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ABSTRACT

In Part I of this study, it is revealed that decadal variations of European blocking, in its intensity, duration, and position, during 1978–2011 are modulated by decadal changes in the frequency of North Atlantic Oscillation (NAO) events associated with background Atlantic conditions. In Part II, reanalysis data are analyzed to first show that a T-bone-type structure of the climatological-mean blocking frequency in the Euro-Atlantic sector roughly results from a combination of the blocking frequency distributions along the southeast–northwest (SE–NW) direction associated with negative-phase NAO (NAO\(^-\)) events and along the southwest–northeast (SW–NE) direction associated with positive-phase NAO (NAO\(^+\)) events.

A nonlinear multiscale interaction (NMI) model is then used to examine the physical processes behind the blocking frequency distributions. This model shows that the combination of eastward- and westward-displaced blocking frequency patterns along the SW–NE and SE–NW directions associated with NAO\(^+\) and NAO\(^-\) events leads to a T-bone-type frequency distribution, as seen in reanalysis data. Moreover, it is found that the westward migration of intense, long-lived blocking anomalies over Europe following NAO\(^+\) events is favored (suppressed) when the Atlantic mean zonal wind is relatively weak (strong). This result is held for the strong (weak) western Atlantic storm track. This helps explain the findings in Part I. In particular, long-lived blocking events with double peaks can form over Europe because of reintensification during the NAO\(^-\) decay phase, when the mean zonal wind weakens. But the double-peak structure disappears and becomes a strong single-peak structure as the mean zonal wind strengthens.

1. Introduction

In Luo et al. (2015, hereafter Part I), we have examined the relationship between the European blocking (EB) activity and the Atlantic background conditions using reanalysis data. We demonstrated that the EB events are intense and long lived and tend to occur in eastern Europe (EE) during 1978–94 (P1), but in western Europe (WE) during 1995–2011 (P2). The decadal changes in the EB events from P1 to P2 are found to be closely related to the decadal changes in the number of the North Atlantic Oscillation (NAO) positive (NAO\(^+\))- and negative (NAO\(^-\))-phase events, with more NAO\(^+\) (NAO\(^-\)) events during P1 (P2) associated with stronger (weaker) Atlantic mean zonal wind and weaker (stronger) western Atlantic storm track during P1 (P2), as seen in reanalysis data. Because the EB events associated with NAO\(^+\) (NAO\(^-\)) events extend more eastward (westward), the decadal changes in the number of NAO\(^+\) and NAO\(^-\) events from P1 to P2

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result in more EB events in eastern (western) Europe during P1 (P2). However, the physical mechanism of the decadal change in the EB events associated with NAO events from P1 to P2 is not fully understood from a theoretical perspective.

Although previous studies have demonstrated that NAO$^+$ events favor the occurrence of EB events (Shabbar et al. 2001; Scherrer et al. 2006; Croci-Maspoli et al. 2007), there does not exist a one-to-one correspondence between NAO$^+$ and EB events (Luo et al. 2007a). Some investigations have also revealed that the climatological blocking frequency pattern in the Euro-Atlantic sector exhibits a T-bone-type structure (Scherrer et al. 2006; Davini et al. 2012a,b; Part I). However, it is unclear why the blocking frequency pattern possesses such a T-bone-type structure. In this paper, we first further examine this structure using reanalysis data and then apply a theoretical model to study the physical processes behind this structure in EB frequency. Then the cause of the decadal variations of EB events is explored using this theoretical model.

At present, there is a lack of a theory or model that can describe the whole life cycle (onset, intensification, maintenance, and decay) of a blocking flow in terms of its duration, intensity, and location, although previous blocking theories can explain the maintenance of a blocking flow (McWilliams 1980; Shutts 1983; Haines and Marshall 1987; Yamazaki and Itoh 2013), with the possible exception of the nonlinear multiscale interaction (NMI) model of Luo (2000, 2005a,b,c), Luo and Chen (2006), and Luo et al. (2014). Besides the NMI model, other theoretical models can only capture a steady state of a dipole blocking flow, because the eddy vorticity flux divergence term is assumed to counteract the dissipation (Haines and Marshall 1987). Obviously, these models are unable to describe the evolution of a blocking flow from onset to decay. The dissipation is too strong to maintain a blocking flow even if the blocking is forced by the synoptic-scale eddies (Maeda et al. 2000). To distinguish from the NMI model, these models are referred to as the steady-state models herein. Thus, the steady-state models cannot be used to investigate how the life characteristics (intensity, duration, and position) of a blocking flow depend on atmospheric conditions. In particular, they cannot describe the mutual relationship between the phase of NAO and the variability of EB events, even though both the NAO and blocking events have the same time scales of 10–20 days (Feldstein 2003; Benedict et al. 2004) and appear to reflect different aspects of the related synoptic activities over the Euro-Atlantic sector.

Stochastic models have been used to depict the mean NAO patterns (Vallis et al. 2004), but these models can hardly describe the complicated nonlinear multiscale interaction process of NAO events. In particular, these NAO models are unable to simulate the evolution of individual NAO events for their duration, intensity, and location, as well as their relationship with European blocking events. In the past decade, the NMI model, originally developed by Luo (2000, 2005a) and further improved by Luo and Chen (2006) and Luo et al. (2007b, 2014) as a mathematical description of the nonlinear interaction among the mean flow and planetary- and synoptic-scale waves during a blocking life cycle, has been extended to investigate the dynamics of NAO events on time scales of 10–20 days driven by synoptic-scale eddies (Luo et al. 2007a,b, 2008a,b, 2011; Luo and Cha 2012). Comparative studies of this theoretical model with observations have shown that the NMI model can capture the essential characteristics of blocking and NAO events. For example, the NMI model results show that the eddy-driven NAO events exhibit time scales of 10–20 days, and the positive (negative) phase of NAO corresponds to the enhanced blocking frequency over the European continent (North Atlantic) (Luo et al. 2007a). These results are consistent with the observational findings of Shabbar et al. (2001) and Scherrer et al. (2006). Further, results from an extended NAO model that includes the effect of land–sea configuration show that the north–south shift of the Atlantic jet during the NAO life cycle is caused by the interaction between the NAO and stationary wave anomalies and is dependent on the phase of the NAO event (Luo et al. 2007b). This explains why the observed Atlantic jet undergoes a northward (southward) shift during the NAO$^+$ (NAO$^-$) life cycle. Interestingly, the latitudinal position of the preexisting Atlantic jet tends to determine the phase of subsequent NAO events (Luo et al. 2008b). Luo et al. (2011) further extended the NMI model to include the effect of the downstream development of synoptic-scale eddies (Chang 1993) and demonstrated that the decline of the positive NAO index during 1991–2009 is closely associated with the marked intensification of the Atlantic storm track. Recently, Luo et al. (2012a,b) found that the strengths of the Atlantic mean zonal wind and storm track are two precursor conditions for whether the NAO event is an individual (in situ, defined as an event that is not followed by an opposite event) event or regime transition event, thus explaining why NAO$^+$ (NAO$^-$) events are dominant during 1978–2010 (1950–77).

The key merit of the NMI model is its ability to unify both blocking and NAO events and to describe how the onset, intensification, maintenance, and decay, as well as the duration, intensity, and position of an NAO event or a blocking flow vary with atmospheric conditions, such as the strength of mean zonal wind, initial conditions,
synoptic eddies, and so on (Luo 2005a; Luo and Cha 2012; Luo et al. 2014). Using this model, NAO$^-$ (NAO$^+$) events can be identified as corresponding to the presence (absence) of cyclonic wave breaking over the North Atlantic (Luo et al. 2008a), in agreement with the observational results of Woolings et al. (2008). Thus, the NMI model is a useful tool for examining how the duration, intensity and position of EB events associated with NAO events depend on the Atlantic background conditions. Unlike in previous studies, in this paper this NMI model is used to reveal the mutual relationship between the NAO events and the frequency distribution of Euro-Atlantic blocking events and to provide an explanation for why intense and long-lived EB events tend to occur in EE (WE) during P1 (P2), as found in Part I of this study.

This paper is organized as follows. In section 2, we present the climatological frequency distribution of blocking events in the Euro-Atlantic sector based on a two-dimensional blocking index developed by Davini et al. (2012a,b). The results show that the climatological blocking frequency distribution is likely to be a combination of the different blocking frequency distributions associated with NAO$^+$ and NAO$^-$ events. Section 3 briefly describes the NMI model. In section 4, we use the NMI model to explain why the climatological blocking frequency distribution is mainly composed of NAO$^+$ and NAO$^-$ related blocking frequency distributions. This validates the ability of the NMI model in capturing the blocking activity. The impact of the Atlantic background conditions, such as the mean zonal wind and storm-track strengths on the EB activity is examined in section 5. Conclusions and discussions are summarized in section 6.

2. Blocking frequency pattern and its link with zonal winds

a. Climatology of blocking frequency

To understand the ability of the NMI model used below, it is useful to first examine whether the NMI model can explain the observed climatological blocking frequency distributions in the Euro-Atlantic sector. We first present the spatial distribution of the climatological blocking frequency derived from reanalysis data before introducing the NMI model.

The NCEP–NCAR reanalysis data and two-dimensional blocking index (Davini et al. 2012a,b) used here are the same as in Part I. Because the life cycles of NAO$^+$ and NAO$^-$ events are of 10–20-day time scales (Feldstein 2003; Benedict et al. 2004; Luo et al. 2007a), the composite of the blocking frequency associated with the phase of NAO is done during the period from lag $-10$ to lag $+10$ days, where lag 0 denotes the day with the strongest NAO anomaly at a given vertical level. The definition of an NAO event has been given in Part I. Figures 1a and 1b show the climatological (1978–2011) mean maps of the winter blocking frequency with and without NAO events, while the composite 500-hPa blocking frequency distributions associated with NAO$^+$ and NAO$^-$ events are shown in Figs. 1c and 1d when low-latitude blocking events excluded [subtropical high ridges are often identified as low-latitude blocking events only if the constraint condition in the blocking index of Davini et al. (2012a,b) is not used]. The time sequences of winter-mean NAO and EB indices during 1978–2011 are shown in Fig. 1e. It is seen in Fig. 1a that the mean blocking frequency exhibits a T-bone-type structure, with a center of high blocking frequency over the North Sea region and a secondary center over southern Greenland. However, this structure undergoes a significant change from P1 to P2, as found in Part I. It also looks similar to that noted by Scherrer et al. (2006), who used an absolute geopotential height index. Figure 1b shows that there are few blocking events in the Euro-Atlantic sector if no NAO events take place, which implies that most of the blocking events in this region are associated with NAO. This reflects a significant contribution of NAO events to the T-bone-type distribution of the climatological blocking frequency. We can further see that the composite blocking frequency is distributed along the southwest–northeast (SW–NE) direction associated with the NAO$^-$ events (Fig. 1c), but along the southeast–northwest (SE–NW) direction associated with the NAO$^+$ events (Fig. 1d). Thus, it is suggested that the climatological blocking frequency distribution as seen in Fig. 1a is a superimposition of the different blocking frequency patterns associated with NAO$^+$ and NAO$^-$ events. The winter EB index exhibits a positive correlation of 0.71 with the winter-mean NAO index during 1978–2011, and the correlation strength changes from 0.92 during P1 to 0.74 during P2 (Fig. 1e). The relatively low correlation during P2 may be associated with the westward migration of EB events associated with NAO events because of weaker mean zonal wind and stronger western Atlantic storm track (the strong storm track in the western Atlantic basin during P2, as found in Part I. Thus, the decadal variation of winter EB events from P1 to P2 is closely related to the decadal change in the winter NAO events. However, there does not exist a one-to-one relationship between NAO and EB events; that is, some EB events are not associated with NAO, and vice versa, although the blocking frequency not associated with NAO is low (Fig. 1b). As revealed in Luo et al. (2012b, their Fig. 1), the composite NAO$^+$ (NAO$^-$) anomaly dipole pattern undergoes a marked eastward (westward) shift as the zonal wind over the
midlatitude Atlantic strengthens (weakens) during its life cycle. As a result, the spatial pattern of the blocking frequency is inevitably influenced by the different phase and position of the NAO events.

While the distribution shape of the blocking frequency associated with NAO is determined mainly by the NAO phase, the westward or eastward displacement of the NAO pattern during its life cycle seems to affect the overall position of the blocking frequency distribution. Some studies have indicated that the eastward (westward) displacement of the NAO pattern is closely related to the enhanced (reduced) zonal wind strength over the mid- to high-latitude Atlantic (Luo and Gong 2006; Luo et al. 2007b; Davini et al. 2012b). Below, we demonstrate that the eastward (westward) shift of the NAO$^+$ (NAO$^-$) pattern likely results from the self-maintained strengthening (weakening) of the midlatitude zonal wind over the Atlantic, and this helps explain the observed distribution of the blocking frequency.

\textbf{b. Zonal wind anomalies associated with NAO$^+$ and NAO$^-$ events}

Here we define the daily zonal wind anomaly as a deviation from the winter-mean zonal wind during the life cycle of an NAO event over the North Atlantic. Figures 2 and 3 show the composite of the daily zonal wind anomalies for the two phases of NAO events during the time period from lag $-4$ to lag $+6$ days. It is seen that for the negative phase the composite zonal wind anomalies exhibit an alternative structure in the meridional direction, with negative anomalies from $40^\circ$ to $65^\circ$N and positive anomalies to the south and north of it. Such
a spatial structure is most evident at lag 0 days, when the negative anomalies reach 18 m s\(^{-1}\) (Fig. 2). However, for the positive phase the zonal wind anomalies show an opposite structure, with positive anomalies around 40\(^\circ\)-65\(^\circ\)N and negative anomalies to the south and north of this band (Fig. 3). Thus, the weakening (strengthening) of zonal winds over the mid- to high-latitude Atlantic during the negative (positive) phase of NAO is an important factor for the westward (eastward) movement of the NAO\(^+\) (NAO\(^-\)) pattern (Luo et al. 2012b).

The zonal movement of NAO patterns can also be understood in terms of the wave breaking concept (Thomcroft et al. 1993). As revealed by Woollings et al. 2008 and Sung et al. (2011), the NAO\(^-\) pattern arises from the retrogression (i.e., westward migration) of the European blocking because of the cyclonic wave breaking (CWB) as the Atlantic jet stream over the middle to high latitudes weakens owing to its southward shift (Franzke et al. 2004). Thus, a westward displacement of the NAO\(^-\) pattern follows the weakening of zonal winds over the mid- to high-latitude Atlantic. In contrast, the anticyclonic wave breaking (AWB) due to the northward shift and strengthening of the Atlantic jet over the middle to high latitudes can excite NAO\(^+\) events (Benedict et al. 2004) and allow their eastward displacement during its life cycle.

3. A weakly nonlinear multiscale interaction model of NAO events

In this section, a simplified NMI model of NAO events developed by Luo et al. (2007a,b) and Luo and...
Cha (2012) is applied to produce the spatial blocking frequency patterns associated with $\text{NAO}^-$ ($\text{NAO}^+$) events (Figs. 1b,c). It is further applied to explain the decadal change in the blocking events over Europe from 1978–94 to 1995–2011 found in Part I.

a. Model description and scale-interaction equations

Benedict et al. (2004) and Franzke et al. (2004) suggested that NAO events are in essence a nonlinear initial-value problem. This hints that the occurrence of the NAO event is a complicated nonlinear multiscale interaction process, which corresponds in essence to the evolution of an initial state to a strong meridional (NAO$^-$) or zonal flow (NAO$^+$) state (Luo et al. 2007a; Woollings et al. 2008). This also motivated us to develop an NMI model to describe life cycles of individual NAO events that evolve from initial states (Luo et al. 2007a,b; Luo and Cha 2012), even though the NAO events have a time scale of 10–20 days (Feldstein 2003). In the following sections, the NMI model is used to identify how the EB activity is modulated by the phase of NAO and how the duration, intensity, and position of EB events vary with Atlantic background conditions.

The NMI model is developed by Luo and his co-authors from a nondimensional equivalent barotropic potential vorticity (PV) equation of atmospheric motions (Luo 2000, 2005a; Luo et al. 2007a). This model was supported by earlier findings (Hurrell 1995; Feldstein 2003; Jin et al. 2006) that a barotropic model can simulate the main features of NAO events, even though the baroclinic processes (i.e., the stirring of baroclinic synoptic eddies) can also affect NAO events to some degree (Vallis et al. 2004). Thus, a barotropic model, such as the NMI model, may be acceptable for mechanism investigations, although the relative importance

![Figure 3](image-url)
of barotropic and baroclinic processes in the dynamics of the NAO requires further examination.

Since this NMI model can establish a link between the EB activity and the phase of the NAO events (Luo et al. 2007a), it can be used to examine the physical processes through which the Atlantic conditions affect the recent decadal relationship between the NAO and EB events during 1978–2011. Here we briefly describe the NMI model; more detailed information can be found in Luo et al. (2007b) and Luo and Cha (2012).

To describe the scale interaction during the NAO life cycle, as in Luo (2005a) and Luo et al. (2007a,b), the total streamfunction \( \Psi \) is split into three parts: the basic-flow \( [\Psi(y)] \), planetary-scale \( (\psi) \), and synoptic-scale \( (\psi') \) streamfunction anomalies. The governing equations of the interaction between planetary- and synoptic-scale components can be obtained under the zonal scale-separation assumption. The condition for the zonal scale-separation assumption was discussed in detail in Luo (2005a).

It has been recognized that the variability of the North Atlantic jet stream is modulated by teleconnection patterns, such as NAO and eastern Atlantic (EA) patterns (Strong and Davis 2008; Woollings and Blackburn 2012). The jetlike structure of the mean zonal wind is weak when there are no teleconnection patterns in the Atlantic (Luo et al. 2007a; Luo et al. 2005a; Luo and Cha 2012).

Figure 4 shows the winter-mean climatological stationary wave (CSW) anomaly during 1978–2012 based on the NCEP–NCAR reanalysis. It is clear that the CSW anomaly exhibits a monopole meridional structure with a wavenumber of 2, in which the strongest positive center is located near 10\(^{\circ}\)W. The CSW anomaly may be considered a winter standing wave induced by the land–sea configuration, as described by h in (1), because its spatial structure is almost independent of the phase of NAO events (Luo et al. 2007b). For this reason, we assume \( h = h_0 \exp \left[ ik(x + x_T) \right] \sin \left( \frac{m}{2} \theta \right) + cc \) according to Charney and DeVore (1979), where \( h_0 = 1/(6.571 \cos \phi_0); m = \pm 2\pi/L_v; \phi_0 = \) the latitude; \( L_v = 5 \) is used in this paper, which is equivalent to 5000 km in a dimensional form; and \( x_T \) denotes the zonal position of the positive anomaly of the topography-induced standing wave relative to the NAO pattern; and \( cc \) denotes the complex conjugate of its preceding term.

Since our focus is on the impact of the Atlantic conditions on the EB variability linked to NAO events, the major mountain ranges, such as the Rockies and the Himalayas, are not considered in the present study, even though they may significantly affect the Northern Hemisphere (NH) stationary waves (Held et al. 2002) and storm track (Brayshaw et al. 2008; Woollings et al. 2011). Nevertheless, the variability of EB events associated with the CSW anomaly and Atlantic storm-track variations can be examined by varying the intensity and location of the CSW anomaly and storm track in our highly idealized model.

\[
\frac{\partial}{\partial t} + u_0 \frac{\partial}{\partial x} (\nabla^2 \psi - F\psi) + J(\psi, \nabla^2 \psi + h) \\
+ (\beta + Fu_0) \frac{\partial h}{\partial x} + u_0 \frac{\partial h}{\partial x} = -J(\psi', \nabla^2 \psi')_P \quad \text{and} \quad (1a)
\]

\[
\frac{\partial}{\partial t} + u_0 \frac{\partial}{\partial x} (\nabla^2 \psi' - F\psi') + (\beta + Fu_0) \frac{\partial \psi'}{\partial x} \\
= -J(\psi', \nabla^2 \psi' + h) - J(\psi, \nabla^2 \psi' + \nabla^2 \psi'^c, (1b)
\]

where \( \beta \) is the nondimensional meridional gradient of the Coriolis parameter, \( h \) is the nondimensional topographic variable, and \( F = (L/R_d)^2 \), where \( L \) and \( R_d \) are, respectively, the characteristic scale and Rossby deformation radius, which is assumed to be of 1000 km in this study. The subscript \( P \) denotes the planetary-scale zonal wavenumber of \(-J(\psi', \nabla^2 \psi')\), which is close to that of the NAO anomaly. Moreover, \( \nabla^2 \psi'^c \) is a prescribed synoptic-scale vorticity source introduced to maintain synoptic-scale eddies in the Atlantic storm track, as described by (2i) presented below. The other parameters and notations can be found in Luo et al. (2007b) and Luo and Cha (2012).
b. Analytical solutions of the nonlinear multiscale interactions associated with NAO events

As in Luo (2005a) and Luo et al. (2007a,b), it is easy to obtain the analytical solutions of planetary- and synoptic-scale components from (1a) and (1b) using a multiple-scale perturbation expansion. In this paper, we assume that the zonal and meridional wavenumbers of the NAO dipole anomaly are \( k \) and \( m \), respectively. Because the NAO event is a zonally isolated mode and undergoes a dominant zonal movement (Luo et al. 2012b, their Fig. 1), here we only consider the impact of the mean zonal wind strength change on it. For this purpose, it is useful to assume that \( u_0 = u_C + \Delta u \), \( u_C = \beta/k^2 + m^2 \), and \( |\Delta u| < u_C \). Here, \( u_C \) is a critical mean zonal wind speed that the planetary wave prior to the NAO onset is allowed to be stationary. Different values of \( \Delta u \) are used to characterize the different strengths of the Atlantic mean zonal wind prior to the NAO onset. Here, \( \Delta u > 0 \) (\( \Delta u < 0 \)) corresponds to a strong (weak) background westerly wind.

Following Luo et al. (2007b) and Luo and Cha (2012), the analytical solution of the atmospheric streamfunction for an NAO event can be expressed as

\[
\Psi_T = -u_0 y + \psi + \psi' = \psi_p + \psi',
\]

\[
\psi_p = -u_0 y + \psi = -u_0 y + \psi_{\text{NAO}} + \psi_C + \psi_m, \quad (2b)
\]

\[
\psi_{\text{NAO}} = B \sqrt[4]{2 \frac{L_y}{B}} \exp(ikx) \sin(my) + cc, \quad (2c)
\]

\[
\psi_C = h_A h_0 \exp[i(kx + x_T)] \sin \left( \frac{m}{2} y \right) + cc, \quad (2d)
\]

\[
\psi_m = \psi_{m1} + \psi_{m2}, \quad (2e)
\]

\[
\psi_{m1} = -|B|^2 \sum_{n=1}^{\infty} \tilde{a}_n (3a_n - b_n) \cos(nmy), \quad (2f)
\]

\[
\psi_{m2} = -h_0 h_A \sqrt[4]{2 \frac{L_y}{B}} (B e^{-ikx} + B^* e^{ikx}) \times \sum_{n=1}^{\infty} \tilde{a}_n (3a_n - b_n) \cos(nmy), \quad (2g)
\]

\[
\psi' = e^{32/3} (\tilde{\psi}_0' + \tilde{\psi}_1') = \psi_1' + \psi_2', \quad (2h)
\]

\[
\psi_1' = e^{32/3} \tilde{\psi}_0' = f_0(x) \left[ \exp[i(\tilde{k}_1 x - \tilde{\omega}_1 t)] \right. \\
\left. + \alpha \exp[i(\tilde{k}_2 x - \tilde{\omega}_2 t)] \right] \sin \left( \frac{m}{2} y \right) + cc, \quad (2i)
\]

\[
\psi_2' = -m 4 \sqrt[4]{2 \frac{L_y}{B}} h_0 \sum_{j=1}^{2} Q \alpha_j \exp[i(\tilde{k}_j + k)x - \tilde{\omega}_j t] \left[ p_j \sin \left( \frac{3m}{2} y \right) + r_j \sin \left( \frac{m}{2} y \right) \right]
\]

\[
+ m 4 \sqrt[4]{2 \frac{L_y}{B}} f_0 h_0 \sum_{j=1}^{2} Q \alpha_j \exp[i(\tilde{k}_j - k)x - \tilde{\omega}_j t] \left[ s_j \sin \left( \frac{3m}{2} y \right) + h_j \sin \left( \frac{m}{2} y \right) \right]
\]

\[
- m 4 f_0 h_0 \sum_{j=1}^{2} \pi_j \alpha_j \exp[i(\tilde{k}_j + k)x + kx_T - \tilde{\omega}_j t] \sin(2my)
\]

\[
- m 4 f_0 h_0 \sum_{j=1}^{2} \sigma_j \alpha_j \exp[i(\tilde{k}_j - k)x - kx_T - \tilde{\omega}_j t] \sin(2my) + cc, \quad (2j)
\]

**Fig. 4.** The 1978–2011 mean winter geopotential height anomaly deviated from a zonal mean at 500 hPa, in which the solid (dashed) lines represent positive (negative) values.
where \( i = \sqrt{-1}; \alpha_1 = 1; \alpha_2 = \alpha; \ h_A = -1/[\beta u_c - (k^2 + m^2/4)]; \ |B|^2 = BB^*; \alpha_1 = 1 (\alpha = -1) \) denotes the positive (negative) phase of the NAO event; \( B^* \) is the complex conjugate of \( B; \hat{k}_i \) and \( \bar{\omega}_i (i = 1, 2) \) are the zonal wavenumbers for each component of the synoptic-scale eddies with a period less than 1 week (Luo 2005a; Luo et al. 2007a), and its corresponding frequency, \( f_0(x) = a_0 \exp[-\mu(x^2 + x_0^2)] \), represents the spatial distribution of the eddy amplitude for \( \mu > 0; a_0 \) is the eddy strength in the western side of the Atlantic basin for \( x_0 > 0 \), referred to as the western Atlantic storm-track strength; and \( x_0 \) represents the zonal position of the maximum eddy intensity. The other coefficients and notations can be found in Luo et al. (2007b) and Luo and Cha (2012), except that \( u_0 \) in those coefficients should be replaced by \( u_c \).

It is clear in (2) that \( \psi_p \) denotes the planetary-scale field of the NAO event, \( \psi_{NAO} \) represents the NAO anomaly, and \( \psi_c \) denotes the topography-induced stationary monopole wave, which is considered as an approximation of the observed CSW anomaly in the NH.

For the positive (negative) phase, \( m = 2\pi/L_y (m = -2\pi/L_y) \) is required in (2d) (Luo et al. 2007a,b). However, in (2i), \( \alpha = 1 \) (\( \alpha = -1 \)) is required so that the NAO\(^+\) (NAO\(^-\)) anomaly can be reinforced by the eddy vorticity forcing \( -f(\psi_1^*, \nabla \psi_1^*)_{x_p} = -\nabla \cdot (\psi_1^* q_1^*)_{x_p} \), where \( \psi_1^* = (-\bar{\omega}_1 \partial_x x_0); \bar{\omega}_1 \bar{\psi}_1^* / \partial x \) and \( q_1^* = \nabla^2 \psi_1^* \). Note that \( \psi_1^* \) represents the streamfunction field of Atlantic storm-track eddies without teleconnection patterns maintained by the synoptic-scale vorticity source \( \nabla^2 \psi_1^* \), while \( \psi_2^* \) in (2j) represents eddies induced by the feedback of the NAO anomaly driven by preexisting synoptic eddies and the CSW anomaly. In (2e)-(2g), \( \psi_{ml} \) and \( \psi_{m2} \) denote the mean westerly wind parts induced by the nonlinear interaction of the NAO anomaly itself and its interaction with the CSW anomaly. For this case, the time-dependent mean westerly wind during the NAO life cycle can be expressed as \( u_M = u_0 - \delta(\psi_{m1} + \psi_{m2})/\partial y \). It is clear that there is \( u_M = u_0 \) in the absence of NAO events. In this case, \( u_M = u_0 \) may be referred to as the mean zonal wind prior to the NAO onset. In fact, observations show that the Atlantic mean zonal wind prior to the NAO onset exhibits a weak jet structure (Luo et al. 2007b, their Fig. 1). Thus, it is reasonable to assume that the Atlantic mean zonal wind is approximately uniform in both the zonal and meridional directions. The case of \( u_0 \) having a weak jet structure has been examined in Luo et al. (2008b). In (2), the evolution of each component in the analytical solution of the NAO event can be predicted once the amplitude \( B(x, t) \) of the NAO anomaly is known from an initial state.

As in Luo and Cha (2012), the evolution equation of \( B(x, t) \) that satisfies a forced nonlinear Schrödinger (NLS) equation can be expressed as

\[
\left( \frac{\partial B}{\partial t} + C \frac{\partial B}{\partial x} \right) + \lambda \frac{\partial^2 B}{\partial x^2} + \delta |B|^2 B + \delta h_0^2 (B + B^* e^{2i\omega t}) + \Delta u \Gamma B + G f_0^2 \exp(-i(\Delta x + \Delta \omega t)) = 0, \tag{3}
\]

where \( \Gamma = -k(k^2 + m^2)(k^2 + m^2 + F), \Delta k = k - (\hat{k}_2 - \hat{k}_1), \Delta \omega = \bar{\omega}_2 - \bar{\omega}_1, \lambda > 0, \delta > 0 \) and the mathematical expressions of \( \lambda \) and \( \delta \) and other coefficients are the same as in Luo et al. (2007b) and Luo and Cha (2012).

In the numerical experiments of Arai and Mukougawa (2002), they considered \( \pi L_y = 10000 \) and \( \pi L_y = 21000 \) km as the narrow and wide channels, respectively, in which the wide channel \( L_y \) is about 6685 km (\( L_y = 6.685 \) in nondimensional form). They found that the blocking solutions that describe the positive-feedback effect of synoptic eddies on block maintenance are sensitive to the width of the channel. However, in our NMI model, the feedback of synoptic-scale eddies on the blocking flow does not depend strongly on the width of the \( \beta \) channel (not shown).

Our calculations show that when \( L_y \geq 6.63 \), there is \( \lambda \delta < 0 \) in (3) because of \( \lambda > 0 \) and \( \delta < 0 \). In this case, the NLS equation [(3)] has no localized soliton solutions. Instead, it has a dark soliton solution (Hasegawa and Kodama 1995). Thus, no localized dipole-structure solutions exist for \( L_y \geq 6.63 \). This suggests that a localized blocking or NAO structure cannot be excited by synoptic-scale eddies when the channel width is too large (\( L_y \geq 6.63 \)). However, when \( L_y \leq 6.62 \), there exists \( \lambda \delta > 0 \) in the NMI model, and a localized dipole blocking or NAO solution can exist in this case. Thus, in numerical experiments of blocking, the width of the channel should not exceed about 6620 km.

In the NMI model, we found that the model results are insensitive to the channel width as long as it is less than \( L_y = 6.62 \) (6620 km in dimensional form). Interestingly, the blocking strength decreases as the channel width increases from \( L_y = 4.8 \) (4800 km) to \( L_y = 5.2 \) (5200 km in dimensional form) (not shown).

It is possible to use solution (2) to examine the relationship between NAO events and downstream blocking circulation if the solution to (3) is obtained by using a finite-difference scheme similar to that used in Luo (2005a) for given initial conditions and parameters. In this model, the strength of the western Atlantic storm track is represented by the eddy strength \( a_0 \), and we consider the value of \( \Delta u \) as a measure of the variation in mean zonal wind over the Atlantic basin. The model parameters and initial conditions used in this study are listed in Table 1. In this paper, we only show results for the \( B(x, 0) = B_0 = 0.4 \) (equivalent to the amplitude of 40 gpm in a dimensional form) case in order to see if
a uniform planetary wave as an initial condition of a blocking or NAO event can be reinforced into a localized blocking or NAO flow by preexisting synoptic-scale eddies. Here, the parameters \( x_T \) and \( x_0 \) are chosen to be different from our previous studies (Luo et al. 2007b; Luo and Cha 2012). The choice of the two parameters is based on the relative position between the source region of NAO\(^+\) (NAO\(^-\)) events and the positive center of the CSW anomaly.

4. Model results

a. Topography-induced standing wave and Atlantic storm track in the NMI model

Because \( m = 2\pi/L_y \) (\( m = -2\pi/L_y \)) is required for the positive (negative) phase of NAO in (2c), \( h_0 < 0 \) (\( h_0 > 0 \)) must be required in order to make the CSW anomaly, as denoted by \( \psi_c = h_A h_0 \exp[ik(x + x_T)] \sin[(m/2)y] \) + cc, independent of the phase of the NAO anomaly (Luo et al. 2007b). In the NMI model, the center of the initial NAO anomaly is assumed to be at \( x = 0 \) for two phases of the NAO event, as shown in Figs. 5a and 5c. Observations show that NAO\(^+\) (NAO\(^-\)) events originate from Greenland (northern Europe) (Luo et al. 2012b, their Fig. 1). Thus, \( x_T < 0 \) must be assumed so that the positive center of the topographically induced standing wave (CSW anomaly) is located in the downstream region of the NAO\(^+\) pattern. In contrast, the source region of NAO\(^-\) events is over northern Europe and almost in the same zonal position as the positive center of the CSW anomaly. So we can choose \( x_T = 0 \) for the negative phase of the NAO event.

As noted above, in the NMI model we have assumed that the observed CSW anomaly (Fig. 4) is represented by the standing wave induced by the wavenumber-2 topography with a monopole meridional structure. To see whether the topographic standing wave is an approximation of the observed CSW anomaly, we show the horizontal distributions of the standing wave in Fig. 5b for the negative phase (\( h_0 = 0.4, m = -2\pi/L_y \)) and \( x_T = 0 \). However, the positive center of the topographic standing wave can be located downstream of the NAO\(^+\) region if a negative value of \( x_T \) is chosen for the positive phase (\( h_0 = -0.4, m = 2\pi/L_y \)), as shown in Fig. 5d for \( x_T = -1.0 \). It is clear (Fig. 4) that the observed CSW anomaly can be better represented by the topographic standing wave (Figs. 5b,d). In the NMI model, synoptic eddies are assumed to preexist, and they can organize to form an Atlantic storm track that could exist without NAO events, although these eddies propagate eastward. Figure 5e (Fig. 5f) shows the horizontal distribution of synoptic eddies at \( t = 0 \) for the parameters \( a_0 = 0.17 \) and \( x_0 = 3 \times 2.87/4 \) \( (x_0 = 2.87/2) \) that drive NAO\(^-\) (NAO\(^+\)) events (the nondimensional wavelength of the wavenumber 2 is \( 4 \times 2.87 \) around the whole circle at \( 55^\circ N \)). In Fig. 5, we only plotted the domain of one wavelength, where \( (x, y) = (1.0, 1.0) \) denotes \( (x, y) = (1000 \text{ km}, 1000 \text{ km}) \) in a dimensional form. The North Atlantic (European continent) is roughly represented by the region from \( x = -5.74 \text{ to } x = 0 \) (from \( x = 0 \) to \( x = 5.74 \)) for the negative phase because the initial NAO\(^-\) anomaly is assumed to be fixed at \( x = 0 \). For this case, the western (eastern) part of the region from \( x = 0 \) to \( x = 5.74 \) is defined as the western (eastern) Europe. Correspondingly, for the positive NAO phase, we choose \( x_T = -1.0 \) so that the positive center of the CSW anomaly is located 1000 km downstream of the initial NAO\(^+\) region. Thus, for the positive phase, the western (eastern) part of the region from \( x = 1.0 \text{ to } x = 6.74 \) (the region for \( x > 5.74 \) is not plotted) is also defined as western (eastern) Europe to ensure that the range of the European continent is the same as that of the negative NAO phase.

In the NMI model, the derivation of the analytical solutions of blocking or NAO events is based upon the zonal-scale separation assumption (Luo 2005a; Luo et al. 2007a,b). This assumption requires that the prescribed synoptic-scale eddies must have zonal wavenumbers greater than or equal to 9 if the blocking or NAO event is assumed to have a wavenumber-2 structure and is nearly quasi stationary. This assumption is acceptable for a mechanism study.

b. Blocking events, NAO phase, and zonal wind anomalies

The strongest synoptic eddies are usually located upstream of or in the same position as the NAO\(^+\) (NAO\(^-\)
Thus, in following calculations, we choose $x_0 = 2.87/2$ ($x_0 = 3 \times 2.87/4$) as the position parameter for the strongest synoptic eddies that drive NAO$^+$ (NAO$^-$) events. However, our main results are insensitive to the choice of $x_0$ (not shown), even though the value of $x_0$ varies slightly with individual NAO$^+$ (NAO$^-$) events.

For given synoptic-scale eddies with $a_0 = 0.17$ and $\Delta u = 0$, the planetary-scale ($\psi_p$), anomaly ($\psi_{\text{NAO}}$), and total streamfunction ($\Psi_T$) fields of an NAO event are shown in Fig. 6 for the negative phase and in Fig. 7 for the positive phase. Figure 6a shows that the planetary-scale field of an NAO$^-$ event exhibits the life cycle of an omega-type blocking structure. This blocking flow undergoes a westward movement as it intensifies (Fig. 6b). Its central position is changed from the source region near $x = 0$ at day 0 to $x = -1$ at day 9. Because the region near $x = -1$ roughly represents the North Atlantic or Greenland, the blocking event over these regions can be considered as an NAO$^-$ event (Luo et al. 2007a,b). The occurrence frequency of EB events is reduced along with the retrogression of the blocking flow. This hints that the EB frequency is reduced over Europe during the growing process of NAO$^-$.

Thus, our result suggests that a blocking event over the North Atlantic is equivalent to an NAO$^-$ event, and they are just two different
FIG. 6. (a) Planetary-scale (CI = 0.3), (b) anomaly (CI = 0.2), and (c) total (CI = 0.3) fields of the streamfunction of an NAO event obtained from the NMI model. The solid (dashed) lines denote positive (negative) anomalies.
ways to describe the same synoptic event (analog to the two faces of the same coin). This result is consistent with Woollings et al. (2008) and Davini et al. (2012a). In particular, the instantaneous field of the total streamfunction (Fig. 6c) resembles the observed evolution of a blocking or NAO+ event (Berggren et al. 1949; Benedict et al. 2004). Moreover, the CWB (Benedict et al. 2004; Woollings et al. 2008), as indicated by the northward (southward)
movement of small-scale warm (cold) air marked with red (green) color in Fig. 6c, can be seen during the establishment of an NAO\(^-\) or blocking event (Fig. 6c from day 0 to day 9). This implies that the simplified model used here can capture the CWB phenomenon and other main characteristics of the observed NAO\(^-\) or blocking events.

Figure 7 shows the evolution fields of an initial NAO\(^+\) pattern under the forcing of synoptic-scale eddies. From day 0 to day 9, an intensification of a zonal flow represented by a low-over-high dipole is seen in the planetary-scale field (Fig. 7a). This is the characteristic of an NAO\(^+\) event. Furthermore, a blocking event is seen to occur over the European continent during the decay phase of the NAO\(^+\) event. Thus, the EB frequency is enhanced during the NAO\(^-\) phase. However, we can see the westward displacement of the NAO\(^+\) pattern in the anomaly field as the NAO\(^-\) intensifies (Fig. 7b). This is different from our finding in Part I, where the composite NAO\(^+\) pattern exhibits an eastward shift. A main reason is that the phase-speed change of the NAO pattern induced by zonal wind anomalies cannot be considered in our simplified model, because the zonal wind change in response to the NAO evolution is considered as a second-order solution of the model equation (Luo et al. 2007a,b). Even so, the results obtained from our theoretical model show a high consistency with those from the reanalysis data, besides the zonal movement of the NAO\(^+\) anomaly. In the total field of an NAO\(^+\) event, the CWB is almost invisible over the North Atlantic (Fig. 7c). Thus, to some extent, the NAO\(^+\) event corresponds to the absence of CWB (i.e., blocking days), supporting the findings of Woollings et al. (2008), who noted that the NAO\(^-\) (NAO\(^+\)) event essentially corresponds to the presence (absence) of blocked days over Greenland (Davini et al. 2012b). Because the Atlantic jet \(-\partial \psi_m/\partial y\) has the cyclonic (anticyclonic) shear (not shown), the NAO\(^-\) (NAO\(^+\)) event can also be explained in terms of the CWB (AWB). This is consistent with the wave breaking explanation of NAO events (Benedict et al. 2004). Thus, the NMI model cannot only capture the main features of NAO events, but also describe the relationship between the EB activity and the phase of the NAO event. Results shown in Figs. 6 and 7 are used below to explain why the observed climatological blocking frequency in the Euro-Atlantic sector exhibits a T-bone-type structure.

To understand if the blocking frequency distribution associated with the phase of NAO depends upon the different spatial structure of zonal winds for different NAO phase, it is necessary to calculate the variation of the zonal component anomaly of geostrophic winds \(u_A = -\partial (\psi_{\text{NAO}} + \psi_C + \psi_m)/\partial y\) obtained from the NMI model during the NAO life cycle. We show the geostrophic zonal wind anomaly during the NAO life cycle in Fig. 8 for the same parameters as in Figs. 6 and 7. We can see that for the negative (positive) NAO phase, the zonal wind is reduced (enhanced) in midlatitude regions (Fig. 8), consistent with the composite result shown in Figs. 2 and 3. In the real world, the enhanced (reduced) zonal wind in middle to high latitudes should increase (decrease) the phase speed of the NAO\(^+\) (NAO\(^-\)) anomaly, which should lead to an eastward- (westward-) displaced NAO\(^+\) (NAO\(^-\)) pattern, as seen from the reanalysis data (Luo et al. 2012b, their Fig. 1). However, the eastward shift of the NAO\(^+\) pattern cannot be seen from our theoretical results (Fig. 7), because the zonal wind-induced phase-speed change cannot be considered in our model if the same parameters are chosen for both phases. Even so, it is concluded that the eastward (westward) shift of the NAO\(^+\) (NAO\(^-\)) anomaly due to strong (weak) zonal wind is a self-maintaining phenomenon in that the enhanced (reduced) zonal wind in middle to high latitudes is concomitant with the intensification of the NAO\(^+\) (NAO\(^-\)) anomaly (Luo et al. 2007b). Of course, the westward shift of the NAO\(^+\) anomaly is suppressed, and even an eastward shift is seen if \(\Delta u\) is chosen to be a positive value in our NMI model (as shown in Fig. 12 below).

c. Physical cause of the T-shape distribution of the Euro-Atlantic blocking frequency

Based on the anomaly fields of the NAO\(^-\) (NAO\(^+\)) pattern shown in Fig. 6b (Fig. 7b), we can draw a schematic diagram (Fig. 9) of the occurrence region of blocking events associated with the life cycles of NAO\(^-\) (NAO\(^+\)) events from the Atlantic basin to the European continent. The observed height anomaly field often exhibits a quadrupole structure in the Euro-Atlantic sector if an EB event can follow the NAO\(^+\), while an opposite quadrupole structure is seen for the NAO negative phase (not shown). For the positive NAO phase, there are positive streamfunction anomalies in the southern North Atlantic and in northern Europe during the life cycle of the NAO\(^+\) event (Fig. 7b). The positive streamfunction anomalies essentially correspond to the regions of active blocking events. Thus, it is likely that the high blocking frequency associated with the NAO\(^+\) life cycle is located in the SW–NE direction of the rectangle domain in our schematic diagram (Figs. 9a,c). Because the observed NAO\(^+\) anomaly is shifted eastward relative to its initial position during its life cycle, the occurrence region of the high blocking frequency is inevitably shifted eastward and distributed along the SW–NE direction from the eastern North Atlantic and western Europe to northeastern Europe and the Ural
region, although the blocking frequency in northeastern Europe and the Ural region is relatively low. This provides an explanation for why the blocking frequency associated with NAO$^+$ events is distributed along the SW–NE direction, as seen from the 500-hPa blocking frequency distribution field (Fig. 1c).

It is also clear that for the NAO$^-$ phase there are positive streamfunction anomalies in the high-latitude North Atlantic and in low-latitude Europe during its life cycle, as these positive anomaly regions in Fig. 6b roughly represent the high-latitude North Atlantic and low-latitude Europe, respectively. This hints that the optimal occurrence regions of blocking are concentrated in these two regions. Thus, the blocking frequency is distributed along the NW–SE direction in Figs. 9b and 9d. Because the observed NAO$^-$ anomaly is retrogressive (i.e., moving westward) along the zonal direction relative to its initial position during its life process, the whole blocking frequency pattern associated with NAO$^-$ events is shifted westward (Fig. 9d).

The overlapping of eastward- and westward-displaced blocking frequency distributions associated with both NAO$^+$ and NAO$^-$ events (Fig. 9e) lead to a maximum blocking frequency in the eastern Atlantic and western Europe, while the blocking frequency is higher over Greenland than over northeastern Europe and the Ural region (Fig. 1a). Thus, a comparison with Fig. 1 indicates that the NMI model can qualitatively capture the geographical distribution of the blocking occurrence frequency in the Euro-Atlantic sector. This suggests that the observed blocking frequency distribution in the Euro-Atlantic sector results primarily from a combination of the different blocking frequency patterns associated with the NAO$^+$ and NAO$^-$ events.

5. Impact of Atlantic conditions on European blocking associated with NAO

a. Temporal relationship between NAO and blocking events in a theoretical model

To understand the causal relationship between the EB activity and the phase of the NAO event, here we construct a local index. The local index can help understand if the NAO events lead or lag EB events. For the local index, in the NMI model we define the difference of the streamfunction anomaly between $y = 1.33$ (low latitude) and $y = 3.67$ (high latitude) at $x = -1.2$ (as a given point in the North Atlantic) as a local NAO index for the NAO negative phase. Its value at $x = 0$ (as a fixed point in the European continent) is defined as a local blocking index. Similarly, for the positive phase, we define the
The difference of the streamfunction anomaly between $y = 1.33$ and $y = 3.67$ at $x = -0.4$ (as a given point in the North Atlantic) as a local NAO index. Their differences at $x = 2.4$ and $x = 3.6$ (as two given points in the European continent) are defined as two local blocking indices, respectively. The local model blocking index is very similar to the TM index as a degenerate form of the two-dimensional index of Davini et al. (2012a) in Part I.

Figure 10a shows the evolution of daily local NAO and blocking indices of an NAO event for $\Delta \mu = 0$ and $\Delta \mu = -0.1$. It is seen that the blocking event over northern Europe ranging from $y = 2.5$ to $y = 5$ and from
$x = 0$ to $x = 5.74$ leads the NAO$^-$ event by about 1 day. Such a lead is more evident for $\Delta u = -0.1$ than for $\Delta u = 0$ (Fig. 10b). In fact, Sung et al. (2011) also found that NAO$^-$ events originate from the westward migration of the amplifying blocking ridge over northern Europe. Figures 10c and 10d show that the decay of the NAO$^+$ event leads to a blocking event over Europe after a long time interval (18 and 21 days for $\Delta u = 0$ and $\Delta u = 0.1$). This can also be seen from the streamfunction field of the NAO$^+$ events shown in Fig. 11c for day 21 (see also Fig. 11a). Thus, the EB occurs about 20 days after the NAO$^+$ peaks. Although the time interval looks different from that of realistic results in Fig. 11 of Part I, the results are qualitatively consistent between the observation and theoretical study. The main reason is that many physical factors, such as baroclinicity, are neglected in our NMI model.

Although the above results show that the NAO$^+$ (NAO$^-$) events lead (lag) the blocking events over Europe (Figs. 9 and 10 in Part I), the strength and longitudinal position of blocking events are modulated by the Atlantic background conditions, as noted in Part I, which are examined below.

b. Impact of mean zonal wind on European blocking

Although the local NAO and blocking indices presented above can identify the lead–lag relationship between the NAO and blocking events, they are unable to track the position and strength of a European blocking event at different times. Here, we introduce a position-tracking index, defined as the difference $B_T = \psi_{\text{BP}} - \psi_{\text{BN}}$ between the strongest negative ($\psi_{\text{BN}}$) and positive ($\psi_{\text{BP}}$) anomalies in the European region $0 \leq x \leq 5.74$ (1 $\leq x \leq 6.74$) for the negative (positive) phases of NAO. The position-tracking index can help identify the relationship between the NAO and EB events in their intensity, location, and duration. At a given time, the zonal position of the center of the blocking with the largest amplitude can be recorded as the blocking position. The basic idea of this model blocking index is similar to the definition of Diao et al.’s (2006) two-dimensional blocking index in Part I. Correspondingly, the value of $B_T$ may represent the blocking intensity at a given time. For a given critical value $B_{IC}$, the time interval from a value greater than $B_{IC}$ to a value less than $B_{IC}$ may be defined as the duration of a blocking event.

Because the initial NAO$^+$ (NAO$^-$) anomaly is assumed to be located at $x = 0$, we define the difference $I_{\text{NAO}} = \psi_{\text{NAO}} - \psi_{\text{NAOP}}$ (an opposite difference) of the dipole structure in the North Atlantic region $-5.74 \leq x \leq 0$ ($-4.74 \leq x \leq 1$) as a daily NAO index for an NAO$^-$ (NAO$^+$) event. Note that $\psi_{\text{NAOP}}$ and $\psi_{\text{NAON}}$ represent the strongest positive and negative anomalies of a dipole pattern in the North Atlantic region.

By calculating $B_T$ and recording the existing region of the maximum of its absolute value $|B_T|$ at each day, the occurrence position of the European blocking event for each day can be determined during its life cycle. Moreover, we define $I_m = (I_{\text{NAO}})_{\text{max}} [B_m = (B_T)_{\text{min}}$ as the maximum NAO index (minimum blocking index) during the life cycle of an NAO (EB) event. Correspondingly, the value and existing region of $B_m$ are defined as the blocking intensity and position, respectively. A similar definition can be made for the NAO index.

As shown in Fig. 6a, the total streamfunction field of the NAO$^-$ event resembles a meandering blocking flow, as first observed by Berggren et al. (1949). Thus, to some degree NAO$^-$ events are identified as corresponding to blocking events over Greenland or the North Atlantic (Woollings et al. 2008; Davini et al. 2012b). The NAO$^+$ event moves westward during its growing stage for $\Delta u = 0$ (Fig. 6a from day 6 to day 15) so that there is an enhanced frequency of blocking events over the North Atlantic, because the positive anomaly is located near $x = -1.0$ (the North Atlantic according to its definition) during the period from day 6 to day 15 (Fig. 6b). The westward migration of the NAO$^-$ anomaly is more evident when the mean zonal wind is weaker (not shown). We also found that the NAO$^-$ pattern is located more eastward for $\Delta u = 0.2$ (not shown) than for $\Delta u = 0$ because of the suppression of its westward migration in the presence of a stronger zonal wind. Below, we will focus on the impact of NAO$^+$ events on the EB activity under different mean zonal wind strengths, because the NAO$^+$ events are followed frequently by EB events, and the decadal variation of EB events from P1 to P2 results primarily from the decadal change in NAO$^+$ events.

The planetary-scale and total streamfunction fields of an NAO$^+$ event for $\Delta u = -0.1$ and $\Delta u = 0.1$ are shown in Figs. 11 and 12, respectively. It is seen that in the Atlantic basin (from $x = -4.74$ to $x = 1.0$) an NAO$^+$ event can arise from the eddy forcing when a weak initial NAO$^+$ anomaly is located at $x = 0$ (Fig. 11a). Along with the decay of the NAO$^+$ event, a blocking high appears over the European continent and eastern Atlantic through the local amplification of a large-scale ridge over Europe (Figs. 11a,b from day 21 to day 45). Such a blocking is the so-called NAO$^+$-induced EB event, whose strength, duration, and zonal position are influenced by the strength of the Atlantic background westerly wind, as observed in Part I. This EB moves westward and enters the Atlantic basin as it intensifies further and finally leads to an NAO$^+$ to NAO$^-$ transition event as the blocking is located over the Atlantic basin (Fig. 11a from day 30 to day 45). This can be seen
Fig. 11. Instantaneous (a) planetary-scale and (b) total streamfunction fields (CI = 0.3) of an NAO$^+$ to NAO$^-$ transition event due to the evolution of an NAO$^+$ event from the NMI model with $a_0 = 0.17$ for $\Delta u = -0.1$. 
FIG. 12. Instantaneous (a) planetary-scale and (b) total streamfunction fields (CI = 0.3) of an NAO+ event from the NMI model for $\Delta u = 0.1$ and other parameters, as in Fig. 11.
from the time evolution of the daily NAO index that varies with the value of $\Delta u$ (Luo and Cha 2012, their Fig. 10d). The snapshots of the planetary-scale and total fields of the NAO$^+$ to NAO$^-$ transition events are also consistent with those from reanalysis data (Luo et al. 2011, their Fig. 9).

The blocking event over Europe is located more eastward as the Atlantic background westerly wind increases, for example from $\Delta u = -0.1$ (Fig. 11a) to $\Delta u = 0.1$ (Figs. 12a,b, from day 39 to day 51). Figure 13 further shows that EB events intensify and are located more eastward as the Atlantic mean zonal wind strengthens during P1, as seen in Part I (Fig. 13a therein). Such a relationship between the blocking position and the mean zonal wind strength can also be detected for initial conditions $B(x, 0) = B_0e^{-x^2}$ for three values of $B_0$ being 0.35, 0.4, and 0.45 and the values of $\nu$ ranging from 0 to 0.06, as given in Table 1 (not shown).

c. Impacts of Atlantic storm track on European blocking

The strength of the Atlantic storm track was found to be able to influence the NAO variability (Lau 1988; Vallis et al. 2004); thus, it can certainly affect the blocking activity over the European continent. For $\Delta u = 0.1$ and $a_0 = 0.21$ (eddy strength, a measure of the storm-track strength in the western Atlantic), the instantaneous planetary-scale streamfunction field of an NAO$^+$ event obtained from the NMI model is shown in Fig. 14. A comparison with Fig. 12 shows that the EB event is intensified and shifted westward when the western Atlantic storm track is enhanced (Figs. 12a and 14a, from day 27 to day 36). Such a westward shift becomes more evident when weaker zonal winds are used (not shown). This is because weaker westerly zonal winds allow the EB event to move westward faster as a result of less suppression by the zonal wind and larger westward displacement for stronger EB events. As a result, the EB events are more likely to be located in western (eastern) Europe as the western Atlantic storm track strengthens (weakens) during P2 (P1), as seen from the reanalysis data in Part I (Fig. 13b therein).

Figure 15 further illustrates how the duration, strength, and position of EB events vary with the intensity of the western Atlantic storm track for $\Delta u = 0$ and $\Delta u = 0.1$. We see that the strength of the EB event increases and its position is shifted westward as the storm-track intensity increases. The westward displacement of the EB event is more evident for $\Delta u = 0$ than for $\Delta u = 0.1$ (Figs. 15a,b). Thus, strengthening of the western Atlantic storm track favors the occurrence of intense EB events in western Europe. The duration of EB events tends to be longer for $\Delta u = 0.1$ and $\Delta u = 0.2$ cases than the $\Delta u = 0$ case (Fig. 15c) when the western Atlantic storm track is relatively weak. However, the lifetime of the EB events is short under weak mean zonal winds when the western Atlantic storm track is extremely weak ($a_0 < 0.16$). This is because few cases under these conditions can develop into a blocking event. Thus, the results here explain the findings presented in Part I, that the EB events are more intense and frequent in EE (WE) during P1 (P2), because P1 has stronger mean zonal winds, while P2 has stronger western Atlantic storm activities. However, it should be noted that the blocking duration from our theoretical model is longer than that seen from the reanalysis data [Diao et al. (2006, their Fig. 13b)]. This may be due to the exclusion of other factors in the present model. For example, the lifetime of blocking events decreases when the basic baroclinicity is included in a nonlinear blocking model (Luo 2005c).

d. Double-peak European blocking events and their possible cause

As shown in Fig. 11a, in the upstream side of the European continent or in WE, the EB event following NAO$^+$ events can behave as a double-peak structure (one peak between days 27 and 33, and another one between days 33 and 45) during its life cycle and become a long-lived event. This can be further seen from the variation of the normalized blocking intensity ($BI = -B_1$) of the EB event, as shown in Fig. 16 (dashed line). However, when the mean zonal wind is relatively strong, the double-peak structure of the EB event almost disappears (Fig. 12a) and it becomes a single peak (Fig. 16, solid line). Thus, it is possible that whether the long-lived EB events that follow NAO$^+$ events can exhibit a double-peak structure depends on the strength of the mean zonal wind over the North Atlantic. When the
Fig. 14. Temporal evolution of (a) planetary-scale and (b) total streamfunction fields (CI = 0.3) of an NAO+ event from the NMI model for $\omega_0 = 0.21$ and $\Delta \mu = 0.1$. 

(a) Planetary-scale streamfunction fields.

(b) Total streamfunction fields.
Atlantic mean zonal wind is relatively strong, no double-peak EB events can be detected; when the zonal wind is weak, long-lasting double-peak EB events may appear. To see if the theoretical result is qualitatively consistent with the real world, again we analyzed the NCEP–NCAR reanalysis data from 1978–2011. Moreover, we use the one-dimensional blocking index of Tibaldi and Molteni (1990; TM) to calculate the blocking intensity of EB events occurring over the domain 0°–90°E during the life cycles of NAO+ events in a winter (NDJFM). The TM index is used to estimate the intensity of a blocking event for a given region, which is approximately the one-dimensional version of the blocking intensity index used in Part I. The statistical analysis shows that for 1978–2011, the total number of NAO+ events is 67, of which 54 (or 81%) correspond to EB events. As shown in Fig. 1b, the frequency of EB events is much less frequent for the case without NAO events. A comparison between Figs. 1c and 1d further shows that the contribution of NAO+ events to the EB frequency seems to be dominant. Thus, the majority of the NAO+ events favor an EB event, supporting the finding of Luo et al. (2007a). Here, we show the composite daily NAO index of all NAO+ events during 1978–2011 in Fig. 15a, in which the time interval A (from lag 20 to lag 10 days) is considered as the earlier stage of an NAO+ event, because its lifetime is 10–20 days, whereas time interval B (from lag 0 to lag +30 days) represents the period of the EB event occurrence due to the NAO+ decay. The calculation of the EB events during period B is reasonable, because NAO+ events often precede EB events by 1 day over southwestern Europe and 3 days over southeastern Europe (Part I). Thus, the mean zonal wind calculated during period A may be considered as the background wind of the NAO+ event. In the statistical analysis of blocking events using the TM index, the maximum value of \( GHGN(\lambda_0, \phi_N) = \frac{Z(\lambda_0, \phi_0) - Z(\lambda_0, \phi_0))}{(\phi_N - \phi_0)} \) on every day may be defined as the daily blocking intensity (http://www.cpc.ncep.noaa.gov/products/precip/CWlink/MJO/block.shtml), where \( \phi_N = 80^\circ N + \Delta, \phi_0 = 60^\circ N + \Delta, \) and \( \Delta = -5^\circ, 0^\circ, \) and \( 5^\circ \). Note that \( Z(\lambda_0, \phi_0) \) is the daily 500-hPa geopotential height at a given grid point with longitude and latitude (\( \lambda_0, \phi_0 \)) (Tibaldi and Molteni 1990).

Using the TM index, the calculation shows that, during the period 1978–2011, there are 34 single-peak and 20 double-peak EB events over Europe that follow NAO+ events for events lasting for at least 3 days. The ratio of single-peak (double peak) events to the total blocking events lasting for at least 3 days is found to change from 69% (31%) to 56% (44%) from P1 to P2.
Thus, the frequency of double-peak or single-peak blocking events can exhibit a decadal variation from P1 to P2. However, there are 25 single-peak and 15 double-peak EB events lasting for at least 5 days, as well as 9 single-peak and 13 double-peak EB events lasting for at least 10 days. This implies that long-lived single-peak blocking events are less frequent than long-lived double-peak events. To clearly see the difference of the blocking intensity between single- and double-peak blocking events, it is necessary to perform the composite of the blocking intensity for two types of EB events.

Here, single-peak and double-peak EB events are picked out during the decaying period (from lag 0 to lag 30 days) of NAO\(^+\) events. Once these EB events are determined, we can perform the composite of the blocking intensity for these events based on the time [lag (0)\(_B\) days] with the largest amplitude of a blocking event during its life cycle [from lag (-10)\(_B\) to lag (+10)\(_B\) days, where lag (+10)\(_B\) denotes the tenth day after the largest blocking amplitude] and the solid (dashed) line represents the single-peak (double-peak) blocking event. The composite blocking intensity is shown in Figs. 17b and 17c for single-peak and double-peak EB events lasting for at least 5 and 10 days. The dark (light) gray shading denotes the regions with positive (negative) values that are above the 95% confidence level based on a two-sided Student’s t test.

As revealed above, whether a single-peak or double-peak EB event can occur is closely related to the strength of the barotropic mean zonal wind in the Atlantic basin (Fig. 16). To understand if this conclusion holds in the real world, it is proper to examine the
difference of the barotropic mean zonal wind (BMW) during the earlier stages of NAO$^+$ events between two types of EB events. Here, we define the zonal wind averaged between 850 and 250 hPa as a barotropic zonal wind. Moreover, the barotropic zonal wind averaged from lag −20 to lag −10 days is also considered as a BMW prior to the NAO$^+$ event, where lag 0 denotes the day with the largest amplitude of each NAO$^+$ event. Figures 17d and 17e show the difference of the composite barotropic zonal wind between single-peak and double-peak blocking events lasting for at least 5 and 10 days. It is obvious that the BMW of the EB event associated with the NAO$^+$ event is different for the two types of blocking events, because its difference is significant. It is interesting to see that the domain-averaged barotropic zonal wind is stronger in the Atlantic basin for single-peak events than for double-peak events. Thus, the reanalysis data show that weak BMW over the North Atlantic favors the occurrence of double-peak EB events associated with NAO$^+$ events, confirming our theoretical findings presented above. As a result, it is suggested that the increased frequency of single-peak (double peak) EB events during P1 (P2) is associated with the stronger (weaker) BMW over the North Atlantic. Because double-peak (single peak) EB events correspond to a longer (shorter) persistence in WE (EE) (Figs. 17a,b), the weaker (stronger) BMW favors the increased frequency of EB events in WE (EE) during P2 (P1).

Lee (1995) showed that the baroclinicity (or vertical wind shear) of the mean state tends to increase with the mean zonal wind strength in the upper troposphere. Thus, compared to the double-peak blocking events, the baroclinicity should be large during the single-peak blocking events, when the upper zonal winds are strong. The large baroclinicity and wind shear shorten the lifetime of the single-peak blocking events, while the relatively weak baroclinicity and wind shear associated with weaker zonal winds could allow the double-peak blocking events to last longer. Therefore, with weak background westerly winds, long-lived, double-peak blocking events can occur over Europe.

6. Conclusions and discussion

In this paper, we first analyzed NCEP–NCAR reanalysis data to depict a T-bone-type distribution of the climatological-mean blocking frequency over the Euro-Atlantic sector. This pattern resembles the shape of an upside-down “T” rotated about 45° to the left (Fig. 1a), and it results primarily from a superimposition of the different blocking frequency patterns associated with NAO$^+$ and NAO$^-$ events (Figs. 1c,d), as the blocking frequency not associated with NAO is low (Fig. 1b). The NAO$^+$-related blocking frequency concentrates in a region covering the eastern North Atlantic and western and central Europe and is distributed along the southwest–northeast (SW–NE) direction, whereas the NAO$^-$-related blocking frequency peaks over a region extending from Greenland to the North Sea along the southeast–northwest (SE–NW) direction and is located more westward. The westward (eastward) displacement of the blocking events associated with the NAO$^-$ (NAO$^+$) events is related to the weakening (strengthening) of the zonal winds over the midlatitude North Atlantic during those events.

We then used a weakly nonlinear multiscale interaction (NMI) model to examine the physical processes leading to the blocking frequency patterns seen in the reanalysis data. It is found that the phase of the NAO event controls the spatial distribution of Euro-Atlantic blocking events in the NMI model, which reproduces the blocking frequency patterns associated with the two NAO phases seen in reanalysis data. Furthermore, in this model the positive (negative) NAO phase corresponds to the concomitant strengthening (weakening) of the zonal wind over the midlatitude North Atlantic. The NAO$^+$ (NAO$^-$)-related blocking frequency pattern along the SW–NE (SE–NW) direction undergoes an eastward (westward) shift, because the strong (weak) zonal wind favors eastward (westward) movement of the EB events as a result of increased (decreased) phase speed because of the enhanced (reduced) zonal winds over the midlatitude North Atlantic. When there are no NAO events, blocking events almost disappear in the Euro-Atlantic sector. As a result, the combination of the NAO$^+$- and NAO$^-$-related blocking frequency distributions can lead to a T-bone-type frequency pattern, as seen in reanalysis data. The results are different from our previous theoretical findings (Luo et al. 2007a,b, 2011; Luo and Cha 2012).

Moreover, the impacts of the mean zonal wind and western Atlantic storm-track strength on the duration, intensity, and position of the EB events resulting from the decay of NAO$^+$ events were also examined using this simplified model. It is shown that when the mean zonal wind (western Atlantic storm track) is relatively strong, the EB event tends to be strong, long-lived, and located in eastern Europe (western Europe), thus providing a theoretical explanation for the decadal changes presented in Part I of this study. Further, it is found that under strong Atlantic mean zonal wind conditions, the duration of the EB events is long and almost independent of the strength of the western Atlantic storm track, which mainly affects the intensity of the EB events when
the mean zonal wind is strong. An interesting result is that the life cycle of the EB events shows double peaks when the Atlantic mean zonal wind is weak but becomes a strong single peak when the mean zonal wind is strong. This is consistent with the reanalysis data (Fig. 17). The strengthening (weakening) of the Atlantic mean zonal wind during P1 (P2) leads to a decadal change in the frequency of single-peaked and double-peaked blocking events associated with NAO$^+$ events over Europe. This result is also a new finding.

On the other hand, we found from our simple model that the NAO$^+$ to NAO$^-$ transition takes place more easily as the Atlantic mean zonal wind (western Atlantic storm track) weakens (strengthens). Such a transition can result in an enhanced EB frequency in western Europe as a result of the marked westward migration of EB events during these conditions. This provides further explanation for why blocking events are frequent in northwestern Europe during 1995–2011 (when the mean zonal winds are weak and the storm activity is strong) relative to 1978–94 (when the mean zonal winds are strong and the storm activity is weak), as noted in Part I.

We recognize that the NMI model used here is a highly simplified barotropic model. Another important assumption is that only a uniform background westerly wind is considered in this weakly nonlinear model, even though the observed Atlantic mean zonal wind exhibits a weak jetlike distribution prior to the NAO onset. However, the magnitude of $\Delta u$ can crudely reflect the strength of the Atlantic jet. Moreover, the change of the phase speed of the NAO anomaly induced by the variation of the zonal wind during the NAO life cycle is not considered in our model because of the model solution limit. As shown in Part I, the strengthening of the mean zonal wind from 1978–94 to 1995–2011 is actually not uniform in the zonal direction, which corresponds to a more marked strengthening over the Atlantic basin than over Europe. To account for this zonal variation, we assumed a variable $\Delta u = \Delta u_0 e^{-\nu x_u + x_u^2}$ (where $\Delta u_0$ is a constant, $x_u$ is the position of the maximum wind, and $0 < \nu \ll 1.0$) but the results (not shown) were similar to those discussed above. For these reasons, the obtained blocking duration is longer than that in reanalysis data, because other factors, such as the baroclinicity of the mean flow, have been excluded in the model. Nevertheless, our model results are qualitatively consistent with the reanalysis data and provide a plausible interpretation of the reanalysis results. Of course, further investigation on the role of other factors in the variation of blocking events in Europe is needed.

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