$–$150^\circ$W, $40^\circ$–$50^\circ$N). Values were divided by $\rho c D$ to give a unit of temperature. (a) The upper 400 m temperature anomaly ($H'/\rho c D$, solid line) and SST anomaly (dashed line). (b) As in (a), but for the temperature tendency (unit: °C/year). (c) As in (b), but for the horizontal mean geostrophic advection $[-1/\rho)[\int_0^L \mathbf{F}_g \cdot \nabla T dx]$. 0.21 (0.06) denotes the correlation coefficient between this term and the upper 400 m temperature (SST) tendency. (d) As in (c), but for the Ekman advection, $-1/\rho[V_e \cdot \nabla T_e']$. (e) As in (c), but for the surface net heat flux ($Q_{\text{net}}'/\rho c D$, solid line). Also shown is the latent and sensible heat flux (dashed line).

Figure 8 shows the upper 400 m heat budget in the core region ($180^\circ$–$150^\circ$W, $40^\circ$–$50^\circ$N) of the fourth SVD mode. Variations of the upper ocean temperature and its tendency are similar to those of the SST anomaly on decadal time scales (Figs. 8(a) and (b)). It is found that the mean geostrophic advection and surface net heat flux are of primary importance (Figs. 8(c) and (e)). The latter is mainly due to the latent and sensible heat flux (i.e., the dashed line in Fig. 8(e)). The Ekman advection also has a positive contribution but with a smaller amplitude (Fig. 8(d)).

The above analysis suggests that the fourth SVD mode might be caused by a similar mechanism to that for the 1988/89 event, as discussed in Luo and Yamagata (2002). This may also be inferred from the similar configuration of their ocean-atmosphere patterns. Such concurrent ocean-atmosphere patterns were suggested as evidence of the SST-wind stress-evaporation feedback process (Liu, 1993; Latif and Barnett, 1996). For example, the atmospheric anticyclonic anomaly shown in Fig. 2(b) suggests that anomalous easterlies appear above the center of the zonally elongated positive SST anomalies between $30^\circ$N and $40^\circ$N. The easterly anomalies reduce the strength of the climatological westerly at that latitude band and therefore decrease latent heat loss from the ocean. So the positive SST anomalies may grow. Forced by the strengthened positive SST anomalies, the atmospheric anticyclone might become stronger, as suggested by the model result published by Latif and Barnett (1996). This induces stronger easterly anomalies there, which in turn reduce heat loss from the ocean. This cycle suggests a positive feedback process between SST, wind stress and evaporation (Liu, 1993).

Figure 9 shows the upper 400 m heat budget for the fifth SVD mode in the KOE region ($140^\circ$–$170^\circ$E, $35^\circ$–$42^\circ$N). Variations of the upper ocean temperature and its tendency are consistent with those of the SST anomaly, especially for the period after the late 1970s (Figs. 9(a) and (b)). The mean geostrophic advection plays a dominant role in the decadal upper ocean temperature variation in the Extension region (Fig. 9(c)). The surface net heat flux also plays a positive role but with a smaller amplitude (Fig. 9(e)). The Ekman advection term is neg-
Using a relatively coarse ocean model, Miller et al. (1997) showed that the local Ekman pumping is important to the thermocline variations in the Extension region since the baroclinic Rossby waves emanated near the eastern boundary northward of 30°N propagate only as far as the date line. To study the role of local Ekman pumping, we plot in Fig. 9(f) the Ekman pumping speed (positive downward) in the Extension region. It has a positive correlation coefficient of 0.4 with the temperature tendency. With a relatively high resolution OGCM, we also found that the baroclinic Rossby waves do not intrude into the KOE region due to the strong mean eastward ocean currents there (see Appendix). In other words, there is a Rossby repeller. As a result, the thermocline depth variation is determined by the local Ekman pumping; the Sverdrup balance does not hold there (Gill, 1982). This suggests that the fifth SVD mode might be related to an air-sea interaction process different from that of the fourth SVD mode. As shown in Fig. 2(d), the atmospheric pattern associated with the positive SST anomalies in the KOE region is an anticyclone just above the SST anomalies. The negative wind stress curl associated with the anticyclone leads to Ekman downwelling and lowers the thermocline. As a result, the positive SST anomalies tend to be strengthened due to reduced cold water entrainment. We note that the SST anomalies are trapped along the KOE region; which suggests the importance of ocean dynamics there. The importance of the thermocline perturbations to the SST field in the KOE region was also reported by the model results of Xie et al. (2000). The SST anomalies might in turn enhance the atmospheric anticyclonic signal as suggested by the model results of Peng et al. (1997) and Ferranti et al. (1994). The whole process suggests a positive feedback between SST and Ekman pumping (Liu, 1993).

Using a 1.5 layer, quasi-geostrophic ocean model plus a linear SST equation, Liu (1993) suggested three unstable air-sea coupled modes in the extratropics which depend entirely on the relative position of the atmospheric response to the SST forcing. The two feedback processes
discussed above may correspond to the unstable SST-evaporation and upwelling modes according to his classification because of the similar ocean-atmosphere patterns. Such positive air-sea interactions could be of crucial importance for decadal signals to grow from the noise background, as suggested by the model result of Handorf et al. (1999). We note, however, that whether or not the atmospheric patterns shown by the above SVD modes are the atmospheric response to the oceanic conditions cannot be clarified in the present analysis; this cause and effect problem requires more observations and model studies.

5. Propagation Features

Results from the CEOF, POP and joint CEOF methods reveal similar traveling features. In the joint CEOF analysis we changed the resolution of SST field from 1° by 1° to 2° by 2° to make the number of its total grid points nearly equal to that of the $Z_{500}$ field, so that the variance contributions from either field are roughly equal. The four leading joint CEOF modes account for 47%, 21%, 12% and 9% of the total variance, respectively.

5.1 Features of the first mode

The first joint CEOF mode of $Z_{500}$ and SST shows a transition from a La Niña-like condition to an El Niño-like condition, which is associated with the 1976/77 climate regime shift. For comparison with the previous observational studies (cf. Nitta and Yamada, 1989; Trenberth, 1990), we have plotted in Fig. 10 the reconstructed evolution for the period around the mid-1970s regime shift. Weak positive SST anomalies originating near 120°W in the eastern equatorial Pacific in 1975–76, as seen in Fig. 10(c), are intensified and gradually enlarged (Figs. 10(d)–(f)). The positive SST anomalies in the central North Pacific last until 1975–76 with anticyclonic atmospheric anomalies above them (Figs. 10(a)–(c)); they are then replaced by negative signals (Figs. 10(d)–(f)). The midlatitude negative anomalies are associated with the positive SST anomalies in the eastern tropical Pacific. We note that the second SVD mode which was eliminated in Fig. 2 is similar to the pattern shown in Fig. 10(d).

The corresponding joint CEOF mode of $Z_{500}$ and subsurface temperature also shows that the subsurface temperature variation in the central North Pacific last until 1975–76 with anticyclonic atmospheric anomalies above them (Figs. 10(a)–(c)); they are then replaced by negative signals (Figs. 10(d)–(f)). The midlatitude negative anomalies are associated with the positive SST anomalies in the eastern tropical Pacific. We note that the second SVD mode which was eliminated in Fig. 2 is similar to the pattern shown in Fig. 10(d).

The corresponding joint CEOF mode of $Z_{500}$ and subsurface temperature also shows that the subsurface temperature variation in the central North Pacific last until 1975–76 with anticyclonic atmospheric anomalies above them (Figs. 10(a)–(c)); they are then replaced by negative signals (Figs. 10(d)–(f)). The midlatitude negative anomalies are associated with the positive SST anomalies in the eastern tropical Pacific. We note that the second SVD mode which was eliminated in Fig. 2 is similar to the pattern shown in Fig. 10(d).

The second joint CEOF mode of SST and $Z_{500}$ shows an ocean-atmosphere pattern similar to the third SVD mode. Zonally elongated positive SST anomalies at midlatitudes are correlated with anticyclonic atmospheric anomalies (not shown). The ocean-atmosphere pattern
moves gradually northeastward, which is reminiscent of the propagation of the northern part of the subtropical gyre mode. To make this point clear, we have plotted the corresponding joint CEOF result of $Z_{500}$ and the subsurface temperature in Fig. 13. The reconstructed evolution for the period around the 1988/89 event is shown to be comparable with the previous observational studies (cf. Tachibana et al., 1996; Koide and Kodera, 1997). In 1985–86, negative thermal anomalies tilting in a southwest-northeast direction appear in the northern North Pacific (Fig. 13(a)). Positive subsurface temperature anomalies are observed in the western tropical Pacific south of 20°N. The negative thermal anomalies move clockwise, passing through the “exchange window” at 30°–40°N near the eastern boundary (Liu et al., 1994), as seen in Figs. 13(b) and (c), and then reaching the western tropics (Figs. 13(d) and (e)). At the same time, positive anomalies appear along the northern edge of the subtropical gyre, which are accompanied by anticyclonic atmospheric anomalies (Figs. 13(b) and (c)). They then gradually move northeastward (Figs. 13(d) and (e)). This mode corresponds to the subtropical gyre mode discussed by several investigators (cf. Latif and Barnett, 1994; Zhang and Levitus, 1997; Tourre et al., 1999). It is interesting that the positive thermal anomalies east of the Philippines observed in 1985–86 reach the equatorial region along the western boundary in 1987–88. They then propagate eastward along the equator in 1989–90 and evolve into a large positive signal in the equatorial eastern Pacific in 1993–94. This may provide an explanation for the long El Niño event during 1990–95 (the dashed line in Fig. 12(b), see also Trenberth and Hoar (1996)). The above feature is consistent with OGCM results showing that signals subducted near the eastern boundary in the extratropics may finally reach the equatorial Pacific via the tropical western boundary (e.g., McCreary and Lu, 1994; Liu et al., 1994). The solid line in Fig. 12(b) shows approximately the temporal phase of this joint CEOF mode derived from the subsurface temperature variation at midlatitudes. The pronounced phase shift around 1988/89 is well captured by this time series, indicating that the midlatitude 1988/89 event is closely correlated with the subtropical gyre mode.

Fig. 13. As in Fig. 11, but for the second joint CEOF mode of $Z_{500}$ and subsurface temperature associated with the 1988/89 event.

Fig. 14. As in Fig. 10, but for the third joint CEOF mode of $Z_{500}$ and SST. The contour interval for $Z_{500}$ is 0.2. Shown is reconstructed evolution for about a half cycle based on the period when the temporal amplitude of this mode is larger and its phase evolution is smooth. The time interval of each map is 1.5-year except for 1-year in (c) and (d) in order to show the evolution during the transition phase more clearly.
5.3 Features of the third mode

Figure 14 shows the typical evolution feature of the third joint CEOF mode of SST and $Z_{500}$ for about a half cycle. The period for reconstruction is arbitrarily selected when the temporal amplitude of this mode is relatively larger, together with an approximate mean phase evolution. A negative SST anomaly appears in the midlatitude North Pacific, and the corresponding atmospheric pattern is cyclonic with its center located a little to the north (Figs. 14(a) and (b)). A positive SST anomaly is located along the northeastern boundary and appears to intrude southward along the Alaskan gyre into the area of the large negative anomaly in midlatitudes (Figs. 14(c) and (d)). This is consistent with the heat budget analysis in the Alaskan stream region, showing that the mean flow advection is important (see Fig. 8(c)). The intruded positive SST anomaly tends to grow (Figs. 14(e) and (f)); the corresponding atmospheric pattern is anticyclonic and also has its center located a little to the north (Fig. 14(f)). This arrangement might indicate the SST-wind stress-evaporation feedback discussed in previous section. Whether or not such an interesting feedback mechanism really exists, however, requires more study. The advection role of the Alaskan gyre was also envisioned by White and Cayan (1998). We call this mode the Alaskan gyre mode since the mean advection of this gyre appears to play an important role in the phase reversal of this mode.

We note that the signal along the Alaskan gyre from the northern boundary to the central North Pacific is weak during the phase transition period. This could be due to the low-pass filtering procedure which tends to reduce the signal severely in the strong current region. To check the evolution of the Alaskan gyre mode, we have also plotted in Fig. 15 the corresponding POP result of the SST field. The period and decay time of this POP mode are about 21 and 38 years, respectively. Evolution of the POP pattern can be interpreted as a cyclic sequence of spatial patterns, ..., $\rightarrow \text{Pi} \rightarrow \text{Pr} \rightarrow -\text{Pi} \rightarrow -\text{Pr} \rightarrow ...$, where $\text{Pi/Pr}$ denotes the imaginary/real part of the POP pattern (Von Storch et al., 1995). The evolution feature shown by this POP mode confirms the above joint CEOF result, namely, negative SST anomalies move southward from the northeastern boundary to the central North Pacific along the Alaskan gyre, and then grow and expand both eastward and westward. Meanwhile, the eastern part of the positive SST anomalies in midlatitudes move eastward and then extend northward along the eastern boundary. In the next stage, they turn southward to reverse the phase as inferred from the cyclic character of the POP method.

5.4 Features of the fourth mode

Figure 16 shows the typical evolution feature of the fourth joint CEOF mode of SST and $Z_{500}$ for about a half cycle. The period for reconstruction is arbitrarily selected just to show the typical evolution during half a cycle as in Fig. 14. A weak cold SST anomaly associated with a cyclonic atmospheric anomaly is found in the first stage (Figs. 16(a) and (f)). This ocean-atmosphere pattern is similar to the fifth SVD mode. The negative SST anomaly is gradually strengthened, which is associated with the cyclonic atmospheric anomaly (Figs. 16(b) and (c)). On the other hand, an initial positive SST anomaly located northeast of the negative anomaly extends westward along the northern boundary (Figs. 16(a)–(d)). It then expands southward along the western boundary to the KOE region (Fig. 16(e)); this is due to the mean flow advection, as suggested by the heat budget analysis (see Fig. 9(c)). Meanwhile, the original negative SST anomaly in the Extension region moves eastward rapidly (Figs. 16(d) and (e)). In the following stage a reversed ocean-atmosphere pattern starts to appear there (Fig. 16(f)). This ocean-atmosphere arrangement reminds us of the SST-Ekman pumping feedback discussed in previ-
ous section. Since the evolution revealed by this mode is greatly affected by the subpolar gyre circulation, we call this mode the subpolar gyre mode.

6. Summary and Discussion

We have used various statistical methods to explore decadal ocean-atmosphere phenomena in the North Pacific. By analyzing both standing and propagating features which complement each other, we focused on the common results confirmed by all methods to obtain a more reliable picture of the decadal phenomena. Since results from different methods are consistent, we conclude that four decadal modes exist in the North Pacific. Their existence is also consistent with physical arguments. The solid statistical significance of the four decadal modes, per se, cannot be established owing to limited observational data. The present analysis should therefore be regarded as a preliminary study of decadal variations in the North Pacific.

The first mode, associated with the atmospheric PNA pattern, shows anti-phase SST variations between the extratropics and the tropics. The result is consistent with previous works (cf. Nitta and Yamada, 1989; Trenberth, 1990). The subsurface temperature anomaly varies out of phase with the SST anomaly in the western tropical Pacific.

The second mode is closely related with the oceanic subtropical gyre circulation as discussed in previous studies (cf. Latif and Barnett, 1994; Zhang and Levitus, 1997; Tourre et al., 1999). The atmospheric pattern associated with this mode, however, is the Arctic Oscillation. It is found that the midlatitude 1988/89 event is closely related to this gyre mode. Another interesting feature of the subtropical gyre mode is that the associated midlatitude thermal anomaly reaches the equatorial Pacific after circulating around the gyre and brings about large temperature changes in the eastern equatorial region. This provides a major contribution to the lingering El Niño event during 1990–95.

The third and fourth decadal modes are also correlated with the ocean gyres such as the Alaskan gyre and the subpolar gyre, respectively. For the Alaskan gyre mode, the surface net heat flux and mean gyre advection play important roles. For the subpolar gyre mode, the mean gyre advection, local Ekman pumping and surface net heat flux are important. The importance of the mean gyre advection has been neglected in coarse resolution model studies (e.g., Latif and Barnett, 1996; Miller et al., 1998; Xie et al., 2000; Seager et al., 2001). Since our results are consistent with the recent analysis with a high-resolution model (Luo and Yamagata, 2002), the role of the mean and anomalous advection needs further studies.

The covariant ocean-atmosphere patterns as well as upper ocean heat budget analysis associated with the three decadal gyre modes suggest that two kinds of air-sea interactions might operate in the midlatitude North Pacific in accordance with the model and theoretical studies (cf. Liu, 1993; Ferranti et al., 1994; Latif and Barnett, 1996; Peng et al., 1997; Handorf et al., 1999). One is the SST-wind stress-evaporation feedback; the other is the SST-Ekman pumping feedback. Whether or not such air-sea interactions really exist is in dispute since the atmospheric response to the low-frequency SST forcing in midlatitudes is still unclear. Peng et al. (1997) found that the atmospheric response depends on its own climatological background. On the other hand, Hasselmann (1976) suggested that the stochastic atmospheric forcing may give rise to low-frequency SST variations. James and James (1989) argued that the nonlinearity of the atmosphere itself might generate weak low-frequency variance. Effects of the atmospheric forcing (Frankignoul, 1985) need to be investigated with more observations and model studies. However, assuming the existence of air-sea interactions has lead us to no physical inconsistency in the present study. Since the decadal signals occupy a thick oceanic layer, ocean dynamics also play important roles. It is possible that the coupled processes in midlatitudes may act to-

Fig. 16. As in Fig. 14, but for the fourth joint CEOF mode. Contour interval for $Z_{500}$ is 0.3.
Fig. A1. Main thermocline depth anomaly averaged between \( \sigma_\theta = 25.5 \) and \( \sigma_\theta = 26.5 \) at \( 35^\circ \text{N} \) based on outputs from a high-resolution OGCM. Zero contour is removed and \( \pm 10 \text{ m} \) contours are added for clarity.

Fig. A1. Main thermocline depth anomaly averaged between \( \sigma_\theta = 25.5 \) and \( \sigma_\theta = 26.5 \) at \( 35^\circ \text{N} \) based on outputs from a high-resolution OGCM. Zero contour is removed and \( \pm 10 \text{ m} \) contours are added for clarity.

gather to generate covariant decadal signals (e.g., Jin, 1997; Zorita and Frankignoul, 1997; Saravanan and McWilliams, 1998; Pierce et al., 2001). Their individual effects, however, are difficult to isolate in the present study; an answer to the cause-and-effect problem requires more fundamental studies, including a well-focused field experiment. Considering the fact that the available model results are quite different and model-dependent, the present analysis may be useful, at least, in evaluating model simulations of the decadal variations in the North Pacific.

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Appendix

An OGCM result
In order to check whether the Rossby waves reach the western boundary at midlatitudes, we analyzed thermocline depth variations using outputs from a high-resolution OGCM (Kagimoto, 1999). Its spatial resolution varies from \( 1/4^\circ \times 1/4^\circ \) near western boundary regions to \( 1/2^\circ \times 1/3^\circ \) in the east. The model is forced by NCEP/NCAR reanalysis daily winds and total surface net heat flux plus a correction term by restoring SSTs to observed values within 60 days. The depth anomaly at each grid point has been low-pass filtered (\( \geq 2 \text{ years} \)) with the wavelet transform. As shown in Fig. A1, the main thermocline depth variations between \( \sigma_\theta = 25.5 \) and \( \sigma_\theta = 26.5 \) at \( 35^\circ \text{N} \) clearly show stationary fluctuations west of \( 170^\circ \text{W} \); westward propagating signals starting from the eastern boundary do not reach the region west of \( 170^\circ \text{W} \).

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