Dominant Anomaly Patterns in the Near-Surface Baroclinicity and Accompanying Anomalies in the Atmosphere and Oceans. Part II: North Pacific Basin

MOTOTAKA NAKAMURA AND SHOZO YAMANE*

Japan Agency for Marine-Earth Science and Technology, Yokohama, Kanagawa, Japan

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ABSTRACT

Variability in the monthly-mean flow and storm track in the North Pacific basin is examined with a focus on the near-surface baroclinicity. Dominant patterns of anomalous near-surface baroclinicity found from empirical orthogonal function (EOF) analyses generally show mixed patterns of shift and changes in the strength of near-surface baroclinicity. Composited anomalies in the monthly-mean wind at various pressure levels based on the signals in the EOFs show accompanying anomalies in the mean flow up to 50 hPa in the winter and up to 100 hPa in other seasons. Anomalous eddy fields accompanying the anomalous near-surface baroclinicity patterns exhibit, broadly speaking, structures anticipated from simple linear theories of baroclinic instability, and suggest a tendency for anomalous wave fluxes to accelerate–decelerate the surface westerly accordingly. However, the relationship between anomalous eddy fields and anomalous near-surface baroclinicity in the midwinter is not consistent with the simple linear baroclinic instability theories. Composited anomalous sea surface temperature (SST) accompanying anomalous near-surface baroclinicity often exhibits moderate values and large spatial scales in the basin, rather than large values concentrated near the oceanic fronts. In the midsummer and in some cases in cold months, however, large SST anomalies are found around the Kuroshio–Oyashio Extensions. Accompanying anomalies in the net surface heat flux, SST in the preceding and following months, and meridional eddy heat flux in the lower troposphere suggest active roles played by the ocean in generating the concomitant anomalous large-scale atmospheric state in some of these cases.

1. Introduction

It has now become our basic knowledge that the extratropical atmosphere is driven strongly by the horizontal potential temperature gradient that arises from the differential solar heating (e.g., Lorenz 1955). The horizontal gradient in the potential temperature, often referred to as baroclinicity, is a measure of upper-level wind steering via thermal wind and a measure of baroclinic instability in the atmosphere. Baroclinicity in the lower atmosphere in classic theories of atmospheric stability is measured by a combination of the static stability and horizontal temperature gradient, the latter of which is equivalent to vertical shear in the horizontal wind through the thermal wind balance (Charney 1947; Eady 1949). In its original form, the Eady’s maximum growth rate for baroclinic instability \( B_{\text{GRMax}} \) is defined by \( B_{\text{GRMax}} = 0.31 \left( \frac{|f|}{N} \right) \left( \frac{\partial U}{\partial z} \right) \) in a zonally homogeneous steady mean state, where \( U \) is the mean zonal flow, \( f \) is the Coriolis parameter, and \( N \) is the Brunt–Väisälä frequency. Charney’s formula is slightly different from the Eady’s, but still incorporates the same effects.

Lindzen and Farrell (1980) first applied the Eady’s parameter to atmospheric data to successfully estimate the maximum growth rate of baroclinic instability in the troposphere. Hoskins and Valdes (1990) used its localized version (i.e., \( U, N, \) and \( f \) are all local Eulerian mean values) as the central parameter in their study of the Northern Hemispheric storm tracks. This local version, or its simplified version, has been used successfully as an indicator of baroclinic wave generation in diagnostic studies of storm tracks in recent years as well (Nakamura and Sampe 2002; Nakamura and Shimpo 2004; Nakamura et al. 2004).

* Current affiliation: Science and Engineering, Doshisha University, Kyotanabe, Kyoto, Japan.

Corresponding author address: Mototaka Nakamura, Japan Agency for Marine-Earth Science and Technology, 3173-25 Showa-machi, Kanazawa-ku, Yokohama, Kanagawa 236-0001, Japan.

E-mail: moto@jamstec.go.jp

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In our study, the North Atlantic part of which was reported in Nakamura and Yamane (2009, hereafter Part I), we define the near-surface baroclinic vector, \( \mathbf{B} = B^x \mathbf{i} + B^y \mathbf{j} \), where \( B^x = -\left(g/\theta N\right)(\partial \theta / \partial y) \) and \( B^y = \left(g/\theta N\right)(\partial \theta / \partial x) \) with \( \theta \) being the monthly-mean potential temperature at 2 m above the surface, and use it as the central quantity of the diagnoses. Unless stated otherwise, “anomalies” refer to deviations from the climatology hereafter. Though its meridional component does not appear in any classic theory of baroclinic instability, a theory that does incorporate the effect of \( B^y \) shows its important role in enhancing baroclinic wave generation locally to the east of the mean trough (Niehaus 1980). In the North Atlantic storm track region, we indeed found that the substantial zonal gradient in the surface temperature in and around the Labrador Sea plays a major role in the large-scale atmospheric state.

The sea surface temperature (SST) is an important factor in determining \( \mathbf{B} \) in the storm-track regions (e.g., Hoskins and Valdes 1990; Nakamura et al. 2004; Part I). SST anomalies (SSTAs) around an oceanic front along the Gulf Stream (GS), Kuroshio Extension (KE), or Oyashio Extension (OE) can have a profound impact on \( \mathbf{B} \) along the storm tracks. A subtle but important point that has to be considered carefully in this regard is the spatial scale and the location of SSTAs with respect to the climatology, since it is the anomalous surface temperature gradient whose structure has a spatial scale of the atmospheric Rossby deformation radius that can exert significant influence on the large-scale atmospheric flow. The high sensitivity of \( \mathbf{B} \) to changes in the temperature contrast across the front and changes in the width of the front, and the uncertainty in the impact of SSTAs of small spatial scales on \( \mathbf{B} \) make it difficult to assess the effective \( \mathbf{B} \) anomalies that are attributable to the SSTAs from the available data. Moreover, it is uncertain exactly how the SSTAs in the presence or absence of the land surface temperature anomalies may or may not produce \( \mathbf{B} \) anomalies that are significant to the atmosphere. The complicating factor introduced by the land surface must be taken into account when studying potential roles of extratropical SSTAs in the extratropical atmospheric anomalies.

Lau (1988) investigated patterns of anomalous storm track activity and associated low-frequency flow anomalies by computing empirical orthogonal functions (EOFs) for high-frequency 500-hPa geopotential height for the Northern Hemisphere winters. He found that both North Atlantic and North Pacific storm tracks have a pattern of meridional shift and a pattern of increased or decreased eddy activity in the first two EOFs. He also found that these changes in the storm tracks have symbiotic relationships with the background flows and have substantial impacts on the mean flow. Part I approached the issue of the storm-track and low-frequency flow variability in connection with SSTAs in the extratropics, focusing on \( \mathbf{B} \) as the key parameter of diagnoses, and found similar patterns of variability in the eddy activity and low-frequency flow in the North Atlantic basin. Much of this variability was connected to SSTAs in the vicinity of the Gulf Stream in cold months (Part I). Since the winter North Pacific basin has a storm track and mean flow that appear to be related to the oceanic fronts, Kuroshio–Oyashio Extensions (KOE) in this case, in a manner essentially the same as those in the North Atlantic basin related to the GS, we have attempted to find similar results for the North Pacific basin. In this regard, we have chosen not to project our results onto the major mode of variability in the extratropical North Pacific basin, the North Pacific decadal variability (PDV), so that our presentation and discussion are mostly confined to the wave–mean flow dynamics of monthly time scale or shorter.

Our approach to the search for a link between anomalies in the KOE and the overlying atmosphere is as follows: (i) identify dominant patterns in anomalous \( \mathbf{B} \) in the storm track for each calendar month and identify years in which the anomaly fits the pattern well, (ii) composite anomalies in the monthly-mean circulation and high-frequency transients in the atmosphere to obtain a typical atmospheric state that accompanies the patterns of anomalous \( \mathbf{B} \), (iii) composite SSTAs to obtain a typical oceanic state that accompanies and precedes the patterns of anomalous \( \mathbf{B} \), and (iv) composite anomalous net surface heat flux that accompanies and precedes the pattern of anomalous \( \mathbf{B} \). With this approach, we obtain typical pictures of anomalous states in the atmosphere and oceans with anomalous \( \mathbf{B} \) as their connecting interface.

Section 2 describes the data and procedure to compute \( \mathbf{B} \). Section 3 describes the climatology and variance of \( \mathbf{B} \). Dominant patterns of \( \mathbf{B} \) are shown in section 4, followed by composited anomalies in various atmospheric fields and SST in section 5. Finally, we present our discussion on the results, examining a potential cause–effect relationship between anomalies in the SST and atmosphere in section 6.

2. Data and calculation procedures

The data used to calculate \( \mathbf{B} \) are the monthly-mean temperature at 2 m above the surface (\( T^{2m} \)) and temperature at pressure levels available from the 40-yr European Centre for Medium-Range Weather Forecasts (ECMWF) Re-Analysis (ERA-40; Uppala et al. 2005). We chose the ERA-40 \( T^{2m} \) data rather than the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis products for its explicit inclusion of the observed near-surface temperature in producing the \( T^{2m} \) data. The monthly-mean surface pressure data from the NCEP–NCAR reanalyses
(Kalnay et al. 1996) were used to determine the pressure levels to be used for $B$ calculation, and to calculate $\theta$ at 2 m above the surface from $T^{2m}$. We used the NCEP–NCAR surface pressure data for convenience, since we had already compiled the dataset for calculating transient eddy fluxes to be mentioned later and the ERA-40 surface pressure data are not readily available. We later compared the NCEP–NCAR monthly-mean sea level pressure with that of ERA-40, and found the difference between the two products to be immaterial for the purpose of the current study. We also used ERA-40 monthly-mean horizontal wind and geopotential height at pressure levels, net surface heat flux, $F_h$ (the sum of latent heat flux, sensible heat flux, solar radiation, and the thermal radiation), and Hadley Centre sea surface temperature data (Rayner et al. 2003) to compile anomaly composites accompanying anomalous patterns in $B$. In addition, we used 6-hourly temperature and wind data from the NCEP–NCAR reanalyses to compute various eddy fields. The accuracy of the $F_h$ data used here is, as true for other reanalyses surface heat flux products, may not be so high to produce reliable anomaly composites.

We computed $B$ near the surface by calculating the horizontal gradient in $\theta^{2m}$, using the centered finite differencing, and calculating $N$ from the lowest three vertical pressure levels that are location dependent because of topography. Both $V\theta^{2m}$ and $N$ were calculated locally as in Hoskins and Valdes (1990) and Nakamura and Shimpo (2004). The entire 45 yr from September 1957 to August 2002 were used for the Northern Hemisphere. To resolve the dominant modes in $B$ arising from the land–sea temperature contrast, one may need much higher horizontal resolution in the data. The relatively coarse horizontal resolution of the data may artificially suppress the significance of the variability associated with the land–sea temperature contrast. One should keep this limitation in mind.

The 6-hourly bandpassed (period of 2–7 days) eddy fields and ultra-low-frequency (period of 30 days and longer) background fields were computed from the NCEP–NCAR reanalyses, using simple time filters (Lau and Lau 1984) first. The filtered time series were then visually examined against the raw time series and, then, used to calculate the slowly evolving bandpassed meridional velocity variance ($V'V'$), meridional temperature flux ($V'\theta'$), and the three-dimensional transient wave activity flux defined on a zonally varying basic state by Plumb (1986). The wave activity flux consists of the zonal and meridional advectional fluxes ($MU$ and $MV$), the zonal and meridional radiative fluxes ($MR^u$ and $MR^v$), and the radiative vertical flux ($MR^z$). The flux is essentially the Eliassen–Palm flux (Eliassen and Palm 1961) in a zonally inhomogeneous mean flow (Plumb 1986). The wave activity flux was calculated from February 1948 to November 2004 only for the extratropics poleward of 20° latitude. Also, it was calculated only from 850 to 30 hPa because of the double differentiation with respect to pressure required for the calculation. The flux of particular interest in this study is the horizontal gradient in $\theta^{2m}$, using the centered finite differencing, and calculating $\phi$ as prime denotes bandpassed component. The 6-hourly time series of wave fluxes was computed by using the time series of ultra-low-frequency fields as the basic-state and high-frequency fields as eddies. In short, the time series was calculated by changing the meaning of an overbar from the time mean state to an ultra-low-frequency state, and changing the meaning of a prime from a departure from the mean to a high-frequency state. The 6-hourly eddy time series was averaged over each month to produce monthly-mean time series. This dataset allows us to examine anomalous eddy fields accompanying anomalous $B$ in specific months. The climatology for the eddy fields was computed from 46 yr, January 1958 to December 2003. The calculation of the wave activity and its flux is described in detail by Nakamura et al. (2010).

3. Climatology and variance

The climatology and variance of $B^u$ and $B^v$ were computed for each calendar month and examined closely for their spatial and temporal structures. The monthly climatology, rather than the seasonal climatology, is used as the reference in our study, to avoid contamination of the diagnostically arising from differences in the climatology. In the following, thus, we focus our presentation on the climatology and variations of $B^u$ and their impact on the large-scale atmospheric state. Figure 1 shows the climatology of $B^u$, $U^{200}$, $V\theta^{850}$, $V'V^{850}$, $MR^{u850}$, $U^{1000}$, SST, and $F_h$ for February and August as examples of the reference state in the winter and summer. The numeric superscript indicates the pressure level in hPa. The seasonal mean is visibly more
diffused in structure, particularly for $B^x$, than those shown in Fig. 1. There are no surprises in the overall picture of $B^x$. The regions of large land–sea temperature contrast and the oceanic fronts show very large $B^x$ in cold months. The position of $B^x$ maximum in the storm track region is found along the KOE region throughout the year. The zonally elongated band of large $B^x$ in the storm track is generally wider in cold months than in warm months. In fact, large $B^x$ values that presumably accompany the Kuroshio and its extension in the cold months vanish in the summer (Figs. 1a,b). The seasonal variation in the $B^x$ values in the storm track basically follows that of the north–south differential heating—largest in the winter and smallest in the summer. As reported by Nakamura et al. (2004), the $U^{200}$ maximum is displaced southward from the $B^x$ maximum visibly in winter months (Fig. 1a), although $U^{200}$ is generally large over the area of large $B^x$ in the core of the storm track. The southward displacement of the $U^{200}$ maximum from the $B^x$ maximum in the storm track is less pronounced or even reversed in warmer months (Fig. 1b). We note that the structure of the climatological $B^x$ generally reflects those of $(\partial \theta^2 m/\partial y)$ with some exceptions where the surface slope contributes significantly to $B^x$ in isolated areas over the land, most notably around the Himalayas.

The position and structure of the storm track as indicated by $\nabla \theta^{850}$ and $\nabla \nabla U^{200}$ are, at least for the winter, essentially the same as those reported in earlier studies on the storm tracks (e.g., Chang et al. 2002). The maxima in $\nabla \theta^{850}$ and $\nabla \nabla U^{200}$ are located in the band of large $B^x$.
along the KOE in general, with a stronger tendency for the eddy fields’ maxima to be displaced southward and downstream in cold months. In the summer, however, they are displaced slightly to the north of the OE. The upper-tropospheric storm track core marked by the maximum $\nabla^2 D$ is located downstream of the jet core marked by the $U$ maximum throughout the year, with the displacement during cold months being much greater than that in the summer. The difference is attributed to the difference in the strength of the upper-tropospheric flow that advects synoptic-scale disturbances as they enter and grow in the storm track. The characteristics of MR$_{e850}$, whose divergence out of and convergence into the planetary boundary layer are related to, respectively, acceleration and deceleration of the surface pseudo-westerlies when the horizontal flux convergence is negligible (Plumb 1986), are roughly in accord with those of $\nabla^2 D$, which is a primary component of MR$_{e850}$. The sense of the surface momentum forcing by MR$_{e850}$ along the storm track is pseudo-eastward, since the potential vorticity gradient at 850 hPa is predominantly meridional and positive and MR$_{e850}$ is almost entirely upward along the storm track. The near-surface climatological $U$ along the storm track is generally positive and is, in part, attributed to this wave forcing.

The climatological SST and $F_p$ fields clearly show the fundamentally different roles of the ocean in the storm-track region between the winter and summer. The atmosphere receives large amounts of heat in the storm track during the winter, with the largest amount found along the Oyashio and its extension. On the other hand, the atmosphere gives moderate amounts of heat to the ocean during the summer, with larger amounts found in the northern ocean. We note that the meridional gradient of the SST across the OE is more or less the same throughout the year, while the gradient across the KE during the summer is considerably weaker than that during the winter.

The variance of $B^x$ is large over the land and in the vicinity of land–ocean boundaries (not shown). Values along the storm track are only moderately large and are roughly half of those in areas of large variance over the land and land–sea boundaries. These patterns arise from large fluctuations in the near-surface temperature over the land because of the much smaller heat capacity of the land surface as compared to that of oceans. Also, the large values to the north of the storm track may reflect the effect of sea ice in the ocean.

### 4. Dominant anomaly patterns of $B^x$}

To identify dominant anomaly patterns of $B^x$ in the vicinity of the storm track, we applied EOF analyses to $B^x$ for each calendar month in the domain with the boundary set by 130°E, 160°W, 20°N, and 45°N for January, February, March, April, and December, and in the domain with the boundary set by 130°E, 160°W, 20°N, and 55°N for May, June, July, August, September, October, and November. These areas of EOF application are indicated by red boxes in many of the figures shown in this paper. These domains have been chosen to include most of the Japanese islands and a part of the Eurasian continent, so that anomalies in $B^x$ arising from the land–sea temperature contrast are included in the diagnosis. The northern boundary of the EOF application for the winter and early spring is lower than that used for the warmer months. This choice was made to avoid the strong influence of the sea ice north of 45°N in the winter and early spring. When the northern boundary is set at 55°N for December, January, February, March, and April, the EOFs are noticeably affected by strong and noisy signals that arise from, evidently, the sea ice. For other months, the choice of the northern boundary between 45° and 55°N does not make any noticeable difference in the first two EOFs obtained. Although the sea ice may play a significant role in generating $B^x$ anomalies in some winter and spring months, we have chosen to remove its signals since our focus is on the impact of Kuroshio, Oyashio, and their extensions on the large-scale atmospheric state.

The percentages of $B^x$ variance explained by its first four EOFs are given for each calendar month in Table 1. The first EOF, or EOF1, of $B^x$ explains 29% or more of $B^x$ variance for winter months, December–March, and 16%–23% for other months. The values are generally greater for cooler months. This is also the case for the second EOF, or EOF2, of $B^x$. The EOF2 of $B^x$ explains 12%–24% of $B^x$ variance for each calendar month, with larger values seen in the cooler months. The third and fourth EOFs of $B^x$ each explain only about 10% or less of the

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variance. These numbers are comparable to those found for the first four EOFs of $B^x$ around the North Atlantic storm track. When the separation of these modes is tested by the North’s criterion (North et al. 1982), all EOF3 and EOF4 are found not to be separated well. According to the North’s criterion, EOF1 and EOF2 are separated well in only 6 out of the 12 calendar months, while EOF2 and EOF3 are separated well in only 4 out of the 12 calendar months. The robustness of the mode separation based on the North’s criterion is indicated by good (GD) and not good (NG) for the first three modes in Table 1. The reader needs to be cautious about this potentially unclean separation between the first four modes in the following presentation. In the following, we focus on the first two modes of $B^x$ for further examination.

Figure 2 shows the regression values of $B^x$ with the first two EOFs for February, May, August, and November, superimposed on the corresponding climatological values. The anomaly patterns for the North Pacific storm track are more diverse than those found in the North Atlantic basin. The first two EOFs of $B^x$ in the cold months, December–March, show two similar patterns as those found in the North Atlantic basin. One of the two patterns indicates a meridional shift of the band of large $B^x$. The second mode for February (Fig. 2b) shown is an example of this pattern. This pattern typically shows an area of large $B^x$ anomalies whose center is displaced by 500 km to 1000 km meridionally from the climatological $B^x$ maximum in the storm track. It is associated with

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**Fig. 2.** Climatological $B^x$ ($10^{-6}$ s$^{-1}$, contours) and regression ($10^{-6}$ s$^{-1}$, color) of $B^x$ with its EOF1 and EOF2. (top to bottom) February, May, August, and November. (left) EOF1 and (right) EOF2. Red rectangles indicate the domain of EOF calculations.
a meridional shift in the position of the tropospheric jet (Fig. 3b). This pattern clearly shows up in either EOF1 or EOF2 in December–March. The other mode shows a fairly clean pattern of increase or decrease in $B^x$ in the band of large climatological $B^x$ along the storm track. An example of this can be seen in the first mode for February (Fig. 2a). This pattern typically has an area of large $B^x$ anomalies whose center is more or less collocated with the climatological maximum in $B^x$. It is also accompanied by an enhanced or suppressed zonal flow (Fig. 3a). Although not as clear as in December–March, November EOF1 and EOF2 also show similar patterns (Figs. 2g,h and 3g,h).

These two anomaly patterns are qualitatively the same as two of the most prominent patterns found by Lau (1988). The EOFs for warmer months are not as well organized as those in the cooler months in general. However, the anomaly patterns associated with EOF1 for July and August show a clear tendency to shift the band of large $B^x$ and the tropospheric jet meridionally (Figs. 2e and 3e). On the other hand, the patterns accompanying EOF2 for July and August are neither the pattern of enhancement nor suppression of $B^x$ in the band of large $B^x$ in the storm track (Fig. 2f). Also, EOF1 for May exhibits an interesting anomaly pattern of a zonally elongated dipole with a visible southwest–northeast tilt, as shown in Figs. 2c and 3c.

The statistical significance of these modes for each of the 12 calendar months is robust. The correlation coefficient of $B^x$ with its first two EOFs is examined for the significance. The threshold values for the 95% and 99% confidence test by the Student's $t$ test are, respectively, 0.29 and 0.38. The correlation of $B^x$ with the first two EOFs is high in areas of large regression values in the vicinity of the
storm track for most months. It is statistically significant in some remote areas in cold months also. Examples are shown for February and August in Fig. 4. As found for the regression values, the correlation tends to be higher in cooler months. Also, the correlation tends to be high in larger areas in the cooler months than in warmer months.

The modes show significant relation with $U_{200}$ as well. The correlation of the EOFs with $U_{200}$ is statistically significant in large zonally elongated areas extending out from the domain of EOF analyses both upstream and downstream, more or less over areas in which the regression value of $U_{200}$ with the EOFs is 2 m s$^{-1}$ or greater in Fig. 3. The regression values of $U_{200}$ with the EOFs are nonnegligible compared to its standard deviation (Fig. 3).

5. Anomaly composites

To identify typical anomalous states that accompany the aforementioned anomalous patterns in $B^x$, we produced sets of anomaly composites for $B^x$, $T^{2m}$, $U$, $V$, $\nabla \theta$, $\nabla \cdot \nabla \nabla$, $MR^z$, SST and $F_h$. We examined the time series of each EOF for each calendar month and selected those years that reach or exceed the threshold absolute value of 0.5 standard deviation for the positive and negative phases separately for compositing. Since the distribution of the EOF time series values is not necessarily normal, some EOF time series exhibit small numbers of very large absolute values while some others exhibit large numbers of moderately large values. It results in the varying number of years used for anomaly composites listed in Tables 2–5. We simply averaged anomalies of these years to create pictures that represent the positive and negative anomalous conditions found by the EOF analyses. This approach may show us some of the nonlinear aspects of the atmosphere in strongly anomalous cases that may remain hidden in simple linear correlation analyses. However, we present the anomaly composites by showing the difference between the averages of the positive and negative phases (positive phase average minus the negative phase average) and, thus, focus on the linear aspects of the results. The years selected for the composites are listed for each calendar month in Tables 2, 3, 4, and 5. Note that some particular months of particular years are used in composites for both EOF1 and EOF2 of $B^x$. Most of these months show very large values in one of the two modes and marginal values in the other mode. To avoid subjective manipulation of the data, we used all months whose EOF time series reach or exceed 0.5 standard deviation in absolute value. Anomaly composites of $U$ and $V$ were made at 50, 100, 200, 500, 700, and 1000 hPa to examine the vertical structure of the anomalies. For the eddy quantities, $\nabla \theta$ at 850 hPa, $\nabla \cdot \nabla$ at 200 hPa, and $MR^z$ at various pressure levels were examined. The former two were chosen as the standard measures of the storm track variation, whereas the third was chosen as a measure of variation in the synoptic-scale wave forcing on the mean flow that originates from the anomalous $B$. The composites for cool months generally show structures that are defined more clearly than those for warm months.
a. Anomalies in the atmosphere

Composited anomalous $B^x$ and other atmospheric fields show interesting structures with substantial anomaly values with respect to the climatology most of the year. We will show some examples below. Figures 5a,b shows composites of anomalous $U^{200}$, superimposed on those of anomalous $B^x$ for the February EOF1 and EOF2. The same anomaly fields for August EOF1 are shown in Fig. 6a. Figures 5e,f and 6e show the corresponding composited anomalous horizontal temperature gradient anomalies in $U_{1000}$ and $MR_{850}$. The composited $U^{200}$ anomalies for each phase (which are roughly the half of the composited values shown in the figures) exhibit substantial values with the maximum absolute values ranging from 7 m s$^{-1}$ in the summer to 13 m s$^{-1}$ in the winter. The anomaly values for each phase are not negligible compared to the local climatology, and are comparable to the standard deviation of $U^{200}$ for each calendar month. The positions of the composited $U$ anomaly maxima are generally near or slightly downstream of the anomalous $B^x$ maxima. It is clear that these anomalies are manifestation of shifts in the latitudinal position and/or changes in the magnitude of the jet. For example, anomalous $U$ for February EOF1 reflects an enhanced or suppressed jet in its climatological meridional position (Fig. 5a), while the anomalous $U$ for February EOF2 reflects a slight meridional shift in the core of the jet (Fig. 5b). These characteristics are verified in the composites of the total $U$ also.

Most of the anomalous $U$ in the mid- to the upper troposphere is in thermal wind balance, especially in the vicinity of the storm track entrance, as one may anticipate from the geostrophic balance as the valid first approximation to the planetary- and synoptic-scale atmospheric motions. There is a general positive correlation between $B^x$ and $U$ in the troposphere and lower stratosphere. However, the anomalous thermal wind associated with the horizontal temperature gradient anomalies in the lowest part of the atmosphere alone does not explain all of the $U$ anomalies in the upper troposphere and stratosphere. The horizontal temperature gradient anomalies, though their magnitudes are generally smaller than those near the surface, exist throughout the troposphere. In particular, the anomalous horizontal temperature gradient anomaly along the core of the winter jet in the mid- and upper troposphere is roughly comparable in magnitude to that near the surface along the KOE. Anomalous $U$ typically shows equivalent-bartropic structures with the maximum $U$ anomalies found near the tropopause, extending up to 50 hPa in cold months and up to 100 hPa in the warmer months, demonstrating that the anomalous $B^x$ identified by the EOF is often associated with the planetary-scale anomalies of the external mode (Panetta et al. 1987; Held et al. 2002).

However, the anomalous $U$ in the summer shows larger vertical shear to the extent that the signs of the $U$ anomalies near the surface and upper troposphere are different in the core of the anomalies (Figs. 6a,e). In other words, the anomalous $U$ in the summer does not show a strong tendency toward a barotropic structure, indicating a weaker role of eddies in the summer anomalies. The core of the anomalous $U_{1000}$ is not visibly shifted very much downstream from the core of the anomalous $U^{200}$ in general, in contrast to the finding for the North Atlantic basin. Nevertheless, other characteristics of the composited anomalous $B^x$ and $U$ are broadly the same as those found for the North Atlantic basin.

Unlike the anomalous $U$, whose qualitative relationship with the anomalous $B^x$ suggests a simple thermal wind balance throughout the year, in some cases, the accompanying anomalies in eddy fields in the winter exhibit patterns that are not in accord with simple linear theories. Figures 5c,d,e,f shows composited anomalies in $\nabla \cdot U_{700}$, $\nabla \cdot U_{850}$, $MR_{850}$, and $U_{1000}$ for the February EOF1 and EOF2, while the composited anomalous eddy fields for

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August EOF1 are shown in Figs. 6c,e. In simple linear frameworks, locally enhanced $\nabla \cdot \theta_{850}$, $\nabla \cdot \nabla \cdot u_{200}$, and $MR_{850}$ for positive $B^\prime$ anomalies, and vice versa, are anticipated. We find this simple relation in most months (e.g., Fig. 6c). In particular, the pattern of anomalous $\nabla \cdot \theta_{850}$ generally suggests the diffusive role of transient eddies, acting to reduce the anomalous meridional gradient in $T_{2m}$ (e.g., Kushner and Held 1998). However, the winter months, January and February in particular, show strong signs of the eddy fields changing in a way opposite to that expected from simple linear theories. This is most noticeable in the anomalous $\nabla \cdot \theta_{850}$ that has the sign opposite of what is expected from the linear theory in zonally elongated bands that originate from the area of large $B^\prime$ anomalies in cases of enhanced–suppressed $B^\prime$ (e.g., February EOF1). It is only far downstream of the area of large $B^\prime$ anomalies that exhibits some anomalous $\nabla \cdot \theta_{850}$ of the “correct” sign in these cases. Also, there is a general tendency for anomalous $\nabla \cdot \nabla \cdot u_{200}$ to be displaced far downstream of the area of large $B^\prime$ anomaly and the maximum anomaly in $U_{200}$ in cold months.

This counterintuitive relationship between the eddy activity and $B^\prime$ is reminiscent of that found for the winter North Pacific storm track by Nakamura (1992). Unlike in the case of enhanced–suppressed $B^\prime$, the relationship between the anomalous $B^\prime$ and eddy fields in the cases of a meridional shift in the band of large $B^\prime$ is in accord with simple linear theories (e.g., February EOF2). The counterintuitive relationship is probably a manifestation of the stronger upper-tropospheric mean flow advection of synoptic-scale transients in the winter. In a paradigm of baroclinic wave growth through interactions of waves at the low level and tropopause (Hoskins et al. 1985), some time is needed for the waves to interact with each other and grow. When insufficient time is allowed for them to interact, wave growth is likely to be locally suppressed. On the other hand, the stronger advection of upper-level waves in the entrance region of the storm track may advect more undissipated energy downstream into the exit region where waves grow with help from the barotropic deformation field (Nakamura 1994; Swanson et al. 1997) or grow via processes demonstrated by Chang (1993) and Chang and Orlanski (1994). The results found here are consistent with the hypothesis of the eddy-trapping effect by the upper-level jet as the cause of the midwinter eddy suppression by Nakamura and Sampe (2002). The narrower–broader jet in the case of enhanced–suppressed $B^\prime$ and accompanying modification of waves by the altered jet may contribute to these changes also (Chang 2001; Harnik and Chang 2004). We also note that the anomalous $U_{200}$ has its peak farther downstream of the climatological maximum in $U_{200}$, affecting the deformation fields associated with the mean flow. The anomalies in the barotropic deformation fields may also be implicated in the enhancement–suppression of eddies in the case of weaker–stronger $B^\prime$ (Deng and Mak 2006). When the jet is shifted meridionally in the cases of meridionally shifted band of large $B^\prime$, such as in February EOF2 (Figs. 6b,d,f), mainly the location of the jet and eddy fields maxima are changed, rather than the strength of the jet and eddy fields. We also note that significant heating anomalies anticipated to accompany the anomalous $\nabla \cdot \theta_{850}$ are bound to modify the stationary waves and, thus, the mean flow throughout the troposphere and lower stratosphere (Held et al. 2002). The reader is referred to Chang et al. (2002) for a comprehensive review of the winter storm track research.

Anomalous wave forcing arising from the $B^\prime$ anomalies can be qualitatively deduced from composited anomalous $MR_{850}$. Also, the sense of anomalous momentum forcing of the underlying ocean by the anomalies in the synoptic-scale transients may be deduced from $MR^2$ near the surface. We use anomalous $MR_{850}$ rather than the divergence of the total transient wave activity flux, which
we have computed, because of very noisy contribution from the horizontal component of the flux. Pfeffer (1987, 1992) demonstrated that the horizontal flux contribution to the total E–P flux divergence is strongly correlated with the actual changes in the zonal-mean flow. Although his results in the zonal-mean framework do not necessarily apply to the zonally varying framework in our study, we note the potentially significant contribution from the horizontal component of the total wave activity flux in the lower troposphere in some areas. Composited anomalous MR\textsubscript{850} for February EOF1 and EOF2, and August EOF1 are shown in Figs. 5e,f and 6e, along with the accompanying anomalous $U^{1000}$. The magnitude of composited anomalous MR\textsubscript{850} is generally nonnegligible compared to its climatology. Positive and negative MR\textsubscript{850} anomalies (i.e., anomalous divergence and convergence of the wave flux in the planetary boundary layer, assuming negligible horizontal flux divergence) act to generate positive and negative $U$ anomalies below, respectively. Note that the anomalous $U^{1000}$ shown is not necessarily the product of the accompanying MR\textsubscript{850} anomalies, since there are other factors at upper levels that can affect $U^{1000}$. The anomalies in MR\textsubscript{850} only contribute to the anomalous $U^{1000}$ in a linear framework, positive and negative MR\textsubscript{850}}
anomalies are anticipated from, respectively, enhanced and suppressed $B^\circ$. While we find this linear relationship between $B^\circ$ and MR$_z^{850}$ to some extent in warmer months (e.g., August EOF1 in Fig. 6) and in cases of a meridional shift in the band of large $B^\circ$ in the winter (e.g., February EOF2), though the phase of the anomalous MR$_z^{850}$ is shifted to the north of where it should be by $10^\circ$ or so), we find practically no relationships in cases of an enhanced–suppressed $B^\circ$ in the winter (e.g., February EOF1). The structures of MR$_z^{850}$ anomalies in the latter cases are often opposite of what we expect from simple linear theories, and are sometimes completely unrelated to $B^\circ$ anomalies. This complicated relationship between the anomalous $B^\circ$ and MR$_z^{850}$ probably arises from the same factors as those mentioned for $V^\theta u^{850}$ and $V^\theta V^V^{200}$ in the preceding paragraph.

For the anomaly composites discussed above, we performed significance tests on the difference in the mean of the anomalies between the positive and negative phases, using the Student’s $t$ test. Examples of the tests are shown.
for February EOF1 and August EOF1 composites in Fig. 7. As found in the North Atlantic analyses, the composited anomalous $U$ is significant at the 95% confidence level or greater in large areas. For eddy fields shown above, the significance is somewhat lower, but not to the degree found in the North Atlantic basin, despite the strong proclivity of the eddies to respond to anomalous $B_x$ in a way opposite to that expected from simple linear theories in the cold months. The composited anomalies in the eddy fields are often significant at 95% confidence level or greater in large areas in the basin.

b. Anomalies in SST

The anomalous states of the ocean surface accompanying the composited anomalous $B^*$ for its first and second modes turn out to be somewhat surprising. The bottom panels of Fig. 5 and the bottom left panel of Fig. 6 show
composites of anomalous SST and $T^{2m}$ for February EOF1 and EOF2, and August EOF1. In a stark contrast to the composited SSTA patterns found in the North Atlantic basin diagnosis in Part I, only two cases in this work show convincing signs of SSTAs along the oceanic fronts accompanying major anomalies in the large-scale atmospheric state in cold months. Composited SSTAs in cold months generally do not have the structure that indicates active participation of the frontal currents in the generation of the SSTAs. Their horizontal structures tend to be broader than those found along the GS by Part I. Of course, this does not necessarily exclude the possibility of the SSTAs in the storm track forcing the large-scale atmospheric anomalies in the cold months. Perhaps the greater variability in the path of the Kuroshio and KE, as compared to the GS, may make detecting the impact of SSTAs along the KE on the large-scale atmosphere with a statistical approach difficult. Indeed, the strong impact of SSTAs along the Kuroshio and its extension does show up very clearly in March EOF2 and, to a lesser degree, in the December EOF2 as shown and discussed later. Also, there are some hints of SSTAs along the KE having impacts on the atmosphere in February composites (Figs. 5g,h). The most interesting finding in the STA composited is, however, a zonally elongated band of substantial SSTAs centered on the OE found in EOF1 of Jul and Aug (Fig. 6g). We refer to this band as Oyashio Extension summer anomaly (OESA) for short. Furthermore, patches of substantial SSTAs in the Sea of Japan and East China Sea are found to accompany OESA in the July and August composites (Fig. 6g).

The accompanying anomalous $T^{2m}$ has the same spatial scale as that of the anomalous SST in the summer composited, whereas the winter anomaly composites show visibly larger scales for $T^{2m}$ anomalies than those of the SSTA. In fact, the pattern and magnitude of anomalous $T^{2m}$ follow those of the underlying SSTA very closely in the summer composites, suggesting a relatively minor role played by the large-scale atmospheric dynamics in determining the anomalous $T^{2m}$ over the ocean in the summer in these cases. It is easy to see that the OESA is directly related with the anomalous $B^2$ (Figs. 6a,g). One should note that if these SSTAs found in the July and August composites were imposed on the winter reference state, they would probably have negligible impact on the large-scale atmospheric state in the winter, since the jet flows farther south of the band of SSTAs and there is another strong temperature front at the KE during the winter. As we emphasized in Part I, the location of SSTAs with respect to the climatological position of the upper-tropospheric jet and maximum $B^2$ in the storm track are critical factors in determining the atmospheric response to the SSTAs. Anomalies in $\nabla^2\theta^{500}$ are generally found to have a diffusive tendency to suppress the accompanying anomalous meridional $T^{2m}$ gradient that has a direct association with SSTAs, implying that the anomalous atmospheric eddy activity tends to destroy the effect of SSTAs on the lower atmosphere.

The difference between SSTAs in the positive and negative phases is significant at 95% confidence level or higher for OESA (Fig. 7j). For other months, particularly in the winter, the difference is significant in larger areas in the basin. Some small patches of highly significant SSTAs off the eastern coast of Japan are found in some cold months also. In particular, large SSTAs in the vicinity of the Kuroshio separation point found in the composites for March EOF2 show up highly significant in the test (not shown).

c. Relationship between the atmospheric and oceanic anomalies

To investigate the cause–effect relationship between the SSTAs in the storm track and the atmospheric anomalies, we compiled composites of anomalous net surface heat flux, $F_h$, using the years given in Tables 2–5. We refer to these composites as month 0 (Mo0) composites. We used the ERA-40 monthly-mean surface flux products, despite the large uncertainty in the surface heat flux produced by various reanalyses. We also produced composites for the 1-month period preceding (referred to as Mo−1) and following (referred to as Mo+1) the months listed in Tables 2–5, the change in the STA over the 3-month period preceding Mo0, which we refer to as dSSTA, and the anomalous $F_h$ integrated over the three months preceding Mo0, which we refer to as $\Sigma F_h A$. Note that we define positive $F_h$ to be upward, out of the ocean into the atmosphere. Therefore, anomalous $F_h$ and SSTA having the same sign indicates the anomalous forcing of the atmosphere by oceans, whereas their having different signs indicates the anomalous forcing of oceans by the atmosphere. We found some cases in which anomalous $U$, $T^{2m}$, and high-frequency storm activity are most likely the products of, at least in part, the direct thermal forcing from the concomitant SSTAs along the oceanic fronts. Following is the list of such cases: March EOF2, July EOF1, August EOF1, September EOF2, and December EOF2. We also found some possibility of direct and/or indirect oceanic thermal forcing (not limited to the SSTAs along the fronts) of the atmosphere in composites of February EOF1 and EOF2, March EOF1, June EOF2, September EOF1, and October EOF1 and EOF2. We focus our further discussion on these cases below.

Figure 8 shows anomalous $F_h$ superimposed on the corresponding SSTA for Mo−1, Mo0, and Mo+1, and
A superimposed on dSSTA for February EOF1 and EOF2, while those fields for August EOF1 are shown in the right column of Fig. 6. Since the March EOF2 case shows a striking picture, we also show these fields along with the climatological $B^*$ and EOF2 patterns of $B^*$ for March, and composited anomalies in $B^*$, $U_{200}$, $V_{850}$, $V_{200}$, SST, and $T_{2m}$ for March EOF2 in Fig. 9. Additionally, composited anomalous SST and $F_h$ for Mo−1 and Mo0 of January EOF1, March EOF1, May EOF1, June EOF2, July EOF1, September EOF1 and EOF2, October EOF1 and EOF2, and December EOF2 are shown in Figs. 10 and 11 to show other potentially interesting cases and two typical cases of atmospheric forcing of oceanic anomalies (January EOF1 and May EOF1).

Generally, the patterns of anomalous $F_h$ for Mo0 in the vicinity of the storm track suggest enhanced and suppressed heat release from the ocean to the atmosphere.
where $B^v$ is anomalously greater and smaller, respectively, except for July and August. Clear examples of this are seen in February EOF1 and EOF2 (Fig. 8). A major part of the heat release from the ocean to the atmosphere is attributed to the activity of synoptic-scale transient waves in the lower troposphere (e.g., Hazeleger et al. 2001). When these waves pass over the ocean, the anomalous surface wind associated with them facilitate the removal of heat from the oceanic surface. Therefore, one may expect a larger (smaller) heat release from the ocean to the atmosphere when and where $B^v$ is anomalously large (small), as synoptic-scale transients are generally more (less) active when and where $B^v$ is larger (smaller), except in the case of locally enhanced or suppressed $B^v$ in the midwinter that shows anomalous eddy fields of the sign opposite to what is expected from simple linear theories as noted earlier. This is indeed what one would find when the composited anomalous $F_h$ shown in Fig. 8d is examined against the anomalous $B^v$ shown in Fig. 5b. The same relationship is, however, found between the anomalous $B^v$ and $F_h$ shown in Figs. 5a and 8c, respectively, despite the suppressed lower-tropospheric eddy activity in the area of enhanced
Fig. 10. Composited anomalous SST (K, color) and $f_\theta \left[10^2 \text{ J m}^{-2} \text{ (6 h)}^{-1}\right]$ for (left) Mo−1 and (right) Mo0 for various cases. (a),(b) January EOF1; (c),(d) March EOF1; (e),(f) May EOF1; (g),(h) June EOF2; and (i),(j) July EOF1. Shown are the composites for the positive phase minus the composites for the negative phase. The years used are given in Tables 2, 3, 4, and 5.
Fig. 11. As in Fig. 10, but for (a),(b) September EOF1; (c),(d) September EOF2; (e),(f) October EOF1; (g),(h) October EOF2; and (i),(j) December EOF2.
anomalous SST and $B^s$ is found in most of the composites for months in which synoptic-scale transients are active, regardless of the sign of the actual anomalies in the synoptic-scale eddy fields. A pattern of anomaly in the vicinity of the storm track found in all of these cases is enhanced (suppressed) $\mathbf{U}^{1000}$ where $B^s$ is enhanced (suppressed). The enhanced (suppressed) $\mathbf{U}^{1000}$ obviously contributes to enhanced (suppressed) upward $F_h$. We thus interpret this relationship between anomalous $F_h$ and $B^s$ as the manifestation of enhanced synoptic-scale wave activity and surface zonal wind causing enhanced heat release from the ocean in the areas of enhanced $B^s$, and vice versa. The anomalous upward $F_h$ in the central North Pacific in the February EOF1 and EOF2 composites for Mo–1 and Mo0, accompanied by enhanced $B^s$, acts to cool the ocean and is responsible for some of the corresponding SSTAs in these composites. This tendency can also be deduced from the sense and magnitude of anomalous $T^{2m}$ and SST (Figs. 5g,h).

At the same time, there are some signs of small-scale anomalous $F_h$ resulting from the SSTAs near the western boundary and in the Sea of Japan. For example, the Mo0 composites for August EOF1 (Fig. 6c) and March EOF2 (Fig. 9d) show, respectively, large $F_h$ anomalies in the Sea of Japan and the vicinity of the Kuroshio separation point acting to force the atmospheric anomalies. Anomaly composites for March EOF1 (Figs. 10c,d), July EOF1 (Figs. 10i,j), September EOF2 (Figs. 11c,d), October EOF1 (Figs. 11e,f), and December EOF2 (Figs. 11i,j) all show signs of small-scale SSTAs forcing the atmosphere in the vicinity of Japan as well. The February EOF1 composites also show some heat release to the atmosphere off the east coast of the Japanese Main Island, despite the small magnitude and unorganized structure of the STA (Fig. 8c). Although these small-scale features may be artifacts introduced by the production process of the reanalysis products, the close correspondence in the structure of anomalous SST and $F_h$ in most of these cases is suggestive of the direct thermal oceanic forcing of the atmosphere to say the least. Among these cases, as mentioned earlier, the March EOF2 (Fig. 9), July EOF1 (Figs. 10i,j), August EOF1 (Fig. 6), September EOF2 (Figs. 11c,d), and December EOF2 (Figs. 11i,j) cases exhibit characters that strongly indicate active participation of the Kuroshio, Oyashio, and their extensions. In fact, the July EOF1 composites for Mo–1 (Fig. 10i) and Mo0 (Fig. 10j) exhibit clearer patterns of anomalous oceanic forcing of the atmosphere than the August EOF1 composites that show weak forcing of the ocean by the atmosphere in both Mo–1 and Mo0 in the KOE (Figs. 6b,d), although there are small areas of the atmospheric forcing of SSTAs along the southern and eastern edges of the area of large positive SSTAs. One may also argue that the February EOF1 and EOF2, March EOF1, and October EOF1 cases may be linked to anomalies in the Kuroshio and/or Oyashio and their extensions to some extent as discussed later.

Aside from the aforementioned cases, June EOF2 anomaly composites (Figs. 10g,h) show some clear signs of anomalous oceanic forcing on the atmosphere, while anomalous atmospheric forcing on the ocean is also visible. In the Mo0 composite (Fig. 10h), the SSTAs evidently force the atmospheric anomalies over a large area. However, at least some of the SSTAs are evidently generated by atmospheric thermal forcing in Mo–1 (Fig. 10g), except for several small areas along the Kuroshio and in the Sea of Japan. This may be a case of air–sea interaction of shorter time scales (less than a few months) involving clouds/radiation effects associated with the baiu–mei-yu Asian monsoon.

In contrast to the above, January EOF1 (Figs. 10a,b) and May EOF1 (Figs. 10e,f) cases clearly show forcing of the SSTAs by the accompanying atmospheric anomalies on the large scale in both Mo–1 and Mo0.

Overall, however, the composited anomalous SST and $F_h$ for Mo0 do not strongly indicate whether the atmosphere is generating the SSTAs or the other way around when details are examined. Also, the corresponding composited anomalous $\mathbf{U}^{1000}$ indicates no systematic relationships between anomalous Ekman transport and STA (not shown). Of course, the apparent lack of systematic relationship between STA and $F_h$ anomalies may be an artifact of poor quality and/or coarse resolution of the $F_h$ data. When the 3-month time series of the composited anomalies in SST and $F_h$ are examined, we begin to learn more about the potential roles played by the atmosphere and oceans. As evident in Figs. 6, 8, 9, 10, and 11, the basic structures of the SSTAs found in Mo0 often exist in Mo–1, suggesting that the SSTAs are preexisting, though are not necessarily accompanied by major atmospheric anomalies in the preceding months. This is often the case even for those cases in which the atmospheric forcing clearly tends to generate SSTAs in Mo0 (e.g., Figs. 10a,b,e,f). Since $B^s$ anomalies in the cases of OESA, March EOF2, and December EOF2 are directly attributed to the $T^{2m}$ anomalies that correspond to the SSTAs, the preexistence of the SSTAs implies that the anomalous atmospheric conditions in Mo0 are much more likely to be forced, at least in part, by the SSTAs, rather than the other way around in these cases. The basic structures of the SSTAs in Mo0 are also found persisting in Mo+1 often, further indicating the important role played by the ocean. In fact, the anomalous $F_h$ in Mo+1 for March EOF2 (Fig. 9f), July EOF1, August EOF1 (Fig. 6f), September EOF1 and EOF2, October EOF1 and EOF2, and December EOF2 in the western
North Pacific and/or Sea of Japan exhibits patterns that suggest the direct thermal forcing of the atmosphere by the ocean in Mo + 1. Among them, substantial anomalies in the mean flow and eddy fields in the atmosphere are found in response to the anomalous SSTAs via anomalous \( B^c \) in Mo + 1 as well as in March EOF1 and EOF2 (Fig. 9), July EOF1, September EOF2, and October EOF1 cases. Thus, the most straightforward interpretation of the composited anomalous SST and \( F_h \) in these cases is that of preexisting SSTAs in the storm track generating \( B^c \) anomalies, resulting in anomalous mean flow, anomalous transient wave activity, and anomalous \( F_h \).

By examining \( \text{dSSTA} \) and \( \Sigma F_h A \), we attempt to see whether the SSTAs in Mo0 are a product of cumulative anomalous \( F_h \) in the preceding three months. The results vary from case to case. As found for the North Atlantic basin, often the sense of \( \text{dSSTA} \) and \( \Sigma F_h A \) grossly smoothed by applying the 9-point averaging to them several times to produce synoptic-scale pictures is such that anomalous atmospheric forcing generates SSTAs in the preceding 3 months. Nevertheless, the details are not so clear cut at all. In the composites for cold months, there are obvious displacements between the location of large \( \text{dSSTA} \) and \( \Sigma F_h A \), indicating a possibility of ocean dynamics contributing to the generation of the SSTAs.

In regard to the systematic southward or southwestward displacement of the area of large \( \Sigma F_h A \) from the area of large \( \text{dSSTA} \) of opposite sign in the vicinity of the oceanic front found for the North Atlantic storm track, we find similar patterns in some of the composites for the winter North Pacific storm track also. A clear example is February EOF2 (Fig. 8h), in which areas of large \( \Sigma F_h A \) are located to the south or southeast of the areas of large \( \text{dSSTA} \) of opposite sign near the separation points of the Kuroshio and Oyashio. We speculate, as we did for the North Atlantic cases, that this pattern may be a signature of the oceanic advection redistributing the anomalous heat provided to the ocean by the atmosphere in such a way that the resulting SSTAs force back the atmosphere in a more effective manner with a time lag. We speculate that the following scenario may be at work in a case of positive SSTAs along the KE or a case of northward shift in the zone of large \( B^c \) due to factors other than the SSTAs. The SSTAs along the KE or some other factors, such as the internal variability of the atmosphere, act to shift the zone of large \( B^c \) northward, thereby acting to shift the zone of synoptic-scale transient wave activity also northward. This northward shift is accompanied by the same shift in the zone of large heat release from the ocean to the atmosphere, because of the large contribution of the synoptic-scale waves to the heat release from the ocean to the atmosphere. Therefore, this shift results in the downward \( F_h \) anomalies to the south of the KE, where \( B^c \) is anomalously small. The extra heat provided to the ocean to the south and east of the Kuroshio and its extension may be collected by westward-propagating eddies and recirculation gyre, transported to the Kuroshio near the western boundary, then transported northward and eastward by the Kuroshio and its extension along their path, and finally modify \( B^c \) in the core of the storm track to force back the atmosphere. In this scenario, the SSTAs may act simply as a trigger for anomalous air–sea interaction via anomalous transient eddies and mean flow, and may not necessarily directly dictate the anomalous \( F_h \). We speculate, therefore, that the anomalous oceanic heat transport may play an important role in cases like the February EOF2, despite the lack of clear signals of the oceanic thermal forcing of the atmosphere. In general, however, this displaced pattern of \( \Sigma F_h A \) and \( \text{dSSTA} \) found here is not so clear cut compared to that found in the North Atlantic basin. We suspect that the more complex structure of the oceanic frontal currents in this region (KE, OE, and Tsushima Warm Current that separates from the Kuroshio and flows into the Sea of Japan) makes it difficult for the atmosphere–ocean system to produce patterns that are easily extracted and identified by a statistical method. This scenario is unlikely in the summer and autumn cases.

6. Discussion and conclusions

The results found for the North Pacific basin presented above are somewhat similar to those found for the North Atlantic basin in regard to the atmospheric anomalies. One notable difference found is in the anomalous eddy activity during the midwinter accompanying the dominant anomalous patterns of \( B^c \). Anomalous SST accompanying anomalous \( B^c \) in the North Pacific storm track, on the other
hand, presents a surprise to us. Contrary to our anticipation of finding significant SSTAs along the KOE in the winter, the season believed to have the strongest air–sea exchanges, the most striking SSTAs we find in the composites turn out to lie along the OE in July and August. This unexpected result is, in fact, not so surprising in the light of various differences in the atmosphere and oceans between the North Atlantic and North Pacific basins.

First of all, in the North Pacific, there are two sharp SST fronts in cold months, one associated with the KE and the other with the OE, while only the front associated with the OE remains sharp in the summer. Thus, the impact of oceanic frontal variations on the atmosphere is expected to appear more clearly in the summer than in the winter. Also, the land–sea temperature contrast created by the Eurasian continent, Sea of Japan, and Japanese islands complicates the situation in cold months. The Sea of Japan also releases large amounts of heat into the atmosphere when cold air mass blows over it during cold months, impacting the atmosphere before the Pacific Ocean modifies the air mass. Furthermore, topographic forcing of the atmosphere by the Tibetan Plateau in the winter is likely to affect far downstream of the plateau via nonlinear processes (Waugh et al. 1994). In fact, it is probably the reason why there is a zonally elongated region of no significant correlation between the upper-tropospheric zonal flow and $B_x$, stretching eastward from the Tibetan Plateau downstream to Japan in most months (not shown). These factors, other than the KOE, which can generate variability in the North Pacific storm track are generally the strongest in the winter and weakest in the summer. Thus, the emergence of OESA in our diagnoses may well be a manifestation of the variability due to the OE becoming clearer or more readily isolated in the summer than in the winter, because the variability due to the other factors in the summer is weaker than it is in the winter. We note that the depth of the mixed layer, which is much shallower in the summer compared to the winter may facilitate generation of large SSTAs in the summer by allowing the ocean to accumulate the impact of anomalous oceanic heat transport or anomalous surface heat flux in the shallower mixed layer.

The results presented here suggest the presence of feedbacks operating between the atmosphere and ocean, probably including the land and ice, which appear more complex than those in the North Atlantic storm track. We intentionally excluded the strong impact of sea ice on $B$ in the winter and early spring in our work to keep our focus on the relationship between SSTAs and the atmosphere in the vicinity of the storm track. Ice plays a major role in controlling the albedo and, when it is present over the ocean, the surface heat flux between the atmosphere and oceans, thereby playing an important role in determining the atmospheric surface temperature in the winter at high latitudes. Thus, a comprehensive study on the atmospheric variability that originates from the surface anomalies in the western North Pacific must treat the effects of sea ice in the winter and early spring in a clever way. The land surface, whose thermal inertia is much smaller than that of the ocean or thick ice, can have a nonnegligible impact on the atmosphere because of its temperature contrast with the ocean and also by modulating the surface heat flux of the atmosphere over the land. The presence of the Japanese islands, whose counterparts are absent in the western North Atlantic, can complicate the interactions between the atmosphere and oceans in the western North Pacific region.

While the extratropical SSTAs over a large area in the central North Pacific associated with the PDV are believed to be linked strongly to the tropical Pacific SSTAs via the “atmospheric bridge” (e.g., Nitta and Yamada 1989; Lau and Nath 1994), those along the subpolar front in the western and central North Pacific seem to have their own mechanisms that generate decadal variability (e.g., Miller et al. 1994; Nakamura et al. 1997). Regarding the oceanic impact on the atmosphere on the decadal time scale, there is some evidence of large-scale atmospheric response to the decadal SSTAs, though fairly weak, to the anomalies along the subpolar front also (Nakamura et al. 1997). Numerical simulation studies on the impact of SSTAs along the subpolar front also suggest, to some extent, that the synoptic- and planetary-scale atmospheric processes respond to the SSTAs (e.g., Kwon and Deser 2007; Taguchi et al. 2009). Interestingly, these SSTAs along the subpolar front are not found to consistently affect the large-scale atmospheric state during the winter in our study. Instead, they are found to have noticeable impacts on the atmosphere in July and August (Fig. 6) here. We speculate that slowly evolving SSTAs due to large-scale oceanic processes (e.g., Seager et al. 2001; Qiu 2003) provide the background condition that tends to aid or suppress the generation of the OESA, but not necessarily generate strong response in the atmosphere in the cold months in a consistent manner, due to the factors discussed throughout this paper. The clear manifestation of the oceanic impact on the atmosphere in the summer may be linked to the strong radiative feedback between the atmosphere and ocean in this region during the spring and summer found by Park et al. (2005). In our analyses, the only robust SSTAs affecting the large-scale atmospheric state in the cold months are found in the vicinity of the Kuroshio and Oyashio separation points, where they modify $B^*$ at its climatological maximum in the storm track, in March and December, when the upper-level jet is not at its peak. When the variability of decadal time scale is the focus, it is probably difficult to extract cases in which
the upper-level jet and $B'$ anomalies generated by SSTAs happen to come together to produce major anomalies in the large-scale atmospheric state in the manner reported here due to the temporal smoothing of the data. With such smoothing of the data, the atmospheric signals found in response to SSTAs in the frontal regions are, we suspect, more likely to be the kind of signals expected from a system in the quasigeostrophic balance. Indeed, quasigeostrophic atmospheric circulation models are known to produce simple surface low pressure anomalies in response to anomalously warm lower boundary (I. Held 1994, personal communication). Because of these factors, we find it difficult to compare our results with the results of studies on the PDV.

We presented above a look at the extratropical atmospheric variability in the North Pacific basin from a perspective of the surface forcing variability. The results are broadly similar to those for the North Atlantic basin. Nevertheless, there are significant differences between the results found for the two basins that may well require two different sets of theories to describe the variability in the two basins that appear qualitatively the same at a glance. Analyses that use $B$ as the key parameter shall, we strongly believe, provide the best framework for extracting the complex mechanisms that generate large-scale atmospheric anomalies in the extratropics from the data or numerical model output.

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