The Link between the North Pacific Climate Variability and the North Atlantic Oscillation via Downstream Propagation of Synoptic Waves

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(Manuscript received 6 August 2014, in final form 20 February 2015)

ABSTRACT

The North Atlantic Oscillation (NAO) response to the northeast Pacific climate variability is examined using the ERA-40 dataset. The main objective is to validate a mechanism involving downstream wave propagation processes proposed in a recent idealized companion study: a low-frequency planetary-scale ridge (trough) anomaly located in the eastern Pacific–North American sector induces more equatorward (poleward) propagation of synoptic-scale wave packets on its downstream side, which favors the occurrence of anticyclonic (cyclonic) wave breakings in the Atlantic sector and the positive (negative) NAO phase.

The mechanism first provides an interpretation of the canonical impact of the El Niño–Southern Oscillation on the NAO in late winter. The wintertime relationship between the Pacific–North American oscillation (PNA) and the NAO is also investigated. For out-of-phase fluctuations between the PNA and NAO indices (i.e., the most recurrent situation in late winter), the eastern Pacific PNA ridge (trough) anomaly modifies the direction of downstream wave propagation, triggering more anticyclonic (cyclonic) wave breakings over the North Atlantic. For in-phase fluctuations, the effect of the eastern Pacific PNA anomalies is cancelled out by the North American PNA anomalies. The latter anomalies being deeper and more centered in the latitudinal band of downstream wave propagation, they are able to reverse the direction of wave propagation just before waves enter the Atlantic domain. The contrasting relationship between the PNA and NAO is similar to what occurs for the two leading hemispheric EOFs of geopotential height: the northern annular mode (NAM) and the cold ocean–warm land (COWL) pattern. The proposed mechanism provides a physical meaning for the NAM and COWL patterns.

1. Introduction

There are numerous evidences that the major modes of climate variability in the North Pacific exert some influence on the leading mode of atmospheric variability in the North Atlantic, the so-called North Atlantic Oscillation (NAO). The last decades were marked by many studies on the influence of ENSO on the North Atlantic atmospheric circulation (Van Loon and Madden 1981; Fraedrich 1990; Fraedrich and Müller 1992; Dong et al. 2000; Pozo-Vázquez et al. 2001, 2005; Lin et al. 2005; Brönnimann 2007; Li and Lau 2012a,b). Using reanalysis datasets, Pozo-Vázquez et al. (2001, 2005) showed that strong La Niña events were associated in winter (December-January) with a significant positive NAO-like phase sea level pressure (SLP) pattern. On the contrary, no significant SLP pattern was found in the North Atlantic during El Niño events. This stronger link between La Niña and the positive NAO-like phase was also observed by Fraedrich (1990) and Fraedrich and Müller (1992). Moron and Gouriand (2003) studied the seasonal modulation of ENSO in the North Atlantic domain: the response to ENSO forcing in the North
Atlantic is opposite in November–December to that of January–March and is stronger in late winter (February–March). This could explain why Pozo-Vázquez et al. (2001, 2005) did not find a significant response to El Niño in their study. The result of Moron and Gourand (2003) is in good agreement with that of Dong et al. (2000), who studied the impact of ENSO on the North Atlantic atmospheric circulation in an atmospheric general circulation model by prescribing sea surface temperature (SST) anomalies in the equatorial Pacific. They managed to reproduce the negative NAO-like phase anomaly observed in the North Atlantic during the strong 1997/98 El Niño winter. Cassou and Terray (2001) found similar results when forcing the ARPEGE global circulation model with prescribed SSTs over the 1948–97 period: the Azores anticyclone tends to be intensified during El Niño events and lowered during La Niña events; the reverse is true for the Icelandic low. As Brönnimann (2007) concluded from all these observational and numerical studies, there exists a “canonical” late winter signal of the ENSO impact on the North Atlantic sector, the negative (positive) NAO phase being excited by El Niño (La Niña) events at this particular period of the year. There are different competing mechanisms to explain this influence. First, the quasi-stationary Rossby wave train excited by ENSO, which takes the shape of the Pacific–North American (PNA) teleconnection, usually extends in the northwestern Atlantic and can be reinforced there by the transient eddy-mean flow interaction (e.g., Cassou and Terray 2001; Pozo-Vázquez et al. 2001). Second, the stratosphere could act as a Pacific–Atlantic bridge to transmit the signal from one basin to another (e.g., Castanheira and Graf 2003; Ineson and Scaife 2009). Finally, a more important role has been recently conferred on synoptic eddies by Li and Lau (2012a, b) through the so-called downstream development process (Chang and Orlanski 1993; Chang 2001), which corresponds to a downstream dispersion of synoptic wave energy. ENSO modifies wave packets propagation over North America, which has a more downstream influence on the North Atlantic storm track and can excite the NAO. El Niño events are characterized by a zonal low-latitude eastward-extended Pacific jet, favoring more downstream development from the North Pacific to the North Atlantic at low latitudes and negative NAO conditions. On the contrary, La Niña events are linked to a higher-latitude, less eastward-extended storm track and less downstream development over North America (Seager et al. 2010; Li and Lau 2012a, b). Beyond this difference in the latitude and intensity of wave propagation between El Niño and La Niña events, Li and Lau (2012b) also noticed a difference in the orientation of synoptic waves. As synoptic waves enter the Atlantic domain, they reach a more southwest–northeast (northwest–southeast) tilt favoring anticyclonic (cycloidal) wave breaking and the positive (negative) NAO phase. The present paper will confirm this finding and will more generally show how various North Pacific low-frequency modes affect the direction of propagation of synoptic wave trains across North America to influence the NAO.

As mentioned above, the PNA teleconnection constitutes one way of interpreting the remote influence of ENSO on the NAO. According to Song et al. (2009), there also exists a significant anticorrelation between the PNA and NAO indices at daily time scales, regardless of the ENSO cycle. Anomalous wave breakings over the North Atlantic found by the authors in the northwestern Atlantic in connection with PNA events support the above findings. This anticorrelation between the two modes was also observed in multicentury coupled general circulation model runs, but only for some subperiods of reanalysis datasets by Pinto et al. (2011). According to them, the growth condition for baroclinic waves is modified by the PNA in the northwestern Atlantic. Indeed, during the negative PNA phase, advection of cold air from North America and warm air from the Gulf of Mexico amplifies the baroclinicity near Newfoundland, leading to a more intense Atlantic storm track in that region and the formation of the positive NAO phase.

Other low-frequency patterns in the Pacific atmospheric flow have been shown to be linked to the North Atlantic circulation. For instance, the Aleutian and Icelandic lows are significantly anticorrelated from February to mid-March (Honda et al. 2001). The Aleutian low first develops and is followed by the formation of a PNA-like wave train that conveys wave activity over North America. This downstream wave activity propagation triggers another quasi-stationary Rossby wave train in the North Atlantic, leading to the development of the Icelandic low. Storm-track activities are shown to reinforce the quasi-stationary Rossby wave trains, which supports the findings of Cassou and Terray (2001) and Pozo-Vázquez et al. (2001).

At weekly time scales, there is also evidence of a connection between the North Pacific low-frequency variability and the NAO, which usually shows that the positive NAO is more likely subject to a remote influence of the Pacific atmospheric flow than the negative NAO (Feldstein 2003; Drouard et al. 2013, hereinafter DRA13). By analyzing daily reanalysis datasets from December to February, DRA13 showed that a north-eastern Pacific ridge reaches its peak amplitude a few days prior to the positive NAO phase. DRA13 proposed a new mechanism to explain the influence of such a northeastern Pacific low-frequency ridge on the NAO by performing
short-term sensitivity numerical experiments with a quasigeostrophic model. The large-scale ridge anomaly creates a zonally asymmetric mean flow in the northeastern Pacific, which deflects synoptic wave trains in such a way that the waves get a southwest–northeast tilt downstream of the ridge, which is maintained until reaching the North Atlantic domain. The more pronounced anticyclonic tilt and equatorward propagation favor anticyclonic wave breaking in the Atlantic and so the positive NAO phase. On the contrary, the presence of a large-scale low-frequency trough in the northeastern Pacific creates a more zonal mean flow leading to more zonal propagation of synoptic waves, which prevents the dominance of anticyclonic wave breaking in the Atlantic sector. This new mechanism is corroborated by the El Niño and La Niña composites of Li and Lau (2012b). The objective of the present paper is to show that this mechanism is systematically active when looking at the influence of various low-frequency modes occurring in the North Pacific–American sector on the NAO.

The previously mentioned studies show that the low-frequency atmospheric circulations in the Pacific and Atlantic domains are correlated and dynamically linked. This suggests that low-frequency variability over the North Pacific and the North Atlantic could be part of the same low-frequency pattern, the so-called northern annular mode (NAM). This mode has been the subject of several studies, and its dynamical meaning is still discussed (Thompson and Wallace 1998, 2000; Thompson et al. 2000; Wallace 2000; Deser 2000; Ambaum et al. 2001; Wallace and Thompson 2002; Zhao and Moore 2009). The idea of a NAM was first introduced by Thompson and Wallace (1998) under the name of the Arctic Oscillation (AO), which corresponds to the first empirical orthogonal function (EOF) of monthly wintertime SLP anomalies north of 20°N. The AO is a zonally hemispheric pattern corresponding to a dipole with a center of action located on the Arctic and a second one in the midlatitudes. The NAO can be viewed as the local manifestation of the AO in the North Atlantic domain and is highly correlated with the AO. The existence of the AO is controversial mainly because of the lack of correlation between the Pacific and Atlantic midlatitude centers of action (Deser 2000; Ambaum et al. 2001), but also because it is not systematically observed in other fields than the SLP (Ambaum et al. 2001). Wallace and Thompson (2002) argued that the lack of correlation between the Pacific and the Atlantic centers of action in midlatitudes is due to the coexistence of the NAM, in which the two centers of action fluctuate in phase, with the second mode of variability (EOF 2) of the Northern Hemisphere, in which the two centers of action have out-of-phase fluctuations. Our study aims at participating in the NAM/AO debate, as it provides a new dynamical link between the North Pacific and North Atlantic centers of action of the NAM/AO and so gives a dynamical interpretation to the NAM/AO in terms of downstream wave propagation. The Pacific center of action of the NAM/AO is shown to modify synoptic wave propagation downstream of it and to influence the type of wave breaking in the North Atlantic sector. As such, the proposed mechanism provides an interpretation of the influence of the Pacific center of action of the NAM/AO on the NAO.

Another major mode of variability in the Northern Hemisphere is the cold ocean–warm land (COWL) pattern introduced by Wallace et al. (1996), which is computed by averaging land surface air temperatures over the whole Northern Hemisphere. The positive (negative) COWL phase corresponds to warm-air (cold air) anomalies and large-scale upper-level ridge (trough) anomalies over land during the winter season. Over the North Atlantic sector, its structure has a south–north dipolar anomaly similar to the NAO. Many studies have shown that the COWL pattern strongly projects onto the second EOF of the midto-upper-tropospheric geopotential height in the Northern Hemisphere (Wu and Straus 2004; Quadrelli and Wallace 2004; Honda et al. 2007). In what follows, the key role played by the North American COWL anomalies in triggering the NAO-like anomalies will be illuminated by the light of the DRA13 mechanism.

To summarize, the present paper addresses the following questions: How general is the mechanism introduced by DRA13? Is this mechanism responsible for the link between PNA/ENSO and the NAO? And can it explain the existence of the NAM and COWL patterns? The data and diagnostic tools are presented in section 2. Section 3 presents different kinds of North Pacific and North American large-scale anomalies influencing the NAO through their impact on downstream wave propagation. In section 4, the dynamical link between ENSO/PNA and the NAO is investigated. A dynamical interpretation of the NAM/O is proposed in section 5. Results are summarized and discussed in section 6.

2. Data and methods

a. Reanalysis

In this study, we use daily and monthly means of ERA-40 field datasets (Uppala et al. 2005) from the European Center for Medium-Range Weather Forecast (ECMWF) on a 2.5° × 2.5° grid for the months from November to March for the period 1957–2002. Several fields (geopotential, temperature, and zonal and meridional winds) at 300 hPa are used. The flow is decomposed into high- and low-frequency parts, with a cutoff period of 10 days to separate the synoptic-scale signal from that of the low-frequency variability patterns.
b. Definition of various modes of variability

The daily NAO is defined as the first EOF of the 300-hPa low-frequency geopotential anomaly over the North Atlantic (20°–80°N, 90°W–40°E). The daily geopotential anomaly corresponds to the daily geopotential minus the seasonal cycle. The monthly NAO is similarly defined as the first EOF of the 300-hPa monthly geopotential anomaly over the North Atlantic. ENSO and PNA monthly indices were obtained from the Niño-3 and the PNA monthly indices provided by the NOAA website. The monthly NAM and COWL patterns are respectively defined as the first and second EOF of the 300-hPa monthly geopotential anomaly north of 20°N. The geopotential anomalies have been weighted by the square root of the cosine of the latitude in the principal component calculation.

c. Diagnostics on Rossby wave propagation

1) INTENSITY OF DOWNSTREAM ENERGY PROPAGATION

The high-frequency eddy kinetic energy and available potential energy per unit mass can be defined as $K' = \mathbf{v}^2/2$ and $P' = (h^2 / c_s^2) \theta^2 / 2$, respectively, where $\mathbf{v}$ denotes the horizontal wind and $\theta$ the potential temperature. The parameters $s^2 = -h \partial \theta / \partial p$ and $h = (R / \rho) \sqrt{\rho / \rho_c}$ depend on pressure only. The variable $\theta_c$ is the averaged potential temperature, $R$ is the gas constant, $\rho$ is a reference pressure, and $C_p$ is the specific heat of the air at constant pressure. Overbars and primes indicate the low-frequency and high-frequency parts, respectively. Following Rivière et al. (2014) and by neglecting diabatic terms, the evolution of $K'$ and $P'$ can be written as follows:

$$\frac{\partial K'}{\partial t} = -\mathbf{v} \cdot (\nabla K') + v'_a \Phi' + \omega \frac{\partial \Phi'}{\partial p} - v' \cdot (\mathbf{v}'_3, \mathbf{v}'_3^2)$$

$$+ v' \cdot (\mathbf{v}'_3, \mathbf{v}'_3 - \mathbf{v}_3 \cdot \mathbf{v}) - \frac{\partial}{\partial p} (\omega K') - \frac{\partial}{\partial p} (v' \Phi'),$$

(1)

and

$$\frac{\partial P'}{\partial t} = -\mathbf{v} \cdot (\mathbf{v} P') - \omega \frac{\partial \Phi'}{\partial p} - \frac{h^2}{s^2 \theta' \Phi' \cdot \nabla \Phi}$$

$$+ \frac{h^2}{s^2 \theta' \Phi' \cdot \nabla \Phi} - \frac{h^2}{s^2 \theta' \Phi' \cdot \nabla \Phi} - \frac{\partial}{\partial p} (\omega \theta'^2),$$

(2)

where $\omega$ is the vertical wind velocity, and $\Phi$ is the geopotential height. The subscripts “3” and “a” represent a three-dimensional vector (or operator) and the ageostrophic wind, respectively.

In the case where the low-frequency part is reduced to a time mean, Eq. (1) is the same as those of Orlanski and Katzfey (1991) and Chang (2001). The term on the lhs of Eq. (1) represents the local tendency of the eddy kinetic energy. The first term on the rhs corresponds to the convergence of the kinetic energy flux (advective flux plus ageostrophic geopotential flux). The ageostrophic geopotential fluxes $\mathbf{v}'_a \Phi'$ have been computed as in Orlanski and Sheldon (1995):

$$\mathbf{v}'_a \Phi' = \mathbf{v}' \Phi' - \mathbf{k} \cdot \nabla \cdot \mathbf{v}' \frac{\Phi'}{2 f(y)}.$$  (3)

The second term on the rhs of Eq. (1) represents the baroclinic conversion from eddy available potential energy to eddy kinetic energy; the third term is the Reynolds stress term, which includes the barotropic conversion; the fourth term is a kinetic energy transfer, for which the time-mean component is zero (Orlanski and Katzfey 1991), and is usually small compared to the others (Rivière et al. 2014); the fifth term is the convergence of vertical advective flux of eddy kinetic energy; and the sixth term is the convergence of the vertical ageostrophic geopotential flux.

The term on the lhs of Eq. (2) represents the local tendency of the eddy available potential energy. The first term on the rhs corresponds to the convergence of the horizontal advective fluxes of potential energy, the second term to the baroclinic conversion from eddy kinetic energy to eddy available potential energy, the third term to the baroclinic conversion from the mean available potential energy to the eddy available potential energy, the fourth term to a potential energy transfer for which the time mean is zero, and the fifth term to the convergence of the vertical advective flux of potential energy.

In what follows, our objective is to estimate the downstream propagation of total energy. This can be done by computing the total energy flux (TEF):

$$\text{TEF} = \mathbf{v} (K' + P') + \mathbf{v}'_a \Phi'.$$  (4)

The convergence of the total energy flux is the sum of the first two terms of Eqs. (1) and (2). The total energy flux allows the downstream transfer of eddy total energy from the Pacific storm track to the Atlantic storm track and has been shown to provide a good estimate of the group velocity (Chang and Orlanski 1994). It should be recalled that the total energy flux contains the advective flux of eddy total energy plus the dispersive flux represented by the ageostrophic geopotential flux. As downstream
development refers to the dispersive part only (Chang 1993), the expression “intensity of downstream wave propagation” is rather used in the rest of the paper and refers to the amplitude of the eastward component of TEF.

2) EDDY ELONGATION AND WAVE TRAIN DIRECTION OF PROPAGATION

To analyze the shape of the eddies, the $\mathbf{E}$ vectors were computed using the formula of Trenberth (1986):

$$\mathbf{E} = \frac{1}{2}(u^2 - v^2)\mathbf{i} - u'v'\mathbf{j},$$

where $u$ and $v$ are the zonal and meridional winds, respectively. Note that the $x$ component of the present $\mathbf{E}$ vector is half that of the $\mathbf{E}$ vector of Hoskins et al. (1983). The $\mathbf{E}$ vectors are classically used to give information on wave propagation, as they point approximately in the direction of wave energy propagation relative to the time-mean flow. Moreover, their divergence and curl show the eddy-induced acceleration of the zonal and meridional winds, respectively. The $\mathbf{E}$ vectors are also important in determining the barotropic conversion from mean kinetic energy to eddy kinetic energy (Rivièrè et al. 2003). In this study, we will mainly use them to infer the tilt of the eddies and the orientation of their propagation. If we denote the angle of the $\mathbf{E}$ vector with respect to $x$ axis as $\epsilon$, one can write $\mathbf{E} = K'(\cos\epsilon + \sin\epsilon)$, and the angle of the eddy major axis with respect to $x$ axis, denoted as $\phi$, is equal to $\phi = \pi/2 + \epsilon/2$ (Rivièrè et al. 2003). The $\mathbf{E}$ vectors pointing equatorward indicate southwest–northeast elongated eddies and equatorward energy propagation (Fig. 1b). On the contrary, $\mathbf{E}$ vectors pointing poleward indicate northwest–southeast elongated eddies and poleward energy propagation (Fig. 1d). To finish, eastward- and westward-oriented $\mathbf{E}$ vectors indicate meridional and zonal directions of elongation, respectively (Figs. 1a,c).

3) WAVE BREAKING

To complete the analysis of synoptic Rossby wave trains, the Rossby wave–breaking detection method of Rivièrè (2009; see his appendix C for a more detailed

![Fig. 1. Qualitative diagram showing the relation between the tilt of the eddies and the direction of the $\mathbf{E}$ vectors for angles of the eddy major axis equal to (a) $\phi = \pi/2$, (b) $\phi = \pi/4$, (c) $\phi = 0$, and (d) $\phi = -\pi/4$.](image-url)
This method detects wave-breaking events, identifies their type (anticyclonic or cyclonic), and computes their frequency of occurrence. Similarly to the algorithm of Strong and Magnusdottir (2008b), this method consists in detecting local overturnings of circumglobal potential vorticity contours on isentropic surfaces or absolute vorticity contours on isobaric surfaces. Each circumglobal contour is oriented from west to east. A wave-breaking event occurs when, locally, a segment of the circumglobal contours is oriented from east to west instead of west to east. Segments oriented from northeast to southwest and southeast to northwest correspond to anticyclonic (AWB) and cyclonic (CWB) wave-breaking events, respectively. Michel and Rivière (2011) checked that the absolute vorticity field on isobaric levels gives qualitatively similar results to the potential vorticity on isentropic surfaces. In the present study, the algorithm is applied to all circumglobal contours of absolute vorticity ranging from $2.4 \times 10^{-4}$ to $4.0 \times 10^{-4}$ s$^{-1}$.

3. The general character of the NAO

Figure 2 describes the low-frequency and synoptic wave propagation anomalies associated with the daily NAO from November to March. The regression of the
low-frequency geopotential onto the daily NAO index shows the classical dipolar geopotential anomaly of the NAO, together with a slight ridge anomaly in the northeastern Pacific. From the downstream side of the ridge anomaly to the eastern Atlantic, the regressed E vectors are all significantly oriented equatorward. This means that synoptic waves propagate more equatorward during NAO+ than during NAO− events over a very broad region spanning from the northeastern Pacific to the northeastern Atlantic. The composite of NAO+ events shows a strong ridge anomaly in the Northeastern Pacific, with waves propagating poleward and equatorward upstream and downstream of the North Pacific ridge anomaly, respectively (Fig. 2c). This reflects the deflection undergone by a synoptic wave train when it travels across a large-scale ridge and compares well with the linear simulations of DRA13 (see their Fig. 6, right column). The more equatorward propagation on the downstream side of the North Pacific ridge anomaly is maintained until it reaches the eastern Atlantic. On the contrary, the composite of NAO− events shows no significant anomalies in the Northeastern Pacific, and there is no unusual E vector orientation upstream of the east coast of North America either (Fig. 2e). In the absence of any large-scale ridge anomaly in the northeastern Pacific, the mean flow is more zonal, E vectors tend to be mainly eastward oriented over North America. The E vectors are mainly poleward oriented in the Atlantic domain during NAO−, but this feature is a local effect, which cannot be attributed to any remote influence of upstream large-scale anomalies. It is a confirmation that the negative NAO phase is more triggered by local processes, whereas the positive NAO phase is more dependent on the upstream flow (e.g., Benedict et al. 2004).

To conclude, the equatorward and poleward orientation of E vectors in the North Atlantic domain confirms the well-established predominance of anticyclonic and cyclonic wave-breaking events for the positive and negative NAO phases, respectively (e.g., Benedict et al. 2004; Martius et al. 2007; Rivièr and Orlanski 2007; Strong and Magnusdottir 2008a; Woollings et al. 2008). It confirms also the occurrence of anticyclonic wave-breaking events in the subtropical eastern Pacific regions during NAO+ events (Benedict et al. 2004; Strong and Magnusdottir 2008b). The new point underlined here is that the dominance of equatorward propagation during NAO+ events precisely starts from the eastern edge of the North Pacific ridge anomaly and is maintained until synoptic waves reach the North Atlantic. It suggests that the large-scale ridge acts to reorient synoptic wave propagation, as in DRA13.

Let us now analyze the difference in the amount of downstream eddy energy propagation between the two NAO phases. No significant signal in the regression of the energy fluxes’ magnitudes on the NAO is observed over North America (Fig. 2b). Again, no major difference in the energy fluxes’ magnitudes is seen over North America between the positive and negative NAO phases (Figs. 2d,f). The only significant feature outside the North Atlantic domain is during NAO+ events, for which there are more intense eastward energy fluxes north of the large-scale North Pacific ridge than south of it (Fig. 2d). But such a difference does not extend in the North American sector. Therefore, the main difference between the two NAO phases over North America lies in the shape and direction of wave propagation rather than in the amount of downstream wave propagation.

4. Influence of ENSO and the PNA on the NAO

The present section aims at showing how the large-scale anomalies of ENSO and PNA in the northeast Pacific and over North America modify the downstream propagation of synoptic waves.

a. ENSO

Figure 3 shows how ENSO modifies the atmospheric circulation from the northeast Pacific to Europe using monthly mean reanalysis datasets. Composites and regressions were computed using the December–February (DJF)-mean Niño-3 index but looking at the late winter (January–March): that is, the period where the influence of ENSO on the North Atlantic atmospheric circulation is the strongest. As shown by Li and Lau (2012a), it is only during late winter that the typical ENSO northeast Pacific atmospheric anomalies appear. This supports the idea that the ENSO impact on the Atlantic is made through atmospheric dynamics linking the Pacific and Atlantic sectors. The DJF-mean Niño-3 index is used to compare our results with the studies of Li and Lau (2012a,b).

Regressed E vectors are poleward oriented from the downstream side of the northeast Pacific low-frequency trough anomaly to the eastern North Atlantic (Fig. 3a), which is opposed to the regressed E vectors on the NAO index (Fig. 2a). Composites of El Niño months (Fig. 3c) are associated with a low-frequency trough anomaly in the northeast Pacific, and E vectors are zonally oriented over North America. This is consistent with Li and Lau’s (2012b) results and the fact that the presence of the low-frequency trough create a more zonally oriented Pacific jet that tends to favor zonal propagation of synoptic waves (Seager et al. 2010). Note that the ridge anomaly located over North America is too far north of the region of downstream wave propagation to be able to modify wave propagation. On the contrary, during La Niña months, a low-frequency ridge anomaly is visible in
the northeast Pacific at the latitude of the Pacific storm track, and waves mainly propagate equatorward downstream of it (Fig. 3e).

The regressed energy flux magnitude exhibits a significant dipolar anomaly over North America (Fig. 3b), with a negative center from 40° to 60°N and a positive center from 20° to 40°N. ENSO corresponds to a latitudinal shift of downstream wave propagation. There is more downstream wave propagation between 40° and 50°N over the northeast Pacific/North America (140°–90°W) during La Niña events when the ridge anomaly deflects the Pacific jet and wave packets northward, whereas, downstream wave propagation occurs at lower latitude (between 25° and 40°N) over the same longitudes during El Niño events. Therefore, major differences between the two phases appear in both the latitude and direction of propagation of the waves, consistent with Seager et al. (2010) and Li and Lau (2012b). However, no difference in the amplitude of downstream wave propagation is seen. The latter point is to be contrasted with the findings of Li and Lau (2012b), who underlined a more important downstream propagation of energy during El Niños than during La Niñas.

Looking at the North Atlantic area, a dipolar geopotential anomaly is visible for the two cases (Figs. 3c,e).
A negative NAO-like phase dipole with a positive anomaly to the north of a negative anomaly is present during El Niño events when E vectors over North America are zonal (Fig. 3c). On the contrary, a positive NAO-like phase dipole with a negative anomaly to the north of a positive anomaly is present during La Niña events when E vectors are oriented equatorward (Fig. 3e). Moreover, as expected, there are more cyclonic and less anticyclonic wave breakings during El Niños than during La Niñas (cf. Figs. 3d and 3f) in the Atlantic sector, in agreement with the more northward position of the Atlantic jet for the latter than the former.

Thus, concluding on ENSO, we again observe a modification of the direction of wave propagation downstream of the northeast Pacific low-frequency anomalies. The low-frequency trough anomaly of El Niño events creates a zonal Pacific jet eastward extending across North America that induces zonal propagation and prevents anticyclonic wave-breaking events in the North Atlantic. The low-frequency ridge anomaly of La Niña events induces a deflected Pacific jet that reorients wave propagation equatorward, downstream of it, to favor anticyclonic wave-breaking events in the North Atlantic.

b. PNA

As for ENSO, the composites and regressions based on the PNA index shown in Fig. 4 are made from January to March. The reason for this choice is that the correlation between the PNA and NAO indices for months between January and March is near 0.21, which is statistically significant at a 98% confidence level, while the correlation for months between November and December is near zero. The anticorrelation between the two indices in late winter is consistent with Honda et al. (2001). Similarly to ENSO, the anomalous E vectors are mostly significantly oriented northwestward at midlatitudes over the northeast Pacific/North America (Fig. 4a). This anomalous orientation starts from the downstream side of the trough, which is located farther westward compared to the one in the regression on the Niño-3 index (cf.
fig. 4a with fig. 3a). During the positive PNA phase (Fig. 4c), downstream of the northeast Pacific trough anomaly, the E vectors are mainly eastward oriented. On the contrary, for the negative PNA phase, downstream of the northeast Pacific ridge anomaly, the E vectors point more equatorward than usual. Low-frequency anomalies of opposite sign to those located in the northeast Pacific are present over North America for the two phases. However, these North American low-frequency anomalies do not impact wave propagation because they are located at a much higher latitude than the mean position of the storm track (see the energy fluxes location in Figs. 4c,e).

The regressed energy flux magnitude has a dipolar structure (Fig. 4b), with a much stronger northward negative anomaly (between 35° and 60°N) compared to the southward positive anomaly (between 20° and 35°N). This is to be contrasted with the Niño-3 regression, for which the equivalent dipolar structure (Fig. 3b) was much more symmetric. This asymmetric behavior in downstream wave propagation intensity between the PNA phases is confirmed by the composites of Figs. 4c and 4e. For the positive PNA phase, downstream wave propagation intensity is significantly weaker than for the negative PNA phase. It is intriguing to observe that less downstream propagation occurs during the positive PNA phase when the Pacific jet is more zonal and extends much more eastward.

To conclude, over North America, the two PNA phases differ in the latitude and intensity of downstream wave propagation, as well as the orientation of wave propagation. Similar to El Niño (La Niña), because of the presence of the northeast Pacific low-frequency anomalies, waves propagate more zonally (equatorward) in the southern (northern) regions of North America. However, in contrast with ENSO, for which El Niño and La Niña show roughly the same intensity of downstream wave propagation, there is significantly more downstream wave propagation during the negative PNA phase than during the positive PNA phase. Part of this difference between ENSO and PNA may rely on the difference in the Pacific jet structure. El Niño events, as diagnosed from Niño-3 index, have a more eastward-extended Pacific jet than the positive PNA phase.

There are also significant differences between the two PNA phases over the North Atlantic. During the positive PNA phase, a positive geopotential anomaly is located to the north of a negative one (Fig. 4c), whereas during the negative PNA phase, the reverse happens (Fig. 4e). These geopotential anomalies are related to a more northward and southward Atlantic jet for the negative and positive PNA phases, respectively. It is accompanied by more anticyclonic and fewer cyclonic wave-breaking events over the whole Atlantic during the former than the latter phase, in good agreement with the anomalous E vectors (Figs. 4c-f). All these distinct features fit well with a more positive (negative) NAO being triggered by a more negative (positive) PNA phase in late winter.

To better understand the dynamical link between the PNA and NAO and its seasonal variations, composite maps of opposite-sign and same-sign PNA and NAO anomalies are shown in Fig. 5 by taking into account all the winter months from November to March. A threshold of ±0.7 standard deviation is here used to get more months in the composites and more statistical significance. There are many more months with opposite-sign PNA and NAO anomalies than same-sign anomalies in late winter (see Table 1), as expected from the significant anticorrelation between the two indices at that period of the year. This is not the case in early winter.

For opposite-sign PNA and NAO composites (hereinafter denoted as expected cases), the E vectors keep the same anomalous orientation from the downstream side of the northeast Pacific anomalies to the Atlantic. They anomalously point equatorward for PNA+ (Fig. 5a) and poleward for PNA+, NAO− (Fig. 5c), leading to more AWB events in the former and more CWB events in the latter.

For same-sign PNA and NAO composites (hereinafter denoted as unexpected cases), the E vectors are not as anomalously oriented on the immediate downstream side of the northeast Pacific anomalies as for the previous composites (cf. Fig. 5a with Fig. 5d and Fig. 5b with Fig. 5c). Part of the explanation relies on the fact that the northeast Pacific anomalies have less amplitude in same-sign than in opposite-sign PNA and NAO composites. Another reason might come from our statistical test, which is based on the y component of the E vectors and not on the angle of the E vectors. For instance, in the PNA+, NAO+ case (Fig. 5b), even though the E vectors are poleward oriented over the eastern North Pacific, the anomalies are less statistically significant because the E vectors have less amplitude in that area. Another major difference between the expected and nonexpected cases is the change in E vectors orientations over North America. Over eastern Canada, their orientation is mainly the opposite to that reached more upstream over western Canada. This can be interpreted as resulting from the localized geopotential anomalies centered over North America. For PNA+, NAO+, the presence of the North American ridge changes the orientation of wave propagation and allows the waves to reach a more equatorward propagation on its downstream side (Fig. 5b).
PNA\(^-\), NAO\(^-\), the same observations can be made with all the signs reversed (Fig. 5d). It is the presence of the North American trough that favors the more poleward-oriented \(E\) vectors on its downstream side. One may argue that opposite-sign PNA and NAO composites also present the same type of geopotential anomalies over North America than the same-sign PNA and NAO composites. However, in the latter composites, the anomalies are localized, which is not the case in the former composites. Such a difference appears in zonal wind anomalies. The localized North American trough in PNA\(^-\), NAO\(^-\) creates a stronger cyclonic shear over Canada than the nonlocalized trough in PNA\(^-\), NAO\(^+\). The same observation can be made by comparing the PNA\(^+\), NAO\(^-\) and PNA\(^-\), NAO\(^+\) cases.

To further illustrate the above idea of the contrasting relationship between the PNA and NAO, a scatterplot of the PNA index versus the NAO index is shown on Fig. 6. As the difference between the expected and unexpected cases seems to arise from the shape and location of the North American anomalies, an average of the low-frequency geopotential anomaly has been made over this sector (35°–55°N, 130°–90°W) which is represented by the red box in Fig. 5d. Months during which a positive (negative) low-frequency North American anomaly is detected are represented by red (blue) crosses. The North American anomaly being usually negative (positive) for negative (positive) PNA, most red and blue crosses are located on the positive and negative PNA side, respectively. However, the mean position of the red (blue) crosses is on the upper-right (lower left) quadrant of the scatterplot. This means that the nonexpected cases are characterized by deeper North American anomalies than the expected cases in the latitudinal band of downstream wave propagation (typically between 35° and 55°N). Such deep anomalies in that region are able to reverse the effect of the northeast Pacific anomalies.

To better underline the role played by North American anomalies in modifying the downstream propagation of synoptic waves, composites of weak PNA and strong North American anomalies have been computed. More precisely, months for which the normalized averaged geopotential anomaly over North

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**TABLE 1.** Winter month distributions of opposite- and same-sign PNA and NAO anomalies. The identified months are used to obtain composites of Fig. 5.
America (NA) exceeds positively (negatively) the 0.7 standard deviation threshold and for which the absolute values of the normalized monthly PNA index is less than 0.7 are considered. The results are shown in Fig. 7. In the case of the North American ridge (Fig. 7a), there is a clear change in the $E$ vectors orientation from the upstream to downstream sides of the ridge, leading to more equatorward wave propagation in the North Atlantic, more AWB events (not shown) and a clear positive NAO phase dipole. On the contrary, in the case of the North American trough (Fig. 7b), $E$ vectors are unusually zonally oriented over the Great Lakes because of the downstream effect of the trough. This leads to more CWB events than usual in the Atlantic sector (not shown) and a clear negative NAO phase dipole appears in the composite. There are therefore several evidences of the role played by the North American anomalies in modifying the downstream propagation of synoptic waves and their ability to trigger one phase or another of the NAO depending on their sign.

One conclusion of the present section is that, depending on the structure of the PNA anomalies, the result in the North Atlantic sector may completely differ. In the expected cases, typically occurring in late winter, the northeast Pacific anomalies of the PNA are strong and largely determine the wave propagation orientation over North America and in the North Atlantic. In such cases, the North American anomalies do not alter wave propagation much. On the other hand, in early winter, there are proportionally more PNA cases (the unexpected cases), for which the northeast Pacific anomalies are less intense and the North American anomalies play a role in modifying the wave propagation orientation. One possible explanation of the seasonal modulation of the PNA impact on the NAO could be thus related to fluctuations in the shape and latitude of the PNA anomalies with the season. These differences might be themselves due to variations of the climatological background flow with the season, as the propagation of low-frequency Rossby wave trains is well known to depend on the background flow in which they propagate (Ambrizzi and Hoskins 1997).
5. Dynamical interpretation of the hemispheric modes

a. NAM

Let us now interpret the structure of the first two modes of the Northern Hemisphere following the same diagnostics. The NAM regression of the geopotential anomaly (contours on Fig. 8a) shows a large-scale ridge anomaly in the northeast Pacific and the classical dipole of the NAO. The regressed $E$ vectors (Fig. 8a) are significantly equatorward oriented from the downstream edge of the northeastern Pacific anomaly to the eastern North Atlantic. During the positive NAM phase, eddies are significantly tilted along the southwest–northeast direction from the downstream side of the northeast Pacific ridge anomaly to the North Atlantic (Fig. 8c). On the contrary, during the negative NAM phase, the trough observed in the northeast Pacific creates a more zonal jet, which tends to meridionally elongate the eddies. Dipolar anomalies are visible in the regression of the energy flux intensity (Fig. 8b) from the northeast Pacific to the northwestern Atlantic. These anomalies are statistically significant at 90% over the eastern North Pacific and North Atlantic and correspond to the latitudinal shifts of the Pacific and Atlantic storm tracks and jet exit regions associated with the NAM, as energy fluxes involve both high-frequency quantities and mean winds. Over North America, even though the statistical significance is not as strong as in the storm-track regions, there is a tendency for more downstream wave propagation between 40° and 60°N and less downstream wave propagation between 20° and 40°N, the former being more pronounced than the latter. Even though it is less marked than for the PNA, there is more downstream wave propagation at a higher latitude for the positive NAM phase over western North America than for the negative NAM phase.

A visualization of synoptic wave trains during the positive and negative NAM phases is provided in Fig. 9. One-point lagged regression maps of 300-hPa meridional wind anomaly are plotted using the base point 45°N, 180°W, which is located on the upstream side of the northeast Pacific anomalies of the NAM. At lag 0, parts of the wave packets lying on the upstream side of the northeast Pacific anomalies are rather similar, some slight differences appearing on the downstream side. As lag increases, differences in the shape of the synoptic eddies increase, with those for the positive NAM phase having a pronounced anticyclonic tilt, whereas those for the negative NAM phase have a more meridional tilt. This is consistent with the more equatorward and poleward-oriented $E$ vectors for the positive and negative NAM phases, respectively, as shown in Figs. 8c and 8e. At lag +4 days, as waves enter the Atlantic domain, the same differences in the tilt of the eddies are still visible.

To conclude, NAM phases slightly differ in terms of the intensity and latitude of downstream wave propagation over North America, but the more statistically significant difference relies on the orientation of wave propagation. More southwest–northeast (meridional) elongated eddies or, equivalently, more equatorward (zonal) wave propagation, from the eastern part of North America to the North Atlantic during the positive (negative) NAM phase, triggers more anticyclonic (cyclonic) wave breakings over the North Atlantic (Figs. 8d,f).

Thus, the mechanism highlighted by DRA13 is in play in the NAM dynamics too and can explain part of the link between the Pacific center of action of the NAM (i.e., the high/low Pacific anomalies) and its Atlantic center of action (i.e., the NAO dipolar anomalies). It is the influence exerted by the northeast Pacific anomalies over the North Atlantic atmospheric circulation through
the modification of synoptic wave propagation that provides a physical meaning for the NAM.

b. COWL

The regression of the geopotential anomaly on the COWL index (Fig. 10a) shows a trough anomaly in the northeast Pacific, a ridge anomaly centered over North America, and the classical positive NAO dipolar structure. The regressed E vectors (Fig. 8a) are significantly poleward oriented on the downstream edge of the northeast Pacific anomaly, but this anomalous orientation is reversed while passing through the North American ridge anomaly. The E vectors reach an anomalously equatorward orientation from the downstream edge of the North American ridge anomaly to the eastern Atlantic. For the positive COWL phase, the same general comments on the E vectors’ orientation can be made around the North American ridge anomaly. However, there is no predominant equatorward orientation all along the North Atlantic; it is only between the longitudes 0° and 40°E that the E vectors appear again to be anomalously equatorward and that there are more AWB events than usual (not shown). For the negative COWL phase, from the downstream edge of the North American trough anomaly to the eastern North Atlantic, E vectors are predominantly poleward oriented (Fig. 10e) and CWB events are more frequent than usual, leading to a negative NAO phase (Fig. 10f). Concerning the location of downstream wave propagation, the regression on the energy flux shows that there is more downstream propagation at a higher latitude than usual over Canada due to the presence of the North American ridge anomaly (Fig. 10b).

The change in wave propagation orientation due to the presence of North American anomalies is well visible in the one-point lagged regression maps of Fig. 11. At lag 0, on the immediate downstream edge of the northeast Pacific anomalies, the eddies are more anticyclonically tilted in the negative COWL phase than in the positive one. However, as wave trains propagate eastward, the

FIG. 8. As in Fig. 3, but for the monthly NAM index from November to March.
tilts are reversed by the North American anomalies. For negative COWL, the waves get a boomerang shape, and the cyclonic tilt on the northward flank of the eddies probably results from the cyclonic shear induced by the North American trough (see in particular the positive meridional wind anomaly). On the contrary, for positive COWL, as waves propagate eastward, the tilt of the eddies becomes more and more anticyclonic. These differences in the tilt of the eddies corroborate the differences between the E vectors’ orientation for the positive and negative NAM phases shown in Figs. 10c and 10e. It is also relevant to compare the negative COWL with the positive NAM and the positive COWL with the negative NAM, because they exhibit similar northeast Pacific anomalies but show drastic differences in terms of wave propagation over North America. The localized trough anomaly over North America of COWL—(Fig. 11, left column) does not have the same effect as the non-localized trough anomaly over North America of NAM+ (Fig. 9, right column) on the tilt of the eddies, because the former increases the background cyclonic shear, which is not the case for the latter.

To finish, there is a striking resemblance between the opposite-sign PNA and NAO composites and the NAM

FIG. 9. Time-lag regressions of the 300-hPa unfiltered meridional wind anomaly (contours; interval: 1 m s$^{-1}$) for (left) the positive NAM phase and (right) the negative NAM phase. The reference time series is the 300-hPa unfiltered meridional wind anomaly on day 0 at 45°N, 180°. Anomalies are constructed by subtracting the monthly mean.
on the one hand and between the same-sign PNA and NAO composites and the COWL on the other hand. PNA−, NAO+ and PNA+, NAO− resemble NAM+ and NAM−, respectively (cf. Figs. 5a,c with Figs. 8c,e); and PNA+, NAO+ and PNA−, NAO− resemble COWL+ and COWL−, respectively (cf. Figs. 5b,d with Figs. 10c,e). This suggests that the two hemisperic modes represent two different ways of PNA influencing NAO, because PNA anomalies in the northeast Pacific and over North America potentially differ from case to case.

6. Summary and discussion

The dynamical link between the North Pacific variability and the NAO in terms of downstream propagation of synoptic waves was investigated using daily and monthly reanalysis datasets. It is not so much the intensity or latitude of downstream wave propagation that matters the most to interpret this link but the direction of wave propagation over North America. Indeed, the most robust distinctive feature of the NAO in terms of downstream wave propagation, as revealed by regression and composite maps, is the anomalous equatorward wave propagation for the positive NAO phase and the more zonal wave propagation for the negative phase. The equatorward propagation during the positive NAO phase starts from the downstream edge of a low-frequency eastern Pacific ridge anomaly to the Atlantic sector, which is a first confirmation of the results of DRA13 on the action of a planetary-scale ridge anomaly on synoptic wave propagation downstream of it.

Then the results of DRA13 on the influence of low-frequency anomalies over the northeast Pacific on the NAO were validated by analyzing the effects of ENSO and PNA during late winter. Both ENSO and the PNA show a low-frequency ridge/trough anomaly over the northeast Pacific, depending on their phase. The low-frequency ridge (trough) anomaly over the northeast Pacific during La Niña (El Niño) events or during the negative (positive) PNA phase deflects the Pacific jet.
poleward (maintains the Pacific jet zonally oriented). Such mean flow configurations give a southwest–northeast (meridional) tilt to the eddies when they enter the North Atlantic, favoring anticyclonic (anticyclonic and cyclonic) wave breakings and the positive (negative) NAO-like phase. This mechanism is in good agreement with the results of Li and Lau (2012a,b), who also highlighted the importance of downstream wave propagation and mentioned the difference in the tilt of the eddies when they propagate over North America between El Niño and La Niña. Our study aims at showing that these are the low-frequency anomalies over the northeast Pacific that are responsible for the orientation of the eddies.

Some differences are noticeable between the composites of the PNA and those of Niño-3. For Niño-3, the eastern Pacific low-frequency anomalies are more eastward located than for the PNA, which leads to a more eastward-extended zonal jet over North America during El Niños than during the positive PNA phase. This difference may explain why the amplitude of downstream wave propagation is more important for the former than the latter. The consequence is that the amount of downstream wave propagation is rather symmetric between La Niña and El Niño years, which is not the case between the two PNA phases for which the negative PNA phase exhibits much more downstream wave propagation than the positive PNA phase. The common

FIG. 11. As in Fig. 9, but for the COWL.
points in terms of downstream wave propagation over North America between the PNA and Niño-3 remain the latitude of downstream wave propagation (lower for El Niños and the positive PNA phase) and the orientation of wave propagation (more equatorward for La Niñas and the negative PNA phase).

As ENSO-induced atmospheric anomalies in the eastern Pacific only appears in mid-to-late winter (Li and Lau 2012a), it makes sense that the canonical relationship between ENSO and the NAO only exists in that particular period of the year. The fact that PNA and NAO are anticorrelated in late winter but not much in early winter is less obvious. Our composites of same-sign and opposite-sign PNA and NAO indices provide some possible explanation. Opposite-sign PNA and NAO cases, which occur more frequently in late winter, have more intense northeast Pacific anomalies than same-sign PNA and NAO cases and thus have a stronger impact on the downstream propagation of synoptic waves. Another major difference occurs over North America as the same-sign PNA and NAO cases present large-scale anomalies centered in the latitudinal band of downstream propagation of synoptic waves, which is not the case for opposite-sign PNA and NAO cases. As these North American geopotential anomalies are localized in the former cases, they alter the meridional shear of the zonal wind, which significantly modifies the orientation of wave propagation and acts to reverse the orientation initially set up by the more upstream northeast Pacific anomalies. The common point over North America between PNA− and PNA+ triggering NAO+ is the presence of intense high-latitude westerlies over Canada, which do not appear in the case of PNA− and PNA+ triggering NAO− (Fig. 5). These results suggest that there are subtle variations in PNA anomalies that significantly affect the way synoptic waves propagate over North America. Therefore, the exact structure/position of the PNA anomaly over both the northeast Pacific and North America might be an important factor to explain its downstream impact on the NAO and potentially the seasonal modulation of this impact.

The present study also provides a physical meaning for the NAM and COWL patterns. Characteristics of downstream wave propagation in NAM composites reinforce the idea that the Pacific center of action of the NAM (i.e., the low-frequency ridge/trough anomaly in the northeastern Pacific) and the Atlantic center of action (i.e., the NAO) are dynamically related. The Pacific ridge anomaly of the positive NAM phase favors the equatorward propagation of synoptic waves from the North Pacific to the North Atlantic, whereas the Pacific trough anomaly of the negative NAM phase favors more a zonal propagation. In COWL composites, the action of the northeast Pacific anomalies on synoptic wave propagation is suppressed by that of the North American anomalies, which reverse the initial orientation of the synoptic eddies coming from the North Pacific. The contrasting relationship between the PNA and NAO has been related to the NAM and COWL patterns. The NAM pattern, which exhibits a ridge (trough) northeast Pacific anomaly upstream of the positive (negative) NAO dipolar structure, corresponds to opposite-sign PNA and NAO composites. On the other hand, the COWL pattern, which shows an eastern Pacific large-scale trough (ridge) and a North American large-scale ridge (trough) upstream of the positive (negative) NAO dipolar anomaly, fits well with the same-sign PNA and NAO composites. It is interesting to note that this COWL variability also corresponds to in-phase fluctuations of the amplitude of the Pacific and Atlantic storm tracks, as shown by Chang and Fu (2002), but the reasons for that are unclear.

The DRA13 mechanism provides a new way of interpreting the two leading modes of variability of the Northern Hemisphere. Honda and Nakamura (2001) emphasized a downstream influence of the Pacific center of action on the Atlantic one but in terms of the Aleutian low–Icelandic low seesaw. Their mechanism differs from ours, as they have shown that the formation of an Aleutian low (high) anomaly leads to the formation of an Icelandic high (low) anomaly via quasi-stationary Rossby wave propagation in mid-to-late winter. They also underlined that synoptic wave activity tends to reinforce the previous quasi-stationary anomalies, as many studies have shown in the past decades (e.g., Lau 1988; Branstator 1995). Our study and that of DRA13 show that synoptic waves do not simply maintain low-frequency anomalies locally but have a remote effect. Upstream low-frequency anomalies modify synoptic wave propagation on their downstream side to trigger new low-frequency anomalies more downstream via wave-breaking processes. The two mechanisms are not exclusive but do not probably act at the same time scales. The propagation of low-frequency Rossby wave trains usually involves greater time scales than 10 days, while the propagation of synoptic waves involves sub-weekly time scales. Future studies should investigate if the present mechanism can explain rapid fluctuations of the NAM and COWL index.

To conclude, it is not because the mean flow exhibits strong zonal asymmetries, as in the COWL pattern or in the same-sign PNA and NAO composites, that downstream wave propagation is not important. This is to be contrasted with one conclusion of Li and Lau (2012a). According to them, zonal asymmetry during their weak cases (i.e., the positive ENSO phase associated with the
positive NAO phase and the negative ENSO phase associated with the negative NAO phase) prevents downstream wave propagation to occur between the North Pacific and the North Atlantic. Our results rather suggest that, even though zonal asymmetries tend to attenuate synoptic wave amplitude, they alter the direction of wave propagation downstream of them and thus determine in large part the type of wave breaking in the North Atlantic sector.

Acknowledgments. This study has benefited from discussions with Pablo Zurita-Gotor, Christophe Cassou, Nick Hall, Masa Kageyama, Francis Codron, and Fabio D’Andrea. The authors would like to acknowledge the two anonymous reviewers for their relevant suggestions, which helped to improve the clarity and significance of the paper.

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