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ORIGINAL PAPER

Signature of the Antarctic oscillation in the northern hemisphere

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Abstract Using the ECWMF daily reanalysis data, this paper investigates signatures of the Antarctic Oscillation (AAO) in the upper troposphere of the northern hemisphere. It is found that during boreal winter, a positive (negative) phase of the AAO is associated with anomalous easterlies (westerlies) in middle-low latitudes (∼30–40°N) and anomalous westerlies (easterlies) in middle-high latitudes (∼45–65°N) of the upper troposphere about 25–40 days later. While there is also a response in zonal wind in the tropics, namely over the central-eastern Pacific, to some extent, these tropical zonal wind anomalies can trigger a Pacific/North American teleconnection patterns (PNA)-like quasi-stationary Rossby waves that propagate into the Northern Hemisphere and gradually evolve into patterns which resemble North Atlantic teleconnection patterns. Furthermore, these quasi-stationary Rossby waves might give rise to anomalous eddy momentum flux convergence and divergence to accelerate anomalous zonal winds in the Northern Hemisphere.

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1 Introduction

The well-known leading modes of extratropical atmospheric variability in the southern hemisphere (SH) is characterized by a near zonally symmetric seesaw in the geopotential height between the polar cap and the surrounding zonal ring along ∼45°S and equivalent barotropic vacillations in the extratropical zonal wind fields between high (center of action located at ∼60°S) and middle latitudes (center of action located at ∼40°S), which is called the Antarctic Oscillation (AAO, Kidson 1988a, b; Gong and Wang 1999). The positive index polarity of the AAO refers to anomalous westerlies along high latitudes with anomalous low-geopotential height in polar cap, and vice versa. Sometimes the "AAO" is alternatively referred to as the "SAM" (Southern hemisphere Annular Mode, Thompson and Wallace 2000). For easy reference, only "AAO" will be used in this paper hereafter.

Although the fundamental dynamical processes that govern the amplitude, meridional scale, and zonal structure of the AAO are not well understood, many studies indicated that variability of the AAO is driven by wave-mean flow interactions in the extratropical circulation (e.g. Hartmann and Lo 1998; Limpasuvan and Hartmann 1999, 2000; Rashid and Simmonds 2004). Results of power spectrum analyses show that the temporal variability of the AAO has no obvious peak period and is indistinguishable from the red noise (Hartmann and Lo 1998). Feldstein and Lee (1998) and Lorenz and Harmann (2001) argued that the fact the AAO exhibits a substantial amount of variance at the low frequencies can be explained by the positive feedback between high-frequency eddies and the mean flow. The long persistent extreme phase of the AAO might help us improve the skill of middle-range weather forecasting in the SH (e.g. Sinclair et al. 1997). The transition of the AAO between phases, however, is unpredictable and the dynamics of it are not fully understood. Efforts have been made to investigate this question and many studies show that the transitional events are mainly driven by eddy momentum flux anomalies and the eddy heat flux is less important (e.g. Lee and Feldstein 1996; Feldstein and Lee, 1996; Kidson and Watterson 1999, Rashid and Simmonds 2004). Shiogama et al. (2005) explored the effects of low- and high-frequency eddies on the transitional processes of the AAO. They found that the roles of low- and high-frequency eddies are different for the poleward (from negative to positive phase) and equatorward (from positive to negative phase) transitional processes. For the poleward transitional events, low-frequency eddies, preceding the high-frequency eddies, drive the transition, but for the equatorward transitional events, low- and high-frequency eddies simultaneously force the transition.

Previous studies have found that the AAO accounts for a large fraction of the extratropical troposphere variability in the SH (e.g., Thompson and Wallace 2000). Unquestionably, main signatures of the AAO are located in the SH extratropical region, but several studies had revealed that it also has a substantial influence on the SH subtropical/tropical latitudes which even extends into the subtropics of the opposing hemisphere as well. Thompson and Wallace (2000) reported that the AAO impacts on the strength of the trade winds in the SH subtropics. Baldwin (2001) pointed out that the AAO is associated with variations in sea level pressure, extending into the deep tropics. Thompson and Lorenz (2004) examined the signature of the AAO in the tropical troposphere. Their results showed that the AAO is related to upper-tropospheric zonal wind anomalies centered about the equator in the SH cold season months. Ho et al. (2005) found that the AAO can alter tropical cyclone activities over the western north Pacific during typhoon season.

Moreover, a number of studies proved that the northern winter atmosphere is very sensitive to tropical forcing such as anomalous sea surface temperature (SST) and tropical convection. The anomalies of the tropical troposphere dramatically influence the extratropical large-scale atmospheric circulation, especially in the Northern Hemisphere (NH, e.g. Horel and Wallace 1981; Blackmon et al. 1983; Branstator 1985; Lau and Phillips 1986; Lau and Boyle 1987; Trenberth et al. 1998). Therefore, an increasing body of evidences shows a substantial role of the AAO in the tropics and raises the question of whether the AAO can impact the NH through the tropics during boreal winter. If there are signatures of the AAO in the NH, how far and through what process does the AAO affect the NH? The objective of this study is to answer these questions.

Zhou and Yu (2004) examined the reproducibility of the AAO by using NCAR CAM2 model forced with historical SST. Their results showed that the reproducibility of the AAO variability is dependent on the tropical Pacific forcing. Codron (2005) investigated the relation between structure of the AAO and the mean state by using observation data and GCM outputs. His results revealed that a change of the background state such as the different phase of ENSO can modify the structure and the dynamics of the AAO. Carvalho et al. (2005) indicated that a negative (positive) phase of the AAO dominates when the 365-day low-pass SST anomalies resemble an El Niño (La Niña) phase of ENSO during austral summer (DJF). L'Heureux and Thompson (2006) pointed out that ENSO cycle can influence the AAO and about 25% of the annual variability in the AAO is linearly related to fluctuations of the ENSO cycle in austral summer. A primary objective of this study is to explore the possible influences of the AAO on the NH, so to avoid an artificial correlation between the AAO and circulation of the NH created by ENSO the influence of ENSO on the AAO should be removed. This paper is organized as follows: Section 2 describes the data and the methods used in this analysis. Observational relationships between the AAO and anomalous zonal winds of the NH are then discussed in Sect. 3. Section 4 gives possible mechanism responsible for the linkage between the AAO and the zonal wind anomalies. Discussion and summary are presented in Sect. 5.

2 Data and methodology

Data used in this study are daily fields of the European Centre for Medium-Range Weather Forecasts (ECMWF) ERA-40 reanalysis for the period of 1979–2002, which is archived on 23 pressure levels spanning from 1000 to 1 hPa, and the horizontal resolution is $2.5° \times 2.5°$ (Uppala et al. 2005). These daily data are available online via the ECMWF website (http://data.ecmwf.int/data/d/era40_daily). A normalized daily AAO index is downloaded from Climate Prediction Center (CPC) of the National Oceanic and Atmospheric Administration (NOAA, http://www.cpc.ncep.noaa.gov) for the period of 1979–2002. The daily AAO index is constructed by projecting the daily mean 700-hPa height anomalies at a horizontal resolution of $2.5° \times 2.5°$ onto the leading Empirical Orthogonal Function (EOF) mode of monthly mean 700-hPa height anomalies poleward of 20° latitude for the SH. A positive (negative) index polarity of the AAO refers to the stronger (weaker) than normal westerlies along $\sim$50°–75° latitudes in the SH. This index is referred to as the AAOI hereafter.

Because ENSO operate mainly on interannual-scale (e.g. Neelin et al. 1998) and because the variability of the AAO is associated with ENSO (L'Heureux and Thompson 2006), the influence of ENSO on the AAO is removed by using a 365-day high-pass Fast Fourier Transform (FFT) filter (e.g. Bloomfield 2000). This is applied to all daily

ERA-40 reanalysis from September 1979 to August 2002, from which the northern cold season (October–April) data is extracted. In this analysis, we use the daily AAOI in the boreal winter months of December–February from 1979/1980–2001/2002, corresponding to 23 winters or 2,070 days. For convenience, 29 February in each leap year is removed from the daily ERA-40 data. At each grid point, the seasonal cycle which is defined as the time mean of each calendar day is removed. In cases where we calculate lead-lag correlation or regression, the calendar dates used in the analysis extend from October to April. For example, when the correlation or regression lags by 31 days, the predictor (the AAO index) is based on 1 December–28 February and the predict and (Zonal wind or geopotential height) on 1 January—31 March (lagging 31 days).

In this study, the statistical significance of all correlation coefficients is established by using the t statistic $t = \frac{r\sqrt{N_{\mathrm{dof}}-2}}{\sqrt{1-r^2}}$. As the AAO index has a high autocorrelation, the effective number of degrees of freedom for two time series was evaluated by using $N_{\mathrm{dof}} = N\frac{1-r_1 r_2}{1+r_1 r_2}$ where N_{dof} is the number of effective sample size; N is the sample size; and r_1, r_2 are the lag-one autocorrelation for the two time series considered (Bretherton et al. 1999). As the analyses are linear, we describe the results in terms of a positive index polarity only.

3 Linkages between the AAO and the zonal mean zonal wind anomalies in the NH

The zonal mean zonal wind anomalies in the SH is closely related to the AAO, and it is convenient to examine the linkages between the AAO and the circulation anomalies in the NH by using the zonal mean zonal wind responses. Therefore, in this section, the relationships between the AAO and the zonal mean zonal wind anomalies in the NH are examined in detail. As shown in Fig. 1, daily global zonal mean zonal wind anomalies at 300-hPa are regressed on DJF daily AAOI for lag ranging from −60 to +60 days (lags 0 days means contemporary regression, lags -*x* days means the zonal wind leads the AAOI by *x* days, lags + *x* days means the zonal wind lags the AAOI by *x* days). There are three features in Fig. 1 we want to highlight. First, the most prominent feature of Fig. 1 is anomalous zonal wind relating to a positive AAO pattern in the middle-high latitudes of the SH from day −20 to day +15. Second, from day +25 to day +40, there are notable statistically significant anomalous zonal wind located at ∼30°–40°N and ∼45°–65°N, suggesting that the AAO is associated with easterly (westerly) anomalies in the NH middle-low latitudes and westerly (easterly) anomalies in the NH middle-high latitudes during a positive (negative) index polarity of the AAO. Thus, we speculate that the AAO can impact the atmospheric circulation of the opposing hemisphere about 25–40 days later. Third, there are weak westerlies (but not significant) associated with the AAO in the tropics from day −15 to day +15. Our ensuing analyses will show that these westerlies are important for connecting the AAO with the anomalous zonal winds in the NH as will be discussed in detail in Sect. 4.

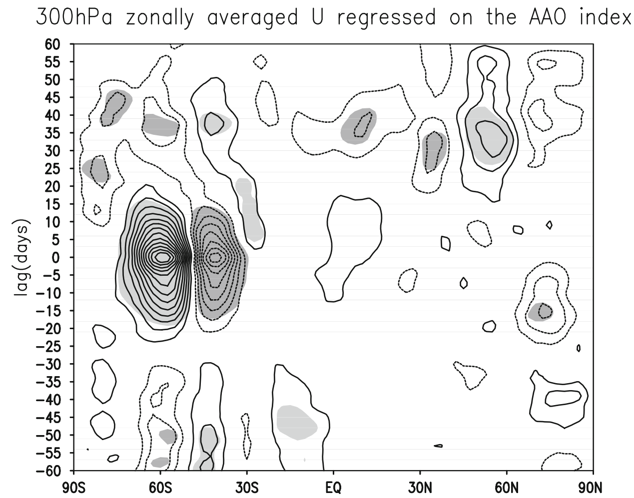

Fig. 1 Zonally averaged zonal wind anomalies at 300-hPa regressed onto daily values of the AAO index. *Positive lags* indicate the zonal wind lags the AAO index time series, and vice versa. *Shading* denotes correlations exceeding the 95% confidence level based on the *t*-statistic with an absolute value of. Contour intervals are 0.2 m/s, zero contours is omitted, *negative contours* are dashed. This result is evaluated for a positive AAO phase

Furthermore, we examine the vertical distribution of the zonal wind anomalies associated with the AAO. Figure 2 is the vertical distribution of zonal mean zonal winds from 1,000 to 1 hPa (23 levels) regressed to daily DJF values of the AAOI at lags 0 (top) and lags +35 days (bottom) for the global domain (90°S–90°N). At lags 0, the AAO pattern is significant from 1,000 to 10 hPa in the middle-high latitudes of the SH. And there are weak westerly anomalies in the upper tropical troposphere. However when the zonal mean zonal wind lags the AAO by 35 days, it is found that the positive AAO-like pattern disappears and changes into a negative AAO-like phase like pattern. The more interesting thing is the emerging westerly anomalies in ∼45°–65° latitudes of the NH. These westerly anomalies are deep, nearly barotropic, from the stratosphere extend to the surface, but only the tropospheric parts are statistically significant. There are also easterly anomalies in ∼30°–40° latitudes of the NH connected with easterly anomalies in the subtropics/tropics but basically confined to the troposphere (see Fig. 2 bottom).

It is well known that, in northern winter, ENSO can dramatically impact the extratropical circulation of both hemispheres. In fact, the NH zonal mean zonal wind anomalies attributed to the AAO which is described above

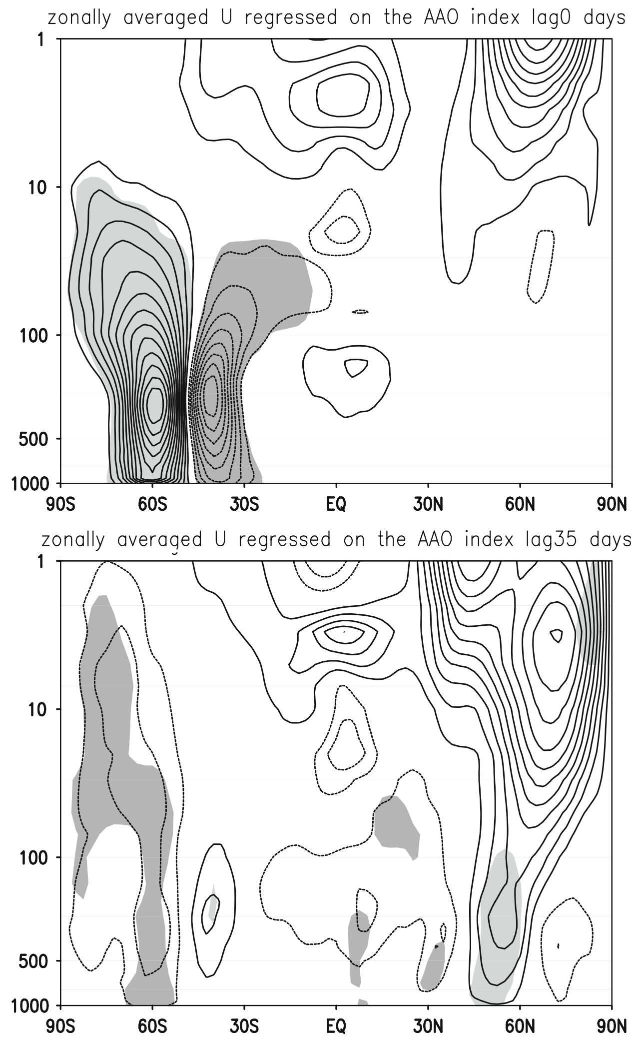

Fig. 2 Zonally averaged zonal wind regressed onto the daily AAO index. Lag 0 day (*top*), lag +35 days (*bottom*). *Shading* denotes correlations exceeding the 95% confidence level based on the *t*-statistic. Contour intervals are 0.2 m/s, zero contours are omitted, and *negative contours* are dashed. This result is evaluated for a positive AAO phase

resemble those associated with ENSO. Although the annual signal is removed by using a 365-day high-pass FFT filter, the possibility that ENSO produce the 'spurious' correlation between the AAO and the NH zonal mean zonal wind might not be completely eliminated. For consolidating the results, data in non-ENSO years[1] from 1979 to 2002 are analyzed with a duplicate methodology. We obtain similar results (not shown). Thus the results of the non-ENSO years confirm that the AAO is associated with anomalous zonal winds in the NH and this phenomenon does not relate to ENSO.

Combining Figs. 1 and 2, we find that the AAO pattern has a possible influence on tropical latitudes. To some extent the influences of the AAO can even extend into the NH middle-high latitudes as well. But the question is, through what mechanism does the AAO exert its impact on the circulation of the NH middle-high latitudes, and what the dynamics is for tropospheric anomalies in the NH observed in Figs. 1 and 2. In the following section we address these issues in detail.

4 Physical explanation

In this section, possible mechanism response for the linkages between the AAO and the NH zonal wind anomalies is presented. Our analyses show that variability of the AAO is associated with anomalous zonal wind in the upper troposphere over the central-eastern tropical Pacific (this will be described in the following paragraph). Anomalous northward propagating quasi-stationary Rossby waves that resemble the Pacific/North American (PNA) and the North Atlantic (NA) teleconnection patterns in the NH can be triggered by this anomalous zonal wind field in the tropics. The anomalous quasi-stationary Rossby waves can propagate into the NH middle-high latitudes and cause anomalous eddy momentum flux convergence (divergence) there, which leads to westerly (easterly) anomalies in the NH. To understand the dynamical mechanism of the tropospheric anomalies in the NH observed in Figs. 1 and 2 in more detail, we consider in turn: a. Anomalous zonal wind in upper troposphere over central-eastern tropical Pacific; b. Quasi-stationary Rossby waves and eddy momentum flux.

4.1 Anomalous zonal wind in upper troposphere over the central-eastern tropical Pacific

We regressed the global domain zonal wind anomalies at 200-hPa on daily values of the AAOI in northern winter (DJF) from lags −30 to +60 days. Three representative days, lags −5 days (top), 0 days (middle) and +5 days (bottom) are picked up and polted in Fig. 3. Obviously there is a positive AAO-like phase pattern over the extratropical area of the SH. An intriguing feature in Fig. 3 is the statistically significant westerly anomalies over the central-eastern tropical Pacific. And these westerly anomalies are consistent with the westerly in tropics that we mentioned in Figs. 1 and 2. It should be noticed that the zonal wind anomalies in tropics are wave like and

[1] The non-ENSO years are picked out based on a Cold Tongue Index (CTI). The monthly CTI defined as the area-averaged over (6°S–6°N, 180°–90°W) SST anomalies which reflect the ENSO cycle is downloaded from the Joint Institute for the Study of the Atmosphere and Ocean at the University of Washington (http://jisao.washington.edu). The criterion for non-ENSO years is half of a standard deviation (|CTI| < 0.5) of northern winter (DJF averaged) CTI. There are 12 non-ENSO years (79/80, 80/81, 81/82, 83/84, 85/86, 87/88, 89/90, 90/91, 92/93, 93/94, 96/97, 01/02) during the period of 1979/1981–2001/2002.

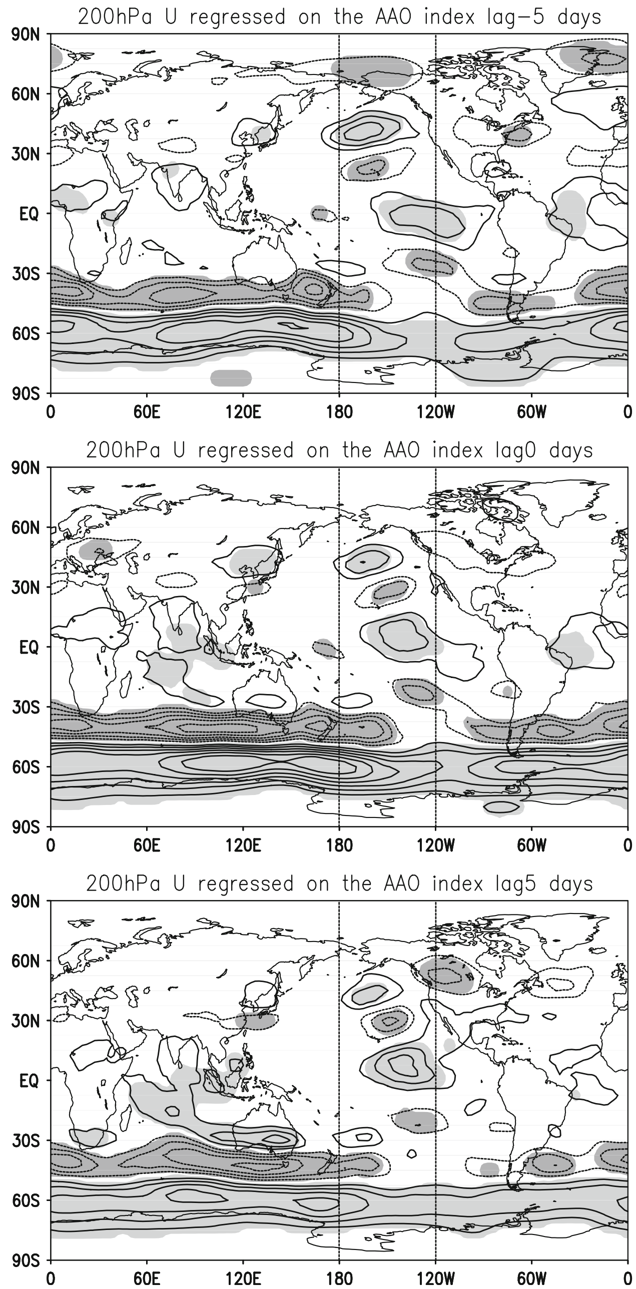

Fig. 3 Zonal wind anomalies at 200-hPa regressed on the daily AAO index at lags -5 days (*top*), lags 0 days (*middle*) and lag +5 days (*bottom*). *Shading* denotes correlations exceeding the 95% confidence level based on the *t*-statistic. Contour intervals are 0.5 m/s, zero contours are omitted, and *negative contours* are dashed. This result is evaluated for a positive AAO phase

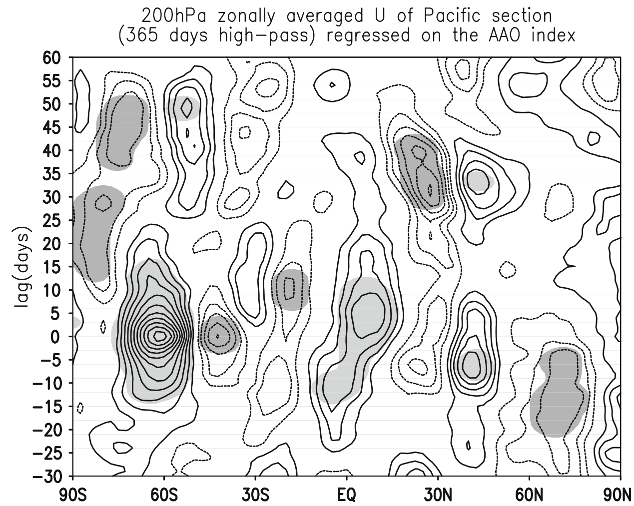

Fig. 4 200-hPa zonally averaged zonal wind anomalies over the central-eastern Pacific (180°–120°W) regressed on the daily DJF AAO index form lag −30 days to lag +60 days. *Shading* denotes correlations exceeding the 95% confidence level based on the *t*-statistic. Contour intervals are 0.2 m/s, zero contours are omitted, and *negative contours* are dashed. This result is evaluated for a positive AAO phase

confined in the area of tropical Pacific climatology westerly 'duct' region where the Rossby waves can propagate into the opposing hemispheres, as suggested by Webster and Holton (1982) The temporal evolution of zonal wind anomalies over the central-eastern tropical Pacific associated with variability of the AAO is also examined. A regression of 200-hPa daily zonally averaged zonal wind anomalies over the central-eastern Pacific (180°–120°W) onto DJF daily values of the AAOI for lags ranging from −30 to +60 days (Fig. 4), reveal that the zonal mean zonal wind anomalies located in the central-eastern tropical Pacific can persist from day −20 to day +20. And it has maximum regression (also the correlation) coefficients with the AAO, when the AAO is leading the zonal mean zonal wind by about +5 to +10 days. These results are consistent with the finding reported by Thompson and Lorenz (2004), which shows that annular modes are associated with upper-tropospheric zonal wind anomalies centered about the equator, and mainly located in the central-eastern tropical Pacific.

As suggested by Thompson and Lorenz (2004), the zonal wind anomalies in the tropics can be attributed to wave-mean flow interactions. For low-frequency Rossby waves, the meridional group velocity of the waves accelerates (decelerates) with the zonal wind (e.g., Hoskins and Karoly 1981). A positive index phase of the AAO weakens the middle-low latitude (~30°S–40°S) zonal flow, so the propagation speed of low-frequency Rossby waves slow down and it takes more time for the dissipation to damp disturbances before the low frequency Rossby waves reach tropical latitude (Thompson and Lorenz 2004). Figure 5 shows high- and low-frequency Eliassen-Palm flux (Edmon et al. 1980) anomalies of the SH regressed onto the daily DJF AAOI at lag +15 days. The E-P fluxes provide a direct measure of Rossby-wave propagation in the meridional plane. Figure 5 (bottom) clearly shows that there are

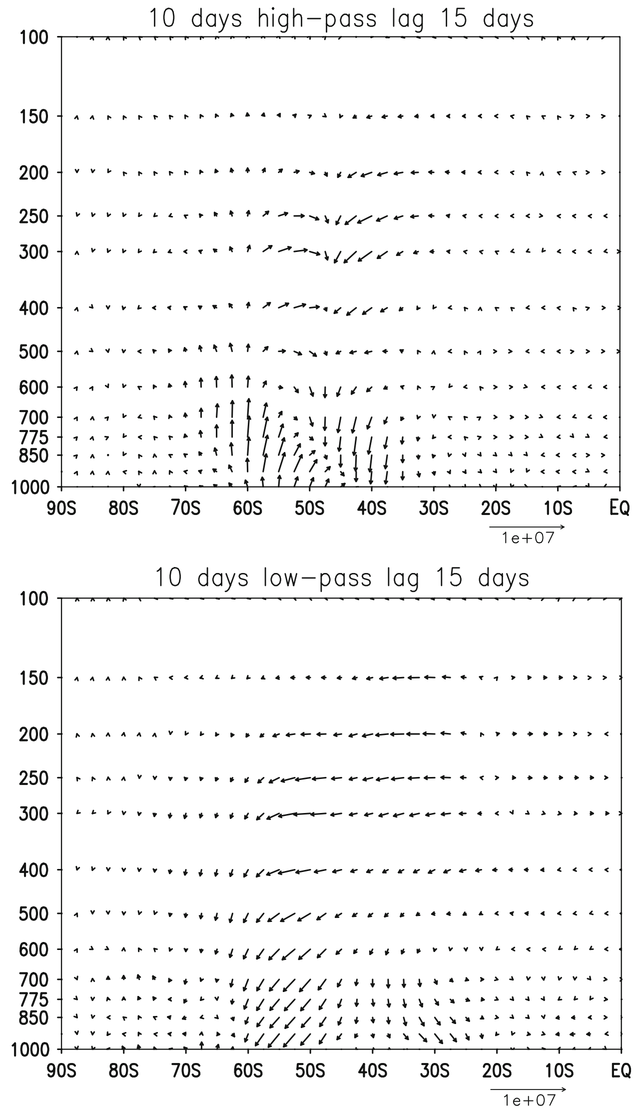

Fig. 5 10-day high- and low-pass E-P fluxes anomalies of the SH regressed on the daily DJF AAO index at lag +15 days. This result is evaluated for a positive AAO phase

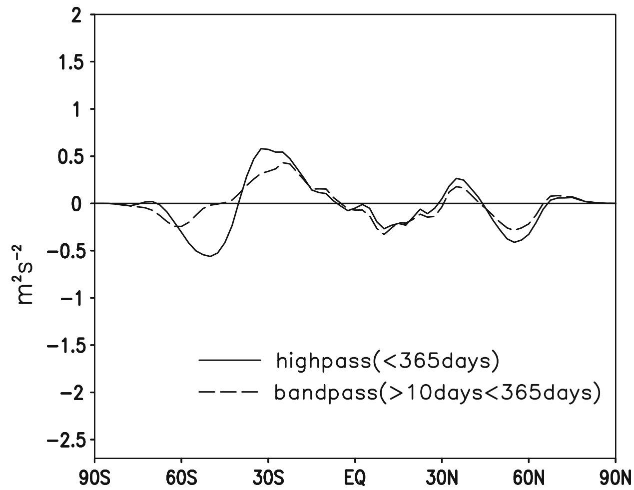

Fig. 6 Zonally averaged 100-hPa eddy momentum flux anomalies $\left(\overline{u'v'}\cos\theta\right)$ regressed onto the daily AAO index averaged from day −20 to day +20. *Solid lines* denote eddy momentum flux by all eddies and *dashed lines* denote eddy momentum flux by eddies with time scales longer than 10 days. This result is evaluated for a positive AAO phase

anomalous southward wave fluxes in the upper troposphere over middle-low latitude of the SH, but these anomalous E-P fluxes do not occur in high frequencies data (Fig. 5 top). These plots validate the hypothesis proposed by Thompson and Lorenz (2004) that, during a positive (negative) phase of the AAO, the AAO not only decreases (enhances) the zonal wind in the middle-low latitude but also meridional group velocity of low-frequency Rossby waves in the middle-low latitudes. So during the positive (negative) phase of the AAO, there is anomalous eddy momentum flux convergence (divergence) at the upper tropical troposphere (see Fig. 6). Consider the quasi-geostrophic zonal mean zonal momentum equation on a beta plane (Holton 2004):

$$\frac{\partial}{\partial t}\overline{u} = -\frac{\partial}{\partial y}\overline{u'v'} + f_0\overline{v} + \text{friction} \tag{1}$$

where u and v denote the zonal and meridional wind, f_0 the Coriolis parameter, over bar the zonal mean, and primes departures from the zonal mean. The term on the lhs corresponds to the acceleration of the zonal-mean wind; the first term on the rhs denotes meridional convergence of zonal eddy momentum flux, which reflects the forcing of the extratropical zonal flow by the eddies; the second term on the rhs, the Coriolis force acting on the meridional component of the flow; and the last term on the rhs, the friction of the planetary boundary layer. When integrated vertically (the Coriolis term equals zero) and neglecting friction, the vertically averaged extratropical zonal-mean wind speed can only be increased by the convergence of the eddy momentum flux. Therefore, through change of the meridional group velocity of the low frequency Rossby waves in the SH middle-low latitudes which can propagate into tropics, the AAO cause anomalous eddy momentum flux convergence (divergence) in the tropics, generating the tropical zonal wind anomalies.

4.2 Quasi-stationary Rossby waves and eddy momentum flux

The anomalous upper-tropospheric zonal wind shown in Fig. 3 gives rise to anomalous divergence field. The anomalous divergence field alters the generation of the horizontal component of atmospheric vorticity, and triggers large-scale atmospheric Rossby waves that can propagate into higher latitudes along a ray (Hoskins and Karoly 1981). These quasi-stationary Rossby waves are usually called teleconnection patterns (Wallace and Gutzler 1981). The propagation and sources of the quasi-stationary Rossby waves can be studied in terms of the wave activity vectors as defined by Plumb (1985), which is an extension of the Eliassen-Palm fluxes (Edmon et al. 1980) to three dimensions. The flux components, which are applicable for quasi-geostrophic stationary waves on a zonal flow, are

$$F_{\lambda} = \frac{p}{P_0 a^2 \cos\phi}\left[\left(\frac{\partial\psi'}{\partial\lambda}\right)^2 - \psi'\frac{\partial^2\psi'}{\partial\lambda^2}\right] \quad (2)$$

$$F_{\phi} = \frac{p}{P_0 a^2}\left(\frac{\partial\psi'\partial\psi'}{\partial\lambda\,\partial\phi} - \psi'\frac{\partial^2\psi'}{\partial\lambda\partial\phi}\right) \quad (3)$$

where streamfunction $\psi = \Phi/2\Omega\sin\phi$, Φ is the geopotential, Ω is the Earth's rotation rate, p is pressure, p_0 is standard constant pressure, ϕ is latitude, λ is longitude, and a is radius of the Earth, primes denote the deviation from the zonal mean (Park and Schubert 1997). Figure 7 shows the global geopotential height field anomalies at 300-hPa regressed on the daily AAOI (contours) in boreal winter and corresponding wave activity flux (vector) at lags +15, +20, +25, +30, +35 days. At first, the northward propagation of Rossby waves from the tropics, which resemble the PNA (lag +20, +25 days) locate in the west coast of North America is evident, subsequently the NA pattern (lag +30, +35 days) appears in the North Atlantic section.

It is well known that Rossby waves play a significant role in the momentum budget. The anomalous quasi-stationary waves in the NH (see Fig. 7) should be associated with eddy momentum flux anomalies. Figure 8 shows the NH eddy momentum flux ($u'v'\cos\theta$, the eddy momentum flux anomalies are multiplied by the cosine of the latitude θ) anomalies at 300-hPa regressed onto the daily AAO index (contours) averaged from day +25 to day +40, and mean wave activity flux (vector) from day +25 to day +40. It is clear that the eddy momentum flux anomalies are mainly located in the North Pacific and the North Atlantic and are accompanied by anomalous wave activity fluxes. Because anomalous Rossby waves, which are aroused by the AAO occur in the North America/Atlantic section, only the anomalous eddy momentum flux in the North Atlantic is associated with the AAO.

The anomalous quasi-stationary waves associated with the AAO make an anomalous momentum flux when considering the climatological stationary waves. Figure 9 shows a composite of the 300-hPa geopotential height averaged from lags +25 days to lags +40 days during the extremely positive/negative AAO phase in DJF. If the AAOI is greater than 2 (94 events) or less than −3 (66 events), the AAO is defined to be an extremely positive or negative event. Note that the quasi-geostrophic theory the geopotential height contour lines are generally a good approximation to the streamlines. In the extremely positive AAO phase, the North America/Atlantic trough line has somewhat northeast-southwest tilt, so there is anomalous northward eddy momentum flux than its climatological value; but contrast, in the extremely negative AAO phase, the trough line has a slight northwest-southeast tilt and there is a pronounced downstream ridge, so that there is an anomalous southward eddy momentum flux (e.g. DeWeaver and Nigam 2000).

Figure 10 show that zonal averaged eddy momentum flux anomalies $\left(\overline{u'v'}\cos\theta\right)$ averaged over 1,000 to 100-hPa regressed on the daily AAO index in northern winter (DJF) for lags ranging from −60 to +60 days (top), and mean regression results from day +25 to day +40 (bottom). Solid lines denote $\overline{u'v'}\cos\theta$ by all eddies and dashed lines denotes $\overline{u'v'}\cos\theta$ by eddies with time scales longer than 10 days (apply 10–365 days band pass FFT filter to the 'u' and 'v' before computing the eddy momentum flux anomalies).[2] As shown in Fig. 10, the AAO is associated with statistical significant anomalous eddy momentum flux in the NH from day +25 to day +40. There are eddy momentum flux divergences in the region from ∼15 to ∼40°N and convergence in middle-high latitude ∼40–65°N (see Fig. 3 bottom). According to the Eq. 1 such distribution of the anomalous eddy momentum flux might lead to the zonal mean zonal wind anomalies in the NH as shown in Figs. 1 and 2.

5 Discussion and summary

In this investigation, we investigate the influence of the AAO on the NH zonal mean zonal winds during boreal winter. In order to remove the influences of ENSO, a 365-day high pass FFT filter is applied to all daily data. It is found that a variability of the AAO has substantial influence on the NH troposphere. During a positive (negative) index phase of the AAO there are anomalous easterlies (westerlies) in ∼30°–40° and westerlies (easterlies) in ∼45°–65° latitudes of the NH after about 25–40 days later (demonstrated in Sect. 3). The physical mechanism responsible for the linkages between the AAO and the zonal mean zonal wind anomalies of the NH is discussed (Sect. 4). It shows that the AAO is associated with anomalous zonal winds in the upper troposphere over the central-eastern tropical Pacific, which forced the anomalous divergence field of the tropical upper troposphere. The

[2] In fact, to remove the influence of ENSO, the 'u', 'v' have been subject to 365 days high pass FFT filter before computing the eddy momentum flux anomalies (see Sect. 2).

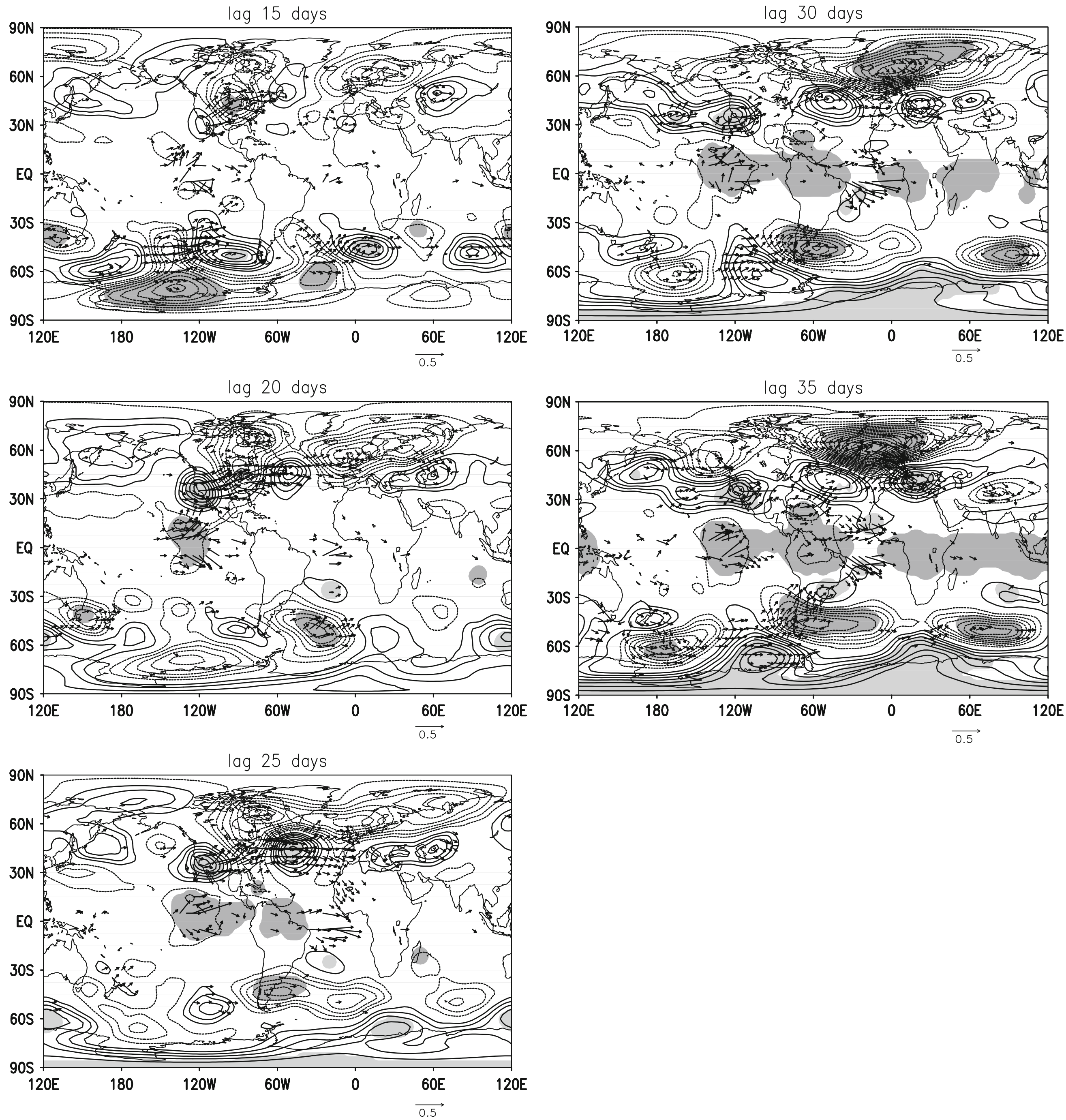

Fig. 7 Geopotential height anomalies at 300-hPa regressed on the daily AAO index (*contours*) and corresponding wave activity flux (*vector*) at lag +20, +25, +30, +35 days. *Shading* denotes correlations exceeding the 95% confidence level based on the *t*-statistic. Contour intervals are 2 m, zero contours and vector less than 0.02 m/s are omitted, and *negative contours* are dashed. This result is evaluated for a positive AAO phase

disturbances in the tropics trigger quasi-stationary Rossby waves that resemble the PNA propagate into the NH and then gradually evolve into an NA-like teleconnection pattern. The anomalous quasi-stationary Rossby waves play a key role in the generation of eddy momentum flux anomalies in the NH, and the zonal wind anomalies discussed in Sect. 3 (see Figs. 1 and 2) are accelerated (decelerated) by the anomalous eddy momentum flux convergence (divergence). But it should be pointed out that from the angle of stationary waves, the anomalous

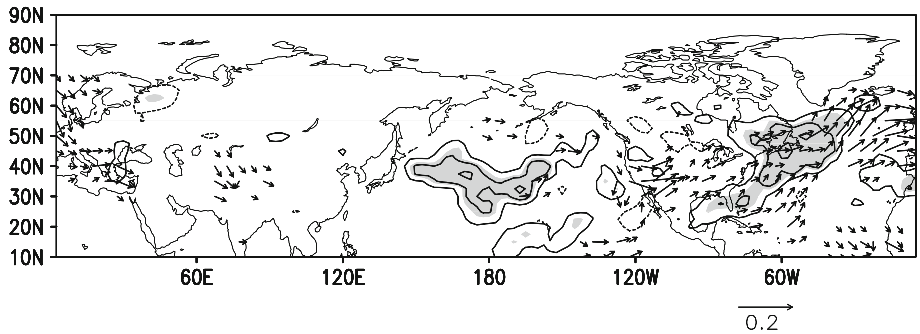

Fig. 8 NH eddy momentum flux ($u'v' \cos \theta$) by eddies with time scales longer than 10-day anomalies at 300-hPa regressed on the daily AAO index (*contours*) averaged from day +25 to day +40 and mean wave activity flux (*vector*) from day +25 to day +40. Contour intervals are 5 $m^2\ s^{-2}$, zero contours and vector less than 0.02 m/s are omitted, and negative contours are *dashed*. Values exceeding $\pm 8\ m^2\ s^{-2}$ are *shaded*. This result is evaluated for a positive AAO phase

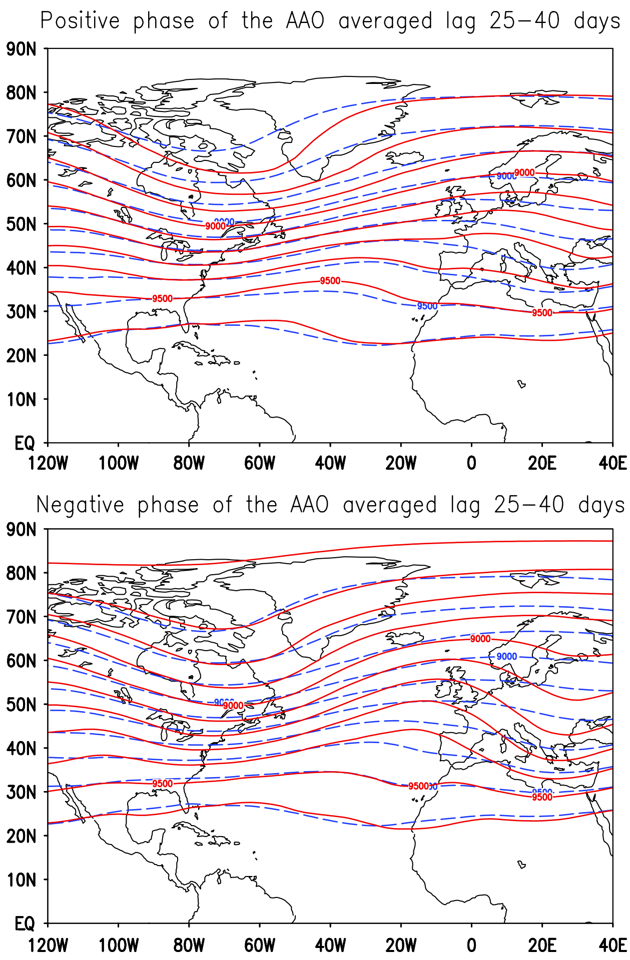

Fig. 9 Composites of 300-hPa geopotential height (*solid line*) for extreme positive (*top*) and negative (*bottom*) phases of the AAO and climatological 300-hPa geopotential height (*dashed*). *Contour intervals* are 100 m (see text for details)

stationary waves can only explain the eddy momentum flux anomalies partly, and we notice that the high- frequency eddy is the main contributor of the total eddy momentum flux anomalies in the NH (see Fig. 10 bottom). The high-

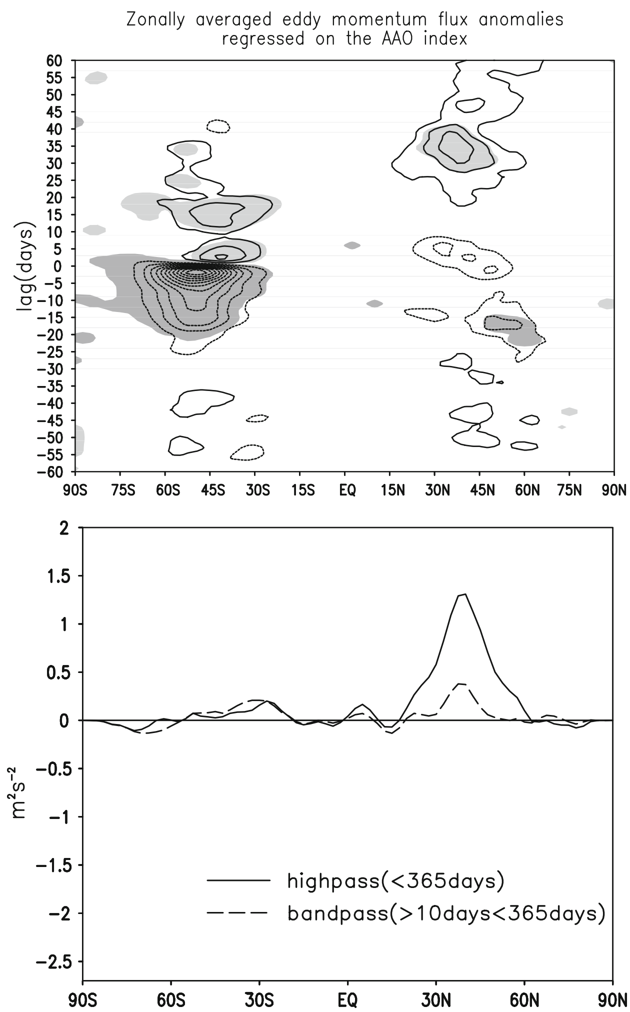

Fig. 10 Zonally averaged eddy momentum flux anomalies $\left(\overline{u'v'} \cos \theta\right)$ averaged over 1,000- to 100-hPa and regressed on the daily AAO index (*top*) and mean regression results from day +25 to day +40 (*bottom*). *Shading* denotes correlations exceeding the 95% confidence level based on the *t*-statistic. *Contour intervals* are 0.5 $m^2\ s^{-2}$, zero contours are omitted, and *negative contours* are dashed. *Solid lines* denote eddy momentum flux by all eddies and *dashed lines* denote eddy momentum flux by eddies with time scales longer than 10 days. This result is evaluated for a positive AAO phase

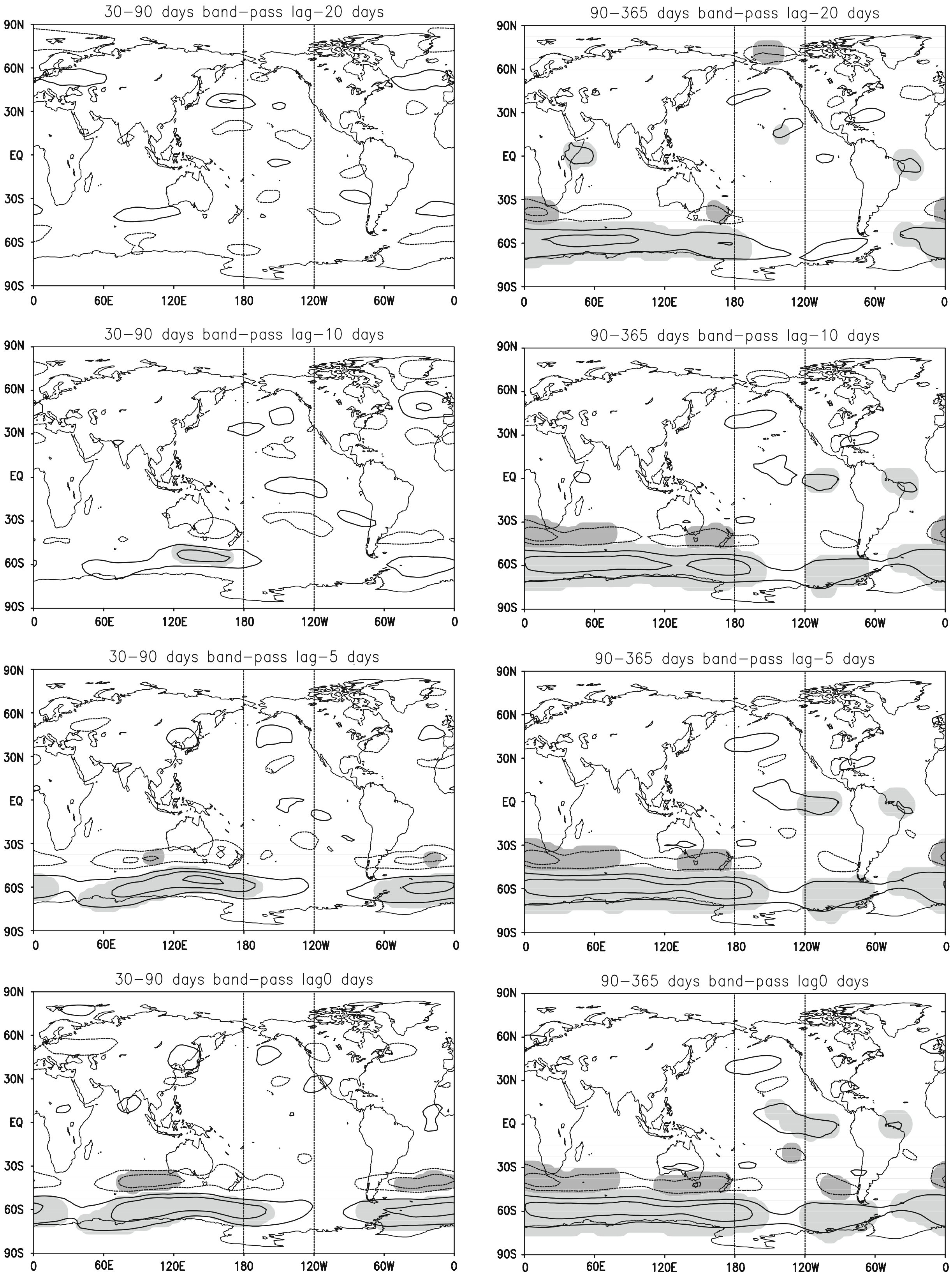

Fig. 11 30–90 and 90–365 days band-pass zonal wind anomalies at 200-hPa regressed on the daily DJF AAO index at lag −20, −10, −5, 0, +5, +10, +20 days. *Shading* denotes correlations exceeding the 95% confidence level based on the *t*-statistic. *Contour intervals* are 0.5 m/s, zero contours are omitted, and *negative contours* are dashed. This result is evaluated for a positive AAO phase

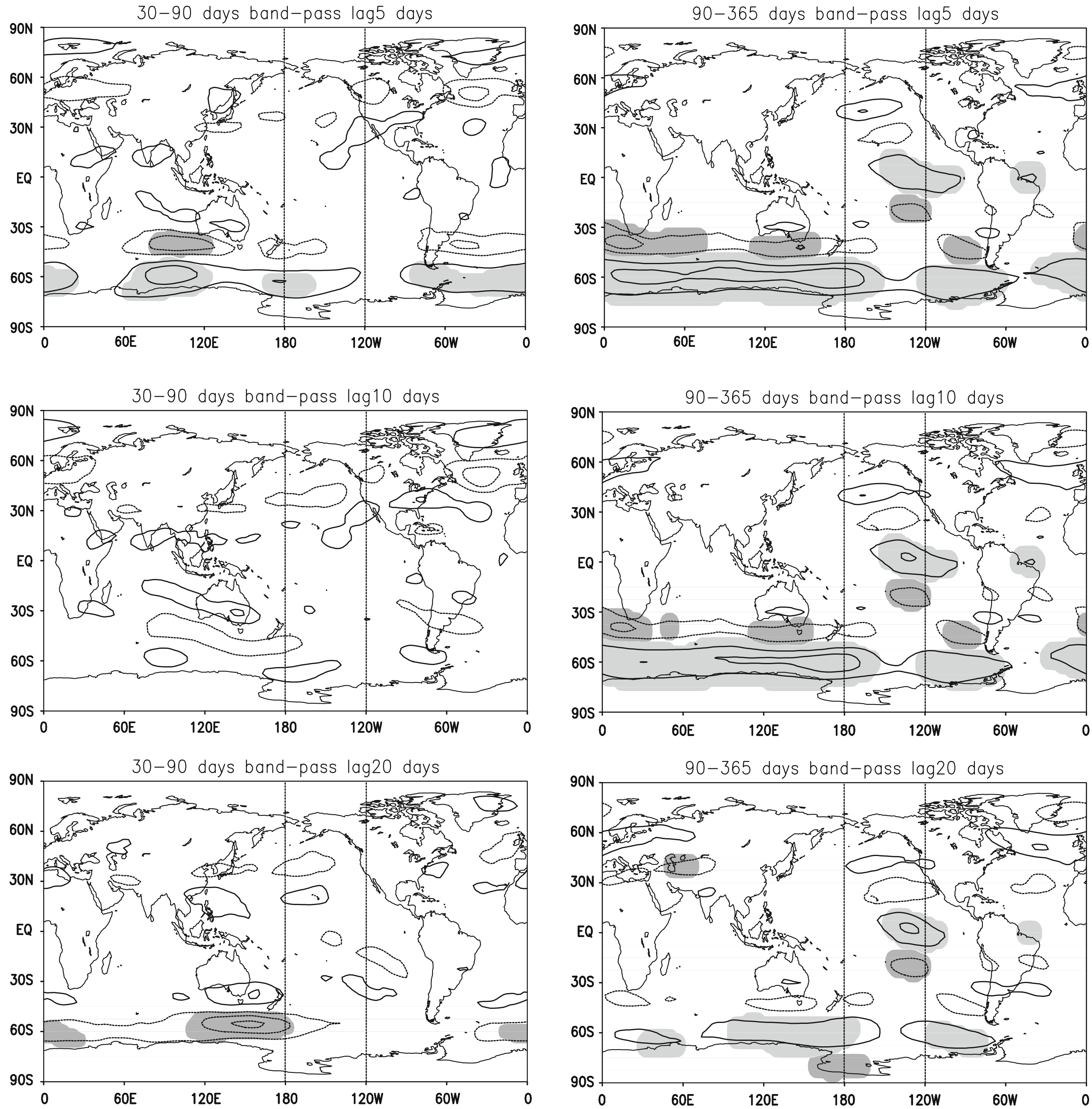

Fig. 11 continued

frequency eddy momentum flux anomalies might due to the interactions between the anomalous quasi-stationary Rossby waves and the Atlantic storm track. In fact, the NA teleconnection also named as North Atlantic Oscillation (NAO, e.g. Hurrell 1995) had been proven that it is associated with the remarkable Atlantic storm track anomalies (Rivière and Orlanski 2007).

There is also an interesting phenomenon that should be noticed, that is, the NA-like teleconnection pattern (at lag +30 to +35 days in Fig. 7) is found to lag the PNA-like quasi-stationary Rossby waves (at lag +20 to +25 days in Fig. 7). Feldstein (2003) analyzed the complete life cycle of positive and negative NAO phases and founded that after anomalous wave train propagation across the North Pacific to the east coast of North America, a positive NAO phase develops. Benedict et al. (2004), Rivière and Orlanski (2007) and Woollings et al. (2008) suggested that the physical essential of the NAO is the remnants of *upstream* waves breaking. Given the previous studies it is reasonable to picture PNA-like quasi-stationary Rossby waves located in the west coast of North America having an influence on the variability of NAO. However, this interesting phenomenon deserves further research, and we will report our results in a future study.

For consolidating our results, the data used in this study are filtered and frequencies corresponding to periods less than 365 days are retained. Therefore, the influences of tropical intraseasonal oscillations such as Madden–Julian oscillation (MJO, Madden and Julian 1971) become the most important sources of wave activity in the tropical Pacific. So the possible influences of MJO on the results of the present study cannot be neglected absolutely. There are studies showing that the zonal mean zonal wind anomalies associated with the MJO are observed to propagated poleward to both hemispheres (e.g. Weickmann et al. 1994, 1997). Carvalho et al. (2005) reported that the propagation of MJO may have an important implication for the onset of the negative AAO phase too. Thus, the daily data and the AAO time series used in this study also include the variability from intraseasonal frequencies that could be influenced by the MJO. Just like ENSO, this might create an artificial correlation between the AAO and zonal winds of the NH.

In order to investigate the potential influence of the MJO and on the basis that the influences of MJO on atmospheric circulation should mainly locate in intraseasonal frequencies band (30–90 days), global 200-hPa zonal wind anomalies in intraseasonal frequencies (30–90 days band-pass, representing the influences of intra-seasonal MJO) and in very low frequencies (90–365 days band-pass completely excluding possible influences of intra-seasonal MJO) are used. Figure 11 shows the 200 hPa zonal wind anomalies in these two frequencies band lead-lag regressed onto the DJF daily AAOI in the global domain at lag −20, −10, −5, 0, +5, +10, +20 days. These results show that the anomalous zonal winds located in the upper troposphere over the central-eastern tropical Pacific provide a "pathway" whereby the AAO extents impact, do not appear in the intraseasonal 200 hPa zonal wind anomalies regression pattern, instead wind anomalies mainly appear in the low frequencies band. Figure 11 also illustrates that the zonal wind anomalies in the tropical Pacific gradually evolve following the AAO anomalies (from lag 0 to lag +20 days in 90–365 days band-pass). Note that in the intraseasonal frequencies, there are no signatures of the zonal wind anomalies in the upper troposphere over the central-eastern tropical Pacific which triggers the inter-hemisphere teleconnection. Therefore, the results reported by this study are not contaminated by intraseasonal tropical convection (MJO).

The findings in this study contribute to an extension of the influences of the AAO, especially outside of the SH. The influences of the AAO on the NH troposphere are most evident during the northern cold season months. However, no notable influences of the AAO on the NH during boreal summer are found. For the AAO's NH equivalent the Arctic Oscillation (AO), there is no evidence that it can impact the SH circulation in any season. The AAO cannot affect atmospheric circulation of the NH on its own and must be assisted by quasi-stationary Rossby waves which resemble the PNA/NA teleconnection patterns. That kind of quasi-stationary Rossby waves are strong in winter and weak in summer, strong in the NH and weak in the SH. This may help us understand why only the AAO has influence on the NH, and why linkages between SH and NH only happen during the northern winter. It must be emphasized that while the inter-hemispheric connections between the AAO and the NH zonal wind anomalies found in this study are statistically significant, they are nevertheless not as significant as ENSO impacts.

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