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Association between Tropical Convection and Boreal Wintertime Extratropical Circulation in 1982/83 and 1988/89

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ABSTRACT

Boreal wintertime extratropical circulation is studied in relation to the tropical convection during the 1982/83 El Niño and 1988/89 La Niña. The anomaly structure of 1982/83 and 1988/89 over the extratropics reveals remarkably different features as the longitudinal tropical forcing region changes. The Rossby wave source (Positive) shows the largest maximum over East Asia in both years due to the persistent heating from the western Pacific warm pool area. However, the sink term shows contrasting features over the subtropics and extratropics between the two years. In the El Niño year, enhanced tropical convection over the eastern Pacific produces the Rossby wave sink at 10°N and shifted eastward over the North Pacific, while in the La Niña year, the sink area is shifted westward over the North Pacific. The contrasting features between the two events in mean-eddy interaction appears especially over the downstream area of the East Asian Jet. The extension (retraction) of the meanflow eastward (westward) to the east (west) of the dateline is related with the effect of the westward (eastward) E-vector and the strengthening (weakening) of the negative anomalies of the barotropic growth of kinetic energy. Hence, almost opposite characteristics between the two events can explain the close relationship of tropical convection and the extratropical internal variability.

Key words: tropical convection, extratropical circulation, rossby wave source, mean-eddy interaction

1. Introduction

The dynamics of the teleconnection between the tropics and extratropics continue to be a central problem in climate research. The tropical weather system is characterized by deep convective activity with large horizontal and vertical scales, and the extratropical climate is dominated by traveling waves with cyclones and anticyclones. Since the work of Horel and Wallace (1981), the tropical-extratropical interaction has been studied in the case of ENSO (El Niño-Southern Oscillation) events (Geisler et al., 1985; Lau and Boyle, 1987; Rasmusson and Mo, 1993; Mattews and Kiladis, 1999; Lee et al., 2002), and from the work of Liebmann and Hartmann (1984) and Weickmann et al. (1985), it has been studied on intraseasonal timescales such as the Madden-Julian Oscillation (MJO) (Schubert and Park, 1991; Park et al., 1995; Higgins and Mo, 1997).

Previous studies which have simulated the atmospheric response to tropical SST anomalies suggested that the problem of the extratropical response to anomalous tropical heating is essentially that of Rossby wave propagation and dispersion in a sheared ambient flow on the sphere, with the wave source located in the tropical upper troposphere (Hoskins and Karoly, 1981; Simmons, 1982; Sardeshmukh and Hoskins, 1988; Yang and Webster, 1990; Grimm and Silva Dias, 1995). Sardeshmukh and Hoskins (1988) systematically considered the Rossby wave source to find out the role of tropical heating in the general circulation and emphasized the importance of modeling the horizontal and vertical structure of heating accurately. Park et al. (1995) and Higgins and Mo (1997) analyzed the response of circulation anomalies over the north Pacific using a Rossby wave source and found the expansion and retraction of the EAJ (East Asian Jet) is closely connected with the Rossby wave source and sink which is associated with the tropical heating. However, Yang et al. (2002) emphasized that the variations of the EAJ were not significant from one year

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to another. The EAJ stream index (30°–35°N, 130°–160°E) was not strongly linked to ENSO and was coupled to a teleconnection pattern that was distinguished clearly from the ENSO-related pattern.

This study follows Park et al. (1995) and Higgins and Mo (1997), and aims to ameliorate our understanding concerning the extratropical connection with the tropics during the boreal winter season. We will first calculate the Rossby wave source function based on a theoretical consideration proposed by Sardeshmukh and Hoskins (1985) and investigate the results by comparing El Niño and La Niña years.

There are many studies (Lau and Boyle, 1987; Ting and Held, 1990; Yu and Huang, 1995; Chen and Van den Dool, 1997) that have put an emphasis on the relationship between tropical convection and the extratropical transient eddy activity. Lau and Boyle (1987) have found that the extratropical response, including the PNA (Pacific North America) pattern, appeared to be more sensitive to the dipole heating with enhanced convection over the maritime continent of Indonesia/Borneo, than the enhanced convection over the central Pacific, which is in agreement with the theoretical findings of Branstator (1985). In addition, they have shown that the extratropical forcing by the transient eddies appeared to be unimportant in contributing to the large-scale circulation anomalies. Yu and Huang (1995) have shown that the variations of intensity in tropical convection could modify the intensity of interaction between low frequency fluctuation and zonal flow over the extratropics. Moreover, results from wave-mean flow energy conversion showed that the zonal flow could transfer its energy to low frequency fluctuation and to concentrate low frequency kinetic energy in the downstream region of the jet. They noted that transient forcing plays an important adjustment role in the interaction between the westerly jet and low frequency kinetic energy in the winters of 1983/84 and 1986/87. Chen and Van den Dool (1997) studied the large asymmetric impacts of the tropical Pacific SST anomalies on the atmospheric internal variability over the North Pacific, and found the reasons for the blocking flows being developed by the high-frequency transients during La Niña more than El Niño years.

Some of the previous studies suggested the insensitivity of the extratropical response by the tropical forcing and some of the other studies have insisted on the importance of tropical forcing on the extratropical mean-eddy interaction. The present study intends to confirm the effect of tropical heating distribution on the extratropical transient eddy activities, and the feedback onto the extratropical mean flow within seasonal timescales. Hence, the change in extratropical circulation can be partly explained by the internal variability that is affected by the change in tropical heat source.

The data used in the present analysis are NCEP/NCAR (National Centers for Environmental Prediction/National Center for Atmospheric Research) reanalysis 2.5°×2.5° daily data for wind, geopotential height and NOAA (National Oceanographic and Atmospheric Administration) $2.5^\circ \times 2.5^\circ$ pentad data for outgoing longwave radiation (OLR) and optimal interpolation monthly sea surface temperature (OI SST). The climatology is made from 1979 to 1999.

2. Anomaly features

In this study, we select the winters of 1982/83 and 1988/89 as representatives of the El Niño and La Niña years, respectively. In 1982/83 (hereafter, El Niño), a warm SST anomaly was persistent for 16 months from April 1982 to July 1983 with the maximum in January, and in 1988/89 (hereafter, La Niña) for 14 months from May 1988 to June 1989 with the maximum in November. In this section, the features of the DJF (December, January, February) anomaly, deviations from the climatological mean (1979/80–1999/2000), serve as the background for the discussions to follow.

Figure 1 shows an anomaly field of SST, OLR, and 200 hPa geopotential height (H200) and zonal wind (U200). Shadings present large anomalies over one and a half times the standard deviation. Significant contrast between the two events can be seen over the tropical eastern Pacific (Figs. 1a and 1b), where the strongest warm (cold) SST anomaly appears in El Niño (La Niña). There are other contrasting features over the Indian Ocean and the North Pacific which show positive and negative anomalies of opposite signs between the two events. Tropical Pacific OLR anomalies (Figs. 1c and 1d) exhibit a simple wavenumber 1 pattern in both event years. During a warm event, deep tropical convection is suppressed over the maritime continent and enhanced over the equatorial central to eastern Pacific. In a cold winter, active convection occurs over the western Pacific, and the anomaly center is located at 10°N. Hence, the ITCZ and SPCZ are shifted equatorward and eastward from their climatological positions during El Niño winter and poleward and westward during La Niña winter. Over the subtropical eastern North Pacific, there is a positive (negative) OLR anomaly, and this seems to result from the downward (upward) compensating flow by the enhanced (reduced) heating at the tropics in the warm (cold) winter. Previous studies have argued that this anomaly is caused by the changes in the transient wave

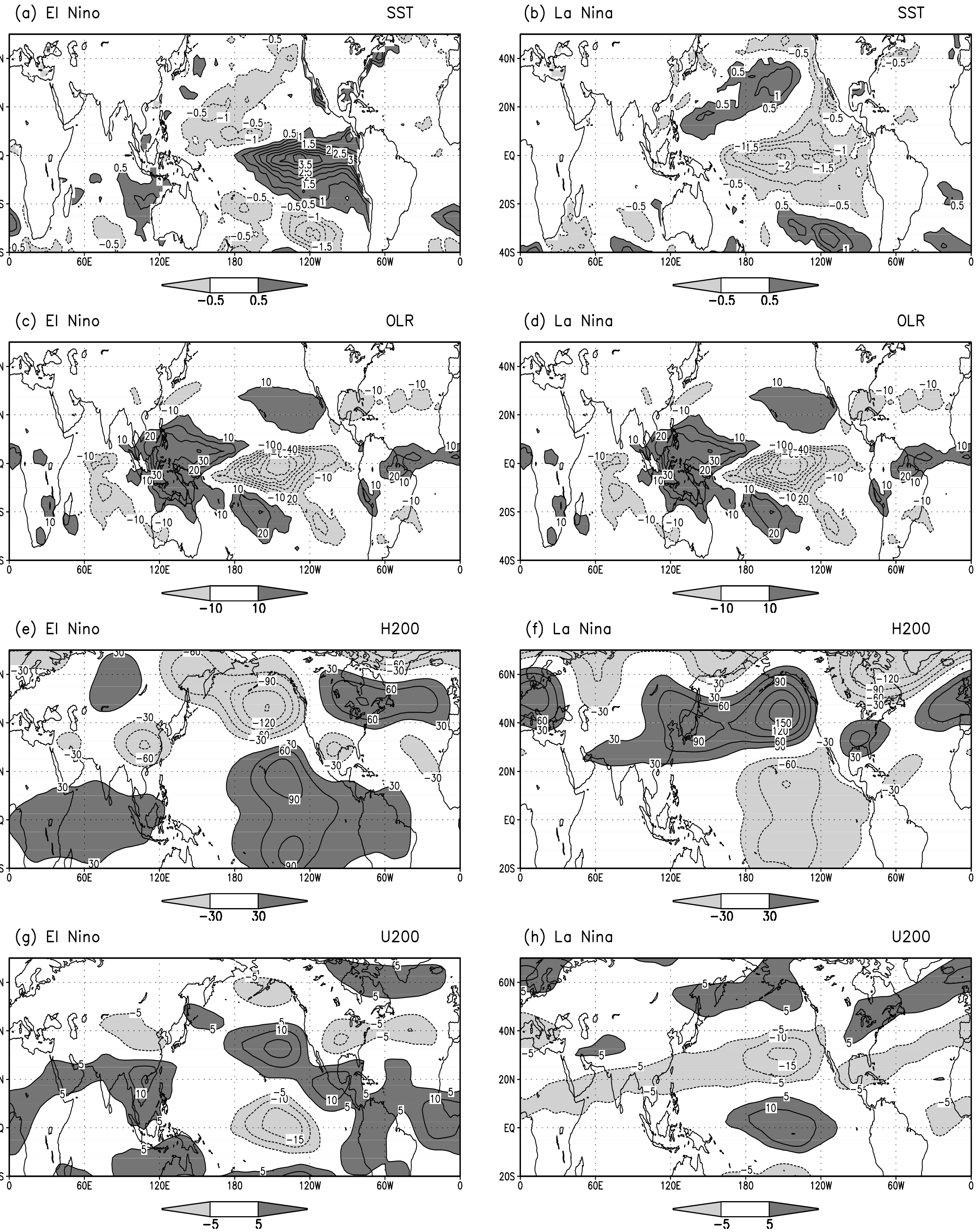

Fig. 1. Composite anomaly in El Niño and La Niña of (a) (b) SST, (c) (d) OLR, (e) (f) H200, (g) (h) U200. Contour intervals are (a) (b) 0.5°C and (c) (d) 10 W m^{-2}, (e) (f) 30 gpm, (g) (h) 5 m s^{-1}. The values in excess of (a) (b) ±0.5°C, (c) (d) 10 W m^{-2}, (e) (f) 30 gpm, and (g)(h) 5 m s^{-1}, which are over one and a half times the standard deviation are shaded.

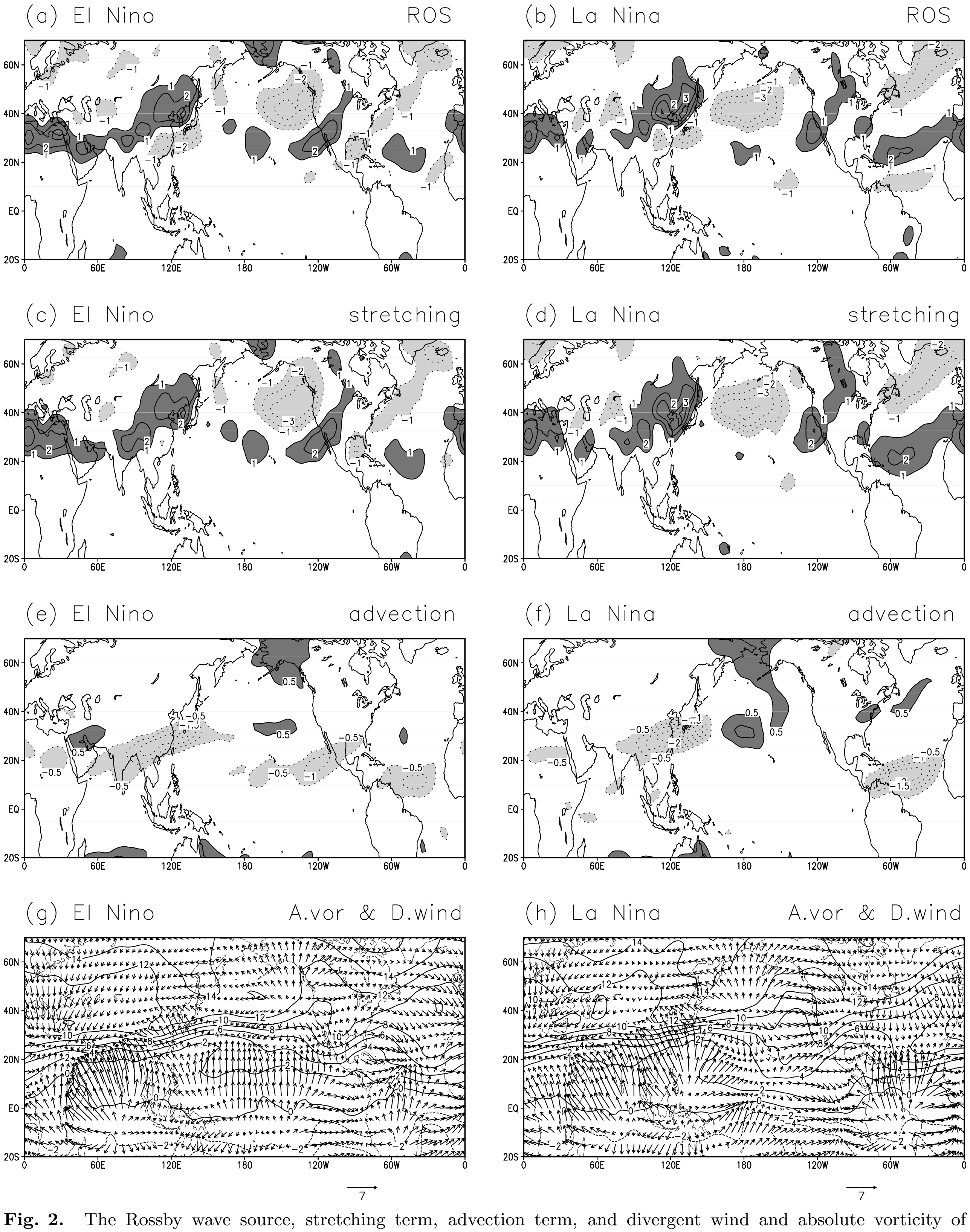

Fig. 2. The Rossby wave source, stretching term, advection term, and divergent wind and absolute vorticity of (a)(c)(e)(g) El Niño, (b)(d)(f)(h) La Niña, and (i)(j)(k)(l) El Niño–La Niña, respectively. Contour intervals are (a)–(d), (i)–(j) 1.0×10^{-10} s^{-2}, (e)–(f), (k) 0.5×10^{-10} s^{-2}, and (g)–(h), (l) 2×10^{-5} s^{-1}. The values in excess of (a)–(d), (i)–(j) $\pm1.0\times10^{-10}$ s^{-2}, (e)–(f), (k) $\pm0.5\times10^{-10}$ s^{-2} are shaded. The unit vector is shown on the lower right-hand side.

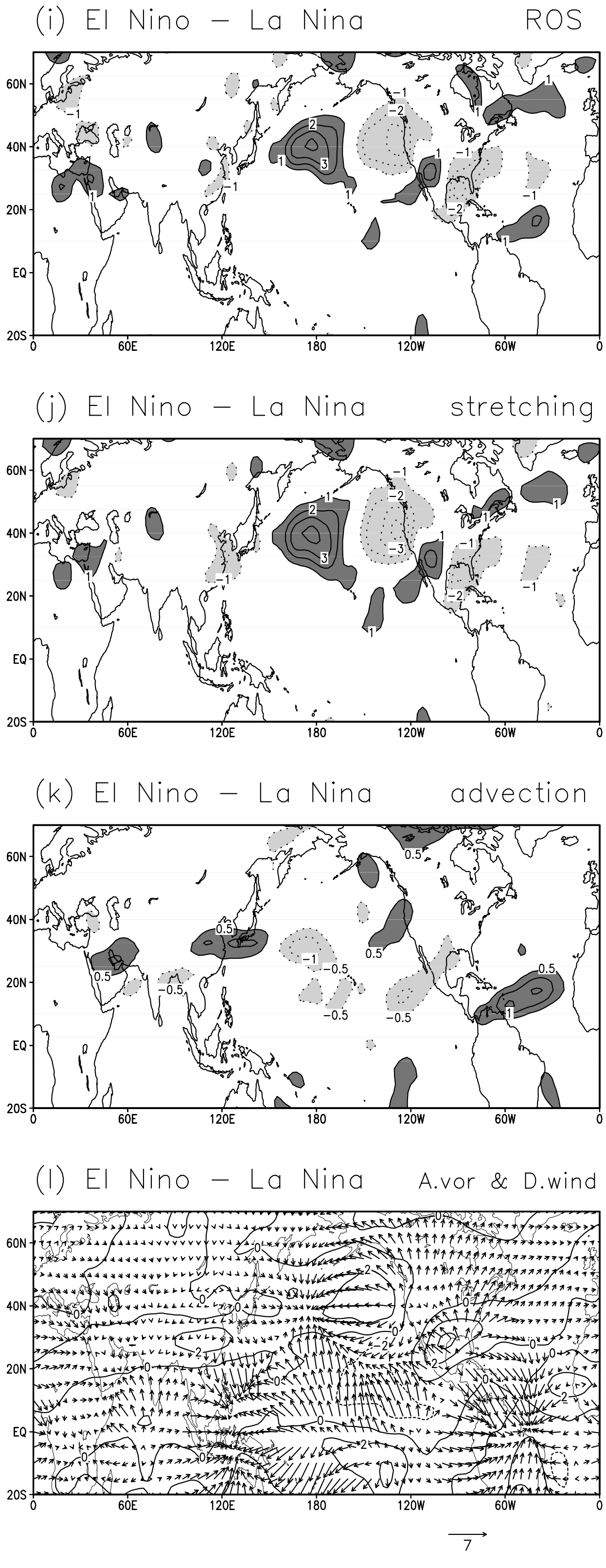

Fig. 2 (Continued)

and convection by tropical forcing activity (Lau and Boyle, 1987; Mattews and Kiladis, 1999).

A sharper difference in the time-mean response during boreal winter extratropical circulation to El Niño and La Niña forcing can be seen from the anomaly shown in Figs. 1e–1h. The anomalies associated with ENSO tend to have opposite polarities with the well-known PNA pattern, and opposite phases from the tropics to high latitudes (Figs. 1e, 1f). In an El Niño year, the negative anomalies over East Asia and the North Pacific indicate an intensified trough and shrunken ridge, and the positive anomalies over the eastern United States indicate a weakened trough. In a La Niña year, the signs are opposite. In a warm winter (Fig. 1g), positive and negative anomalies in the eastern Pacific represent the extension of the subtropical jet exceeding 30 m s^{-1} to North America and the tropical easterly far eastward to 150°W. During a cold winter (Fig. 1h), similar patterns can be found with the sign reversed. This means that a distinct break occurs in the subtropical jet over the central Pacific along 30°N and the region of the tropical easterly retracts westward to the dateline. The most dominant feature of U200 is the wind anomaly over the EAJ exit region over the extratropics. Anomaly features in El Niño (La Niña) will be simply referred to as stronger (weaker) wind due to an enhanced (reduced) thermal contrast in the Asian jet exit region.

3. Tropical-extratropical interaction

3.1 *Rossby wave source*

Since the kinetic energy for the mid-latitude flow has a peak near 250 hPa, one expects there to be an equivalent barotropic level which is perhaps the most appropriate one for studying the tropical-extratropical interaction with a one-level model (Sardeshmukh and Hoskins, 1988). This study selects 200 hPa to investigate the Rossby wave source using a one-level vorticity equation model in a barotropic sense. Following Sardeshmukh and Hoskins (1985), the nonlinear vorticity equation at this level can be written as

$$\frac{\partial \zeta}{\partial t} + \boldsymbol{v} \cdot \bigtriangledown \zeta = -\zeta D$$

$$(\boldsymbol{v} = \boldsymbol{v}_\psi + \boldsymbol{v}_\chi\ ,\ \ \boldsymbol{v}_\psi = \kappa \times \bigtriangledown^\psi\ ,\ \ \boldsymbol{v}_\chi = \bigtriangledown \chi)\,. \qquad (1)$$

Here, ζ is the absolute vorticity, D is the horizontal divergence, and the velocity field is decomposed into a purely rotational part ($\boldsymbol{v}_\psi$) and a purely divergent part ($\boldsymbol{v}_\chi$). Then equation (1) becomes,

$$\frac{\partial \zeta}{\partial t} + \boldsymbol{v}_\psi \cdot \bigtriangledown \zeta = -\zeta D(\text{Streching term})$$
$$- \boldsymbol{v}_\chi \cdot \bigtriangledown \zeta(\text{Advection term})\,. \qquad (2)$$

Equation (2) represents the correct partitioning between Rossby wave propagation terms, which involves just the rotational part of the wind field on its left-hand side, and the forcing terms, involving the divergent part of the wind, on the right-hand side. The propagation of Rossby waves is a result of advection of the absolute vorticity by the rotational, not the divergent part, of the wind field. The forcing term is referred to simply as the "Rossby wave source term (or simply as S)", which may be reduced to the compact form (James, 1994):

$$S = -\nabla \cdot (\boldsymbol{v}_\chi \zeta) \,. \tag{3}$$

Upper level divergent wind in the tropics may play a role to transfer the Rossby wave to mid-latitudes from the tropical convection region. The Coriolis torque acts to make anticyclonic circulation with divergent flow, and this anticyclone has a role to propagate the Rossby wave. Eventually the vorticity anomaly in the extratropical latitudes will be enhanced, even though the vorticity is not significant in the tropics (Sardeshmukh and Hoskins, 1988; Rasmusson and Mo, 1993).

Figure 2 represents the Rossby wave source, stretching term, advection term, and absolute vorticity and divergent wind during the El Niño and La Niña years, with the difference between the two years. During boreal winter, the heating has a large maximum over the Indonesian region where ζ and $\nabla\zeta$ are small. However, the divergent wind is strongest north of this region around the southeast coast of Asia. As the Rossby wave source is remote from the heating region, it can be considered as the effect of a steady localized divergent perturbation on a basic flow. This effect belongs to the teleconnection between the tropics and extratropics in a barotropic sense. The Rossby wave source (S) has a large maximum in the subtropical westerlies, such as the African Jet, East Asian Jet, and North American Jet, in both years, where the stationary Rossby waves can be efficiently generated. The largest Rossby wave source appears over East Asia, which is the region at the downward branch of the local Hadley circulation, and does not change its location between the two years. This may be due to the persistent divergent flow from the western tropical Pacific where the SST anomaly change is below 0.5°C (Figs. 1a and 1b). In Fig. 2i, the contrast between the two years appears to the east of 150°E, where the main difference of the tropical SST anomaly exists. The Rossby wave sink (negative) over the North Pacific moves eastward (westward) during El Niño (La Niña) years compared to normal years, which seems to be the effect of the longitudinal shift in the tropical forcing. According to Higgins and Mo (1997), this movement of the sink can be related with the extension and retraction of the EAJ exit. They have argued that the subtropical Rossby wave source anomalies combine with mid-latitude Rossby wave source anomalies to form a quadrupole pattern consistent with the jet retraction (extension) in the jet exit region. In this study, a similar result is obtained in the seasonal mean field by comparing El Niño and La Niña conditions. It can be confirmed from the zonal wind (contour line) at 200 hPa in Figs. 3b and 3c that the exit region of EAJ is consistent with the movement of the Rossby wave sink in Figs. 2a and 2b, respectively.

The stretching and advection terms, which constitute the Rossby wave source, have different characteristics at different latitudes. The stretching term (Figs. 2c–2d) is large mainly in the extratropics. The advection term is significant in the subtropics as the gradient vorticity and divergent wind becomes larger approaching the subtropics away from the heating maximum, roughly 10 degrees south of the stretching term. In Figs. 2e–2f, negative anomalies (sinks) show significant contrast between the two years. In an El Niño year, the sink region appears over 20°–30°N in the western Pacific, 10°–20°N in the eastern Pacific, and at the north of the South Atlantic Convergence Zone (SACZ). In a La Niña year, the sink region at the eastern Pacific disappears by the cooling of the SST anomaly and intensifies at the western Pacific and north of the SACZ. This suggests that the advection of absolute vorticity by the divergent wind is directly connected with the magnitude and location of the heating anomalies over the tropics.

3.2 *Mean-eddy interaction*

The response of the time-mean flow to external forcing has been studied in various ways and has been well documented (Opsteegh and Van den Dool, 1980; Hoskins and Karoly, 1981; Lau and Boyle, 1987; Ting and Held, 1990). Recently, several studies have dealt with the effect of tropical forcing on the internal variability. Chen and Van den Dool (1997) explained the possible reason for the large difference in the magnitude of the low-frequency variability over the eastern North Pacific between El Niño and La Niña years by using the local barotropic energy conversion term. They found that low-frequency components extract energy from the time-mean flow during La Niña winters, and lose energy to the time-mean flow in El Niño winters over the eastern North Pacific. This dynamical process has contributed significantly to the larger magnitude of low-frequency variability during La Niña years.

This section intends to find out the impact of tropical Pacific SST anomalies on the extratropical atmospheric internal variability, and to examine the contrasting features of the connection between the extratropical circulation and atmospheric internal variability during the two event years.

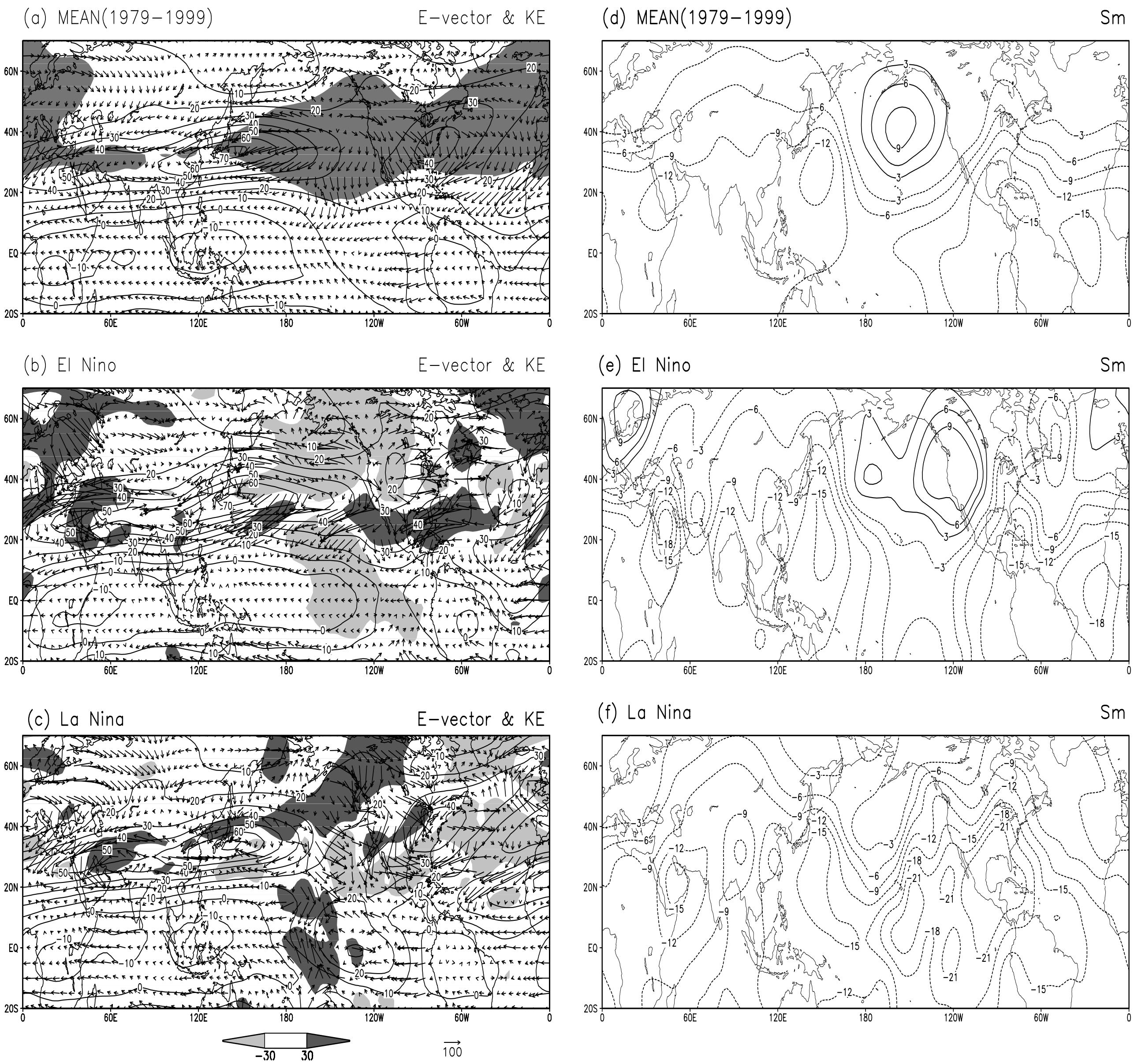

Fig. 3. E-vector and zonal wind at 200 hPa during (a) MEAN (1979–1999), (b) El Niño, and (c) La Niña; $S_{\rm m}$ of (d) MEAN (1979–1999), (e) El Niño, and (f) La Niña. Contour intervals are (a)(b)(c) 10 m s^{-1}, (d)(e)(f) 3.0×10^7m^3 s^{-2}. The values in excess of 150 m^2 s^{-2} are shaded. The unit vector is shown on the lower right-hand side in (a)–(c).

3.2.1 *E-vector and kinetic energy*

To investigate the role of mean-eddy interaction on the extratropical circulation by the longitudinal change of tropical heating, the E-vector, which is a very useful quantity for diagnosing the nature of the mean-eddy interaction, has been used. The E-vector can give information about the shape and propagation of eddies and the feedback of eddies onto the mean flow. Following Hoskins et al. (1983), the E-vector is defined to be,

$$\boldsymbol{E} = -(\overline{u'^2 - v'^2}, \overline{u'v'}) \,, \tag{4}$$

where, “ $'$ ” is the daily deviation from the seasonal (90-day) mean and “ $^-$ ” is the seasonal (90-day) mean. In order to emphasize large-scale structure, the inverse Laplacian of the E-vector is used and depicted as “$S_{\rm m}$” in equation (5).

$$S_{\rm m} = [\bigtriangledown^{-2}(-\bigtriangledown \cdot \boldsymbol{E})] \,. \tag{5}$$

In the region of $S_{\rm m}$ maxima (minima), there is a tendency for westerly (easterly) acceleration of the mean flow by the eddies.

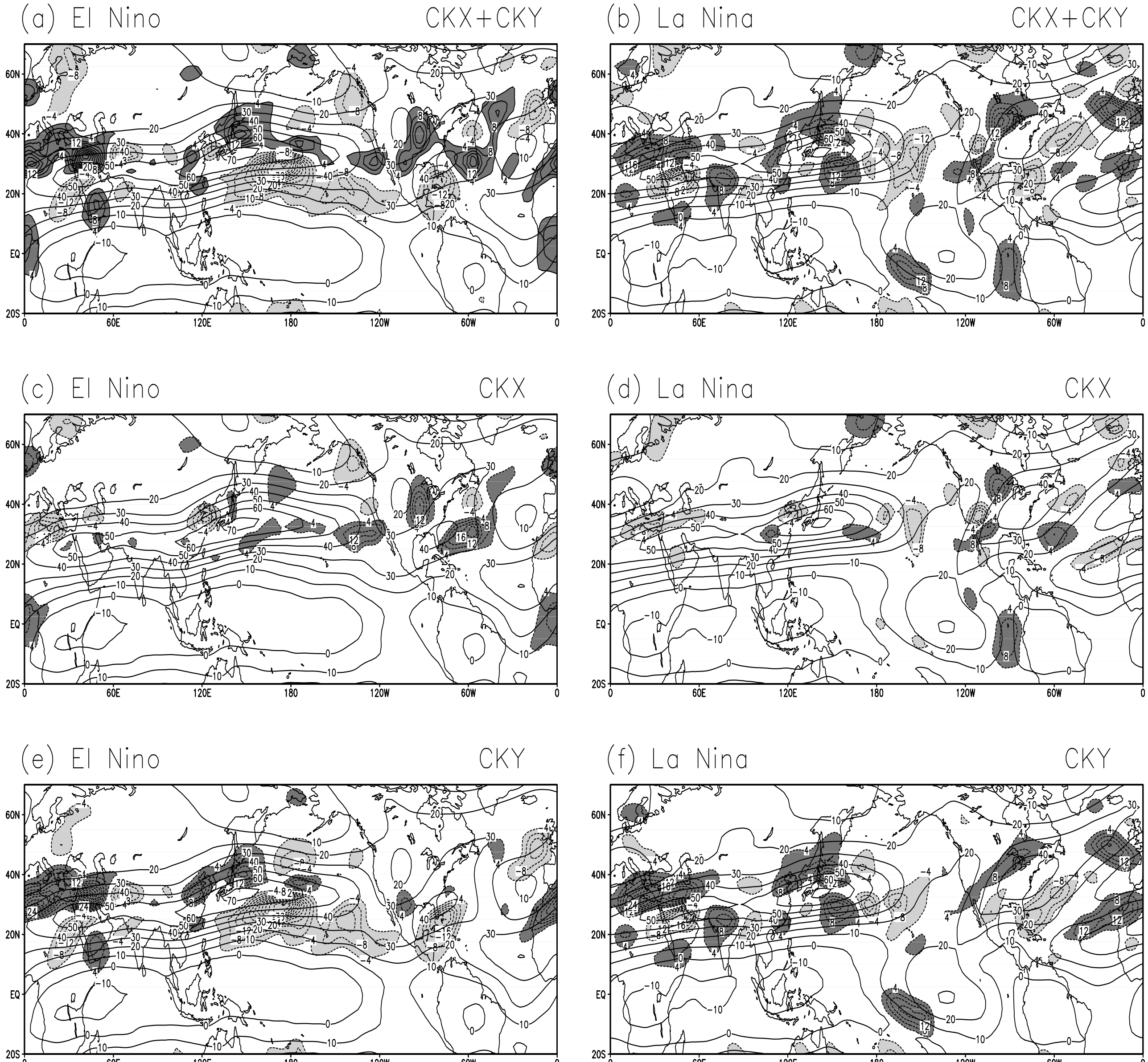

Fig. 4. CKX+CKY, CKX, and CKY in (a) (c) (e) El Niño, (b) (d) (f) La Niña, and (g) (h) (i) El Niño–La Niña, respectively. Contour (zonal wind at 200 hPa) interval is 10 m s^{-1}. The values in excess of $\pm 4.0\times10^{-4}$ m^2 s^{-3} are shaded.

Figure 3 shows the composite mean of the E-vector and kinetic energy with zonal wind and S_{m} at 200 hPa during the two event years. Figure 3a is the climatological DJF mean feature of the E-vector and kinetic energy and zonal wind. In the Northern Hemisphere, the E-vector tends to be easterly in most of the region, except for the storm track area over the North Pacific and North Atlantic. According to Hoskins and Karoly (1981), low-pass transients (periods greater than 10 days) are predominantly elongated zonally (easterly E-vector) and high-pass eddies (periods shorter than 10 days) are mostly elongated meridionally (westerly E-vector). Hence, at the downstream of the EAJ, eddy activities are nearly dominated by the low-frequency transients, and at the east of the EAJ exit, high-frequency eddy dominate. At the EAJ maximum, there is strong convergence of the E-vector near 150°E, and a strong divergence further downstream near 150°W. The corresponding eddy-mean flow feedback can be interpreted as an intense easterly mean flow forcing near 150°E, where the westerly is strong, and the westerly mean flow forcing

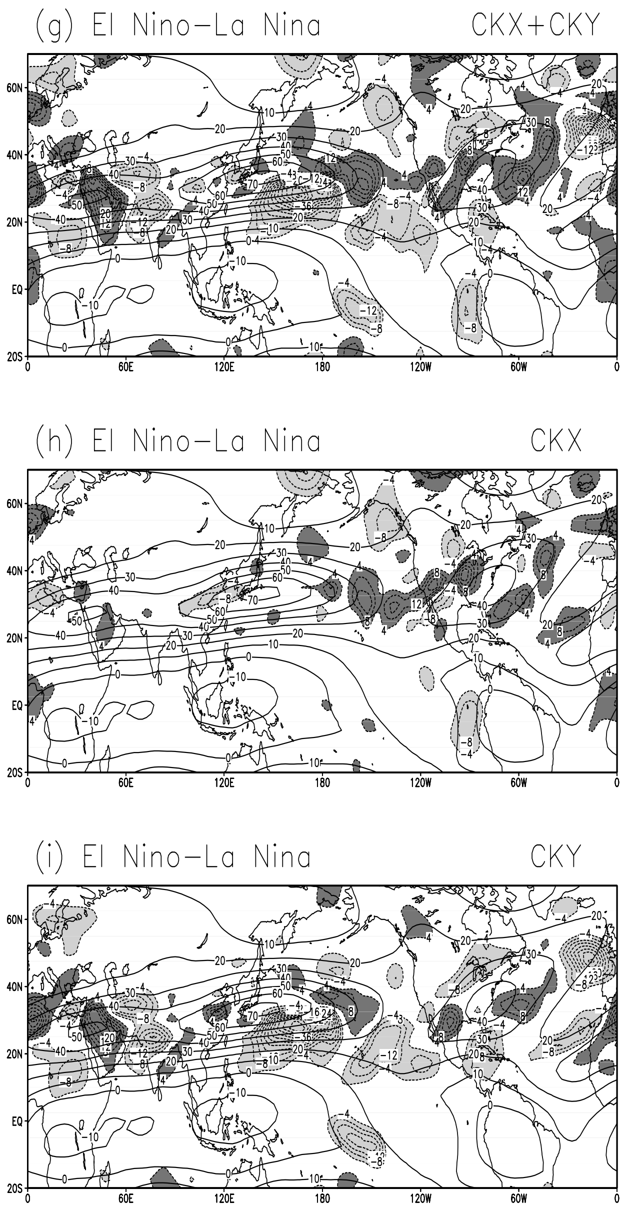

Fig. 4 (Continued)

to the downstream, where the westerly is weak. In the region of the meridionally elongated energy vector in 160°–100°W, the E-vector points equatorward which is related with poleward eddy momentum flux that plays to feed energy from the eddy to the westerly mean flow in the region east of the dateline.

During an El Niño year (Fig. 3b), the easterly E-vector extends far eastward, which plays to feed energy from the eddy to the westerly mean flow in the region east of the dateline. The E-vector that points from the extratropics to the equator (Fig. 3a) has almost disappeared. These features can be confirmed in $S_{\rm m}$ of Figs. 3d–3e. The dipole structure of the maximum center over the downstream region of EAJ and minimum center to the south of the EAJ maximum can be seen. Compared with the result from the climatological mean, the maximum (divergence of E-vector) $S_{\rm m}$ center has moved to the east, indicating that the region of westerly forcing of the mean flow by the eddies has shifted eastward. In a La Niña year (Fig. 3c), the area of easterly E-vector almost disappears and the equatorward energy vector at the southeast side of the EAJ and poleward energy vector to the northeast of the EAJ strengthens. In the large-structure of the E-vector (Fig. 3f) during a cold-event, the positive $S_{\rm m}$ disappears, the minimum center at the EAJ core intensifies, and the center to the east of Mexico extends to the dateline. From Figs. 3d to 3f, it is found that the region of maximum $S_{\rm m}$ is sensitive to the tropical forcing region, and the minimum $S_{\rm m}$ varies in magnitude.

To obtain a clear picture of the boreal eddy behavior, the eddy kinetic energy $K_{\rm E}=\overline{(u'^2+v'^2)}/2$ is compared in Figs. 3b and 3c in shading with the climatological mean. Large eddy kinetic energy appears in the downstream region of the westerly jet and over the region of the meridionally elongated E-vector. The two event years show opposite phases over the North Pacific and North Atlantic in their anomaly fields (Figs. 3b and 3c). While the eddy kinetic energy is developed far east ward compared with the region of normal westerly mean flow in the El Niño year, it is distributed at the meridional boundary of the EAJ exit during the La Niña year, indicating the eddy activity is dominated on the eastside of the EAJ exit in the warm event year, and on north and south sides in the cold event year due to the close connection with the extension and retraction of the mean flow.

3.2.2 *Barotropic local growth of energy*

In this section, mean-eddy energy conversion is investigated following Simmons et al. (1983), Hoskins et al. (1983), and Yu and Huang (1995) by comparing the El Niño and La Niña years. According to Yu and Huang (1995), time evolution of eddy kinetic energy with two components for zonal and meridional energy conversions can be represented as follows.

$$
\begin{aligned}
\frac{\partial K_{\rm E}}{\partial t} = \quad & -\frac{\overline{(u'^2-v'^2)}}{a}\left[\frac{1}{\cos\Phi}\frac{\partial\overline{u_b}}{\partial\lambda}-\overline{v_b}\tan\Phi\right] \\
& -\frac{\overline{u'v'}}{a}\left[\cos\Phi\frac{\partial}{\partial\Phi}\left(\frac{\overline{u_b}}{\cos\Phi}\right)+\frac{1}{\cos\Phi}\frac{\partial\overline{v_b}}{\partial\lambda}\right] \\
\approx \quad & -\overline{(u'^2-v'^2)}\left(\frac{1}{\cos\Phi}\frac{\partial\overline{u_b}}{\partial\lambda}\right) \\
& -\overline{u'v'}\left(\frac{\partial\overline{u_b}}{a\partial\Phi}+\frac{\overline{u_b}\tan\Phi}{a}\right)\,.
\end{aligned}
\tag{6}
$$

Hence,

$$\begin{cases} \mathrm{CKY} = -\overline{(u'^2 - v'^2)}\left(\dfrac{1}{\cos\Phi}\dfrac{\partial \overline{u_b}}{\partial\lambda}\right), \\ \mathrm{CKY} = -\overline{(u'v')}\left(\dfrac{\partial \overline{u_b}}{a\partial\Phi} + \dfrac{\overline{u_b}\tan\Phi}{a}\right), \end{cases} \tag{7}$$

where, $\overline{u_b}$ and $\overline{v_b}$ are the seasonal (90-day) mean zonal and meridional component wind in each year and u' and v' are daily deviations from the seasonal mean, respectively. CKX (CKY) is associated with the effect of the zonal inhomogeneity (meridional shear) of the time-mean flow. Hence, a generation of eddy energy is implied by zonally elongated eddies ($u'^2 > v'^2$) in regions of confluence ($\partial u_b/\partial x < 0$), and by meridionally elongated eddies ($v'^2 > u'^2$) in regions of diffluence ($\partial u_b/\partial x > 0$).

Figure 4 shows the growth of kinetic energy in the El Niño and La Niña years, and the difference between them. A positive conversion of energy means that the kinetic energy of perturbation gets supply from the basic flow so that barotropical instability will develop there. In Figs. 4a and 4b, there are large positive/negative pairs over the African Jet (AJ), EAJ, and NAJ maxima, in which the positive anomaly is to the north and the negative anomaly is to the south. Therefore, the region just north of the AJ, EAJ, and NAJ is the major energy source for forcing the barotropic eddies. To compare CKX and CKY, each term is shown in Fig. 4. The CKX term, involving the zonal variation of the basic state, confirms that the dominant contributions to this term come from the regions of entrance and exit in the EAJ. Hence, the eddy energy becomes larger at the exit of the EAJ where the diffluent flow is dominant, and this feature shows some differences between the two event years. That is, the eddy energy growth extends far east ward in the El Niño year, while it almost disappears in the La Niña year. The CKY term shows a similar pattern as CKX+CKY, indicating that eddy energy conversion is more dependent on the barotropically unstable effect of the meridional shear of the zonal flow. In this case, the eddy energy growth region exists to the north of the jet core, which does not change its location throughout the years, however, the region of losing eddy energy, which means the basic flow gets energy from the eddies, is seen to become large over the southeast side of the EAJ exit, and extend to the NAJ in the El Niño year and becomes very weak and retracted in the La Niña year.

From the barotropic growth of eddy energy, it is found that CKX, which is the term of energy conversion by the zonally inhomogeneous flow shear and describes an additional energy transfer between the basic state and the perturbation, was located at the exit of the EAJ and NAJ, and showed some differences between the two event years in the extension/retraction of the positive area. The CKY term, with eddy energy growing when the meridional eddy transfer of zonal momentum $(u'v')$ is in the direction of a weakening westerly, seems to contribute mainly in determining the barotropic growth of kinetic energy. A striking contrast appears to the southeast of the EAJ with a negative sign, where much mean flow energy is obtained during El Niño years. This seems to affect the mean flow to extend more eastward during El Niño years.

4. Summary

The association between tropical convection and boreal extratropical circulation is studied in order to enhance our understanding of winter circulation variability in mid-latitudes and its interaction with ENSO. First, the heating and circulation anomalies during the El Niño and La Niña years revealed remarkably contrary features. Tropical convection becomes stronger (weaker) over the eastern Pacific during the El Niño (La Niña) year. In relation to the tropical variability, the EAJ exit shows extension (retraction) to the east (west) during the El Niño (La Niña) year.

Second, the contrasting features of the extratropical response due to the tropical forcing are examined by using several methods, such as Rossby wave source function, transient eddy flux, and mean-eddy kinetic energy conversion. In the El Niño year, upper-level divergence intensifies over the eastern tropical Pacific and the absolute vorticity gradient increases mainly due to the effect of enhanced convection during the warm event year. This produces another sink to the east of the dateline over the subtropics and an eastward shift of the sink region over the North Pacific. In the extratropics, the Rossby wave source forms over East Asia. This source appears in all years due to the persistent divergence flow from the western Pacific warm pool area. In the La Niña year, the Rossby wave sink over the North Pacific moves west. Hence, the contrasting results in both years can explain the sensitive response of the extratropics by the effect of the longitudinal change in tropical forcing.

The E-vector is used in order to investigate the mean-eddy feedback. The contrasting feature between the two events in the E-vector appears downstream of the EAJ. During the El Niño year, the easterly E-vector intensifies and extends from the EAJ maximum to the far eastern Pacific, which seems to play a key role in extending the mean flow far eastward. In the La Niña year, the easterly E-vector changes its direction to westerly, which indicates the strengthening of meridionally elongated eddies in the storm track areas. From the barotropic growth of eddy energy, it is found that CKX, which is related with zonally inhomogeneous flow, is located at the exit of the EAJ and NAJ. The location of the eddy energy growth is found to be sensitive to the EAJ behavior identical to the movement of the tropical heat source. The CKY term, mostly from the meridional shear of the mean flow, contributes mainly to determining the barotropic growth of kinetic energy. A remarkable contrast between the two event years appears with negative sign on the southeast side of the EAJ. Much more meanflow energy is obtained from the eddy energy in El Niño than La Niña, which seems to play an important role in enhancing the meanflow energy there. This shows that the mean-eddy interaction over the extratropics acts in secondary role to influence the extratropical circulation by the change of westerly flow, which seems to be affected by the tropical heating far away.

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