The Variability of the Atlantic Storm Track and the North Atlantic Oscillation: A Link between Intraseasonal and Interannual Variability

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ABSTRACT

The winter-mean North Atlantic Oscillation (NAO) index has been mostly positive since the 1980s, with a linear upward trend during the period from 1978 to 1990 (P1) and a linear downward trend during the period from 1991 to 2009 (P2). Further calculations show that the Atlantic storm-track eddy activity is more intense during P2 than during P1, which is statistically significant at the 90% confidence level for a t test. This study proposes a hypothesis that the change in the trend of the positive NAO index from P1 to P2 may be associated with the marked intensification of the Atlantic storm track during P2.

A generalized nonlinear NAO model is used to explain the observed trend of the positive NAO index within P2. It is found that even when the Atlantic storm-track eddies are less intense, a positive-phase NAO event can form under the eddy forcing if the planetary-scale wave has an initial value with a low-over-high dipole structure during P1 and P2. A blocking flow can occur in the downstream side (over Europe) of the Atlantic basin as a result of the energy dispersion of Rossby waves during the decay of the positive-phase NAO event. This blocking flow does not strictly correspond to a negative-phase NAO event because the blocking stays mainly over the European continent. However, when the Atlantic storm-track eddies are rather strong, the blocking flow occurring over the European continent is enhanced and can retrograde into the Atlantic region and finally become a long-lived negative-phase NAO event. In this case, the NAO event can transit from the positive phase to the negative phase. Thus, the winter-mean NAO index during P2 will inevitably decline because of the increase in days of negative-phase NAO events in winter because the Atlantic storm track exhibits a marked intensification in the time interval. The transition of the NAO event from the positive phase to the negative phase can also be observed only when the downstream development of the Atlantic storm-track eddy activity is rather prominent.

Thus, it appears that there is a physical link between intraseasonal and interannual time scales of the NAO when the Atlantic storm track exhibits an interannual variability.

1. Introduction

The North Atlantic Oscillation (NAO) is the most prominent and recurrent low-frequency dipole pattern of atmospheric variability in the mid-to-high latitudes of the Northern Hemisphere (NH). The NAO is often referred to as a seesaw of the atmospheric sea level pressure between the Arctic and subtropical Atlantic (Hurrell 1995; Hurrell et al. 2003). This phenomenon has been an important research subject because a positive trend of the NAO may be related to warming over the Northern Hemisphere landmasses (Hurrell 1995; Cohen and Barlow 2005).

The NAO exhibits a trend in its index from large-amplitude anomalies of the negative phase in the 1960s.
to large-amplitude anomalies of the positive phase since the early 1980s (Feldstein 2002; Cohen and Barlow 2005). Many studies have suggested that the positive trend of the NAO index is related to increased greenhouse gas concentrations (Palmer 1993; Shindell et al. 1999; Ulbrich and Christoph 1999) and other external forcings (Rodwell et al. 1999; Hoerling et al. 2001; Gillett et al. 2003). Hall et al. (1994) noted with a high-resolution general circulation model (GCM) that the storm-track eddy activity exhibits downstream intensification under doubled carbon dioxide (2\textsuperscript{3}CO\textsubscript{2}). Ulbrich and Christoph (1999) found in an atmosphere–ocean coupled model that under anthropogenic forcing the positive trend of the NAO index is associated with an increase in the storm-track eddy activity in the northeastern Atlantic because of the increased baroclinicity in that region.

To clearly see the long-term variability of the NAO index, the winter [December–February (DJF)] mean NAO index during 1950–2009 is shown in Fig. 1, which is constructed by averaging the daily DJF NAO index at 500 hPa, available from the Climate Prediction Center (CPC; http://www.cpc.noaa.gov/). It can be seen that the winter-mean NAO index is mostly positive during the period from 1978 to 2009, in which the positive NAO index exhibits an upward trend during 1978–90 (P1) and a downward trend during 1991–2009 (P2). In particular, for the 2009/10 winter, the winter-mean NAO index is extremely negative (Wang et al. 2010). Cohen and Barlow (2005) speculated that such a downward trend of the NAO index may be related to an increase in the October Eurasian snow cover. However, the dynamical mechanisms underlying the strong reversal in the NAO trend remain unclear. Recent theoretical studies sufficiently indicate that the interaction of storm-track eddies, a planetary wave, and the mean flow can excite NAO events (Luo et al. 2007a,b). Thus, it is inferred that the variability of the Atlantic storm track may play an important role in the variability of the NAO trend. It has been recognized that the variability of the North Atlantic storm-track eddy activity during 1978–2009 (Fig. 2) may be associated with an increased baroclinicity due to growing greenhouse gas concentrations (Hall et al. 1994; Ulbrich and Christoph 1999; Paeth et al. 1999). In some sense, however, the mechanism by which the increased greenhouse gas concentration affects the NAO trend may also depend on the variability of the Atlantic storm track. As this study will suggest, the Atlantic storm-track activity can exert a positive feedback on the upward trend of the NAO index if the Atlantic storm-track intensity is not too strong, but a negative feedback if it is sufficiently strong.

In this study, we propose a hypothesis that the strong downward trend of the NAO index within P2 may be related to the intensification of the storm track over the North Atlantic, which induces a change in the sign of the NAO index (on intraseasonal time scales). In this sense, the interannual trend of the NAO index can be explained in terms of the intraseasonal variability of the NAO. Furthermore, some observational evidence is provided to test our hypothesis. In addition, the highly idealized weakly nonlinear NAO model of Luo et al. (2007b) is extended to examine if the downstream development of the Atlantic storm track can affect such a transition.

The paper is outlined as follows: in section 2, we describe a possible link between the NAO trend and the variability of the Atlantic storm track. In section 3, the weakly nonlinear NAO model of Luo et al. (2007b) is extended to include the downstream development of the Atlantic storm-track eddy activity. Then, in sections 4
and 5, we examine the relationship between the Atlantic storm-track variability and the phase of the NAO. It is shown that a marked enhancement of the Atlantic storm-track activity can lead to a westward drift of European blocks followed by an increase in the number of days of negative-phase NAO events. In particular, when the downstream development of Atlantic storm-track eddies is rather strong, the transition from positive- to negative-phase NAO events can be seen. In section 6, observational evidence is provided to show that there is a large increase in days of negative NAO events during P2, which is based on the assumption that negative-phase NAO events and Atlantic blocking events are the same phenomenon. Moreover, the statistical analysis performed shows that most of the relatively long-lived negative-phase NAO events during P2 arise from the westward drift of the amplified European blocks. The conclusions and discussion are presented in section 7.

Moreover, it is noted that the difference of the EKE between P2 and P1 is positive over part of the European continent (0°–30°W) and negative over the eastern Atlantic near 10°E, although it is not statistically significant. The positive EKE anomalies over Europe can crudely be seen as being induced by the downstream development associated with the Atlantic storm track. This downstream development essentially corresponds to the group velocity being greater than the phase velocity (Chang 1993). It is thus anticipated that the Atlantic storm track can undergo a stronger downstream development during P2 than during P1. As a result, we conjecture that the reversal in the NAO trend from P1 to P2 is, to a large extent, due to the more marked enhancement of the EKE strength in the northwestern Atlantic during P2 than during P1, while the downstream development of the Atlantic storm track may affect the NAO variability.

2. A possible link between the NAO trend and the Atlantic storm track

To search for a possible relationship between the NAO trend and Atlantic storm-track variability, the winter-mean high-pass (2–7 days) eddy kinetic energy (EKE) at 250 hPa is calculated using National Centers for Environmental Prediction (NCEP)–National Center for Atmospheric Research (NCAR) reanalysis data for the period from January 1958 through December 2009. Because our attention is focused on understanding why the winter-mean NAO index undergoes a change in its trend from P1 to P2, it is necessary to understand the low-frequency variability of the Atlantic storm track. Here, the maximum of the (DJF) time mean standardized EKE in the domain 25°–80°N, 90°W–20°E is defined as the maximum storm-track intensity in the winter for the Atlantic region. Correspondingly, the time mean standardized EKE averaged over this domain is defined as the mean Atlantic storm-track intensity in winter. Figure 2 shows the time series of the maximum and mean winter Atlantic storm-track intensities in the North Atlantic (25°–80°N, 90°W–20°E). It can be seen that the maximum and mean Atlantic storm-track intensities exhibit a monotonic upward trend during 1978–2009, with a correlation coefficient of 0.75. Figure 3 shows the spatial distributions of the winter-mean EKE at 250 hPa over the North Atlantic during P1 and P2, and their difference. It is found that the difference of the storm-track intensity between P2 and P1 upstream of 30°W in the North Atlantic is statistically significant above the 90% confidence level for a two-sided Student’s t test. Thus, the variability of NAO and blocking events are associated with a change in the Atlantic storm track upstream.

3. Extended weakly nonlinear NAO model

To investigate the dynamical processes that govern the reversal of the NAO trend during P2, we will use a highly simplified weakly nonlinear NAO model parallel to that of Luo et al. (2007b), as this model was designed to focus on the role of the Atlantic storm-track variability in the development of the NAO. At the same time, this highly simplified NAO model is also extended to include the downstream development of the Atlantic storm-track eddy activity over the North Atlantic. The basic dynamical framework of this model is also similar to that used in Luo et al. (2007b).

To examine the above problems, in our simplified model the variability of the Atlantic storm track is assumed to be a superimposition of two parts: one is the variation of the Atlantic storm track in both intensity and position, and the other is the downstream development contributing to the storm-track variability. The downstream development of the Atlantic storm track was not considered in our previous models. In this paper, such an effect will be included to examine if the downstream development can contribute significantly to the NAO variability. In addition, in the present paper, a uniform background westerly wind is assumed to emphasize how the Atlantic storm-track variability affects the NAO evolution. The effect of different basic flows on NAO variability has been examined in Luo et al. (2008).

As in Luo et al. (2007a,b), in a uniform background flow $u_0$ the nondimensional perturbation barotropic vorticity equations of the interaction between planetary- ($\psi$) and synoptic-scale ($\psi'$) waves scaled with the characteristic length $L$ and velocity $U$ can be expressed as
(\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 \psi}{\partial x} + u_0 \frac{\partial h}{\partial x} = -J(\psi', \nabla^2 \psi')_p, \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' + \psi^p), \quad (1b)

where \(u_0\) is the constant westerly wind; \(\beta\) is the non-dimensional meridional gradient of the Coriolis parameter centered at a given latitude \(\phi_0\); \(F = (L/R_d)^2\), in which \(R_d\) is the radius of Rossby deformation; \(h\) is the non-dimensional topographic height; the subscript \(P\) in (1a) denotes the planetary-scale projection; and \(\psi^p\) indicates a synoptic-scale vorticity source designed to maintain synoptic-scale waves prior to the NAO onset. The parameter values of the characteristic scales, the notation, and boundary conditions used can found in Luo (2005) and Luo et al. (2007a,b).

It should be noted that the term \(-J(\psi', \nabla^2 \psi')_p\) in (1a) is a planetary-scale component of the eddy vorticity fluxes that drive the NAO pattern growth and decay (Luo et al. 2007a,b).

As in Luo (2005) and Luo et al. (2007a,b), we can assume \(\psi = \varepsilon \psi, \psi' = \varepsilon^{3/2} \hat{\psi}', h = \varepsilon h',\) and \(\psi^p = \varepsilon^{5/2} \hat{\psi}^p\) in order to obtain analytical solutions of the eddy-driven NAO life cycle for planetary and synoptic scales. In this case, (1a) and (1b) can be reduced to

\[
\frac{\partial}{\partial t} \frac{\partial \hat{\psi}'}{\partial x} \left( \nabla^2 \hat{\psi}' - F \hat{\psi}' \right) + \varepsilon J(\hat{\psi}', \nabla^2 \hat{\psi}' + \varepsilon \hat{\psi}^p) \\
+ (\beta + Fu_0) \frac{\partial \hat{\psi}'}{\partial x} + u_0 \frac{\partial \hat{\psi}'}{\partial x} = -\varepsilon^{2} J(\hat{\psi}', \nabla^2 \hat{\psi}')_p, \quad (2a)
\]

where \(\hat{\psi}\) and \(\hat{\psi}'\) denote the planetary and synoptic scales, respectively.

To derive the asymptotic solutions of (2), we introduce the slowly varying time and space coordinates

\[T_1 = \varepsilon t, \quad T_2 = \varepsilon^2 t, \quad X_1 = \varepsilon x, \quad X_2 = \varepsilon^2 x, \quad (3)\]

where \(\varepsilon\) is a small parameter required to satisfy \(0 \leq \varepsilon \ll 1.0\).

We can expand \(\hat{\psi}\) and \(\hat{\psi}'\) as

\[\hat{\psi} = \hat{\psi}_0(x, y, t, T_1, T_2, X_1, X_2) \\
+ \varepsilon \hat{\psi}_1(x, y, t, T_1, T_2, X_1, X_2) \\
+ \varepsilon^2 \hat{\psi}_2(x, y, t, T_1, T_2, X_1, X_2) + \cdots, \quad (4a)\]

\[\hat{\psi}' = \hat{\psi}'_0(x, y, t, T_1, X_1) \\
+ \varepsilon \hat{\psi}'_1(x, y, t, T_1, T_2, X_1, X_2) + \cdots. \quad (4b)\]

Substituting (3) and (4) into (2) with \(\hat{\psi}^p = \psi^p\) yields...
\[ O(e^0)N(\psi'_0) = \left( \frac{\partial}{\partial t} + u_0 \frac{\partial}{\partial x} \right) \left[ \nabla^2 \psi'_0 - F(\psi'_0) \right] + (\beta + Fu_0) \frac{\partial(\psi'_0)}{\partial x} = 0, \] (5a)

\[ O(e^1)N(\psi'_1) = \left( \frac{\partial}{\partial t} + u_0 \frac{\partial}{\partial x} \right) \left[ \nabla^2 \psi'_1 - F(\psi'_1) \right] + (\beta + Fu_0) \frac{\partial(\psi'_1)}{\partial x} = \]

\[ = - \left[ \frac{\partial}{\partial t} + u_0 \frac{\partial}{\partial x} \right] \left( \nabla^2 \psi'_0 - F(\psi'_0) \right) + 2 \left( \frac{\partial}{\partial t} + u_0 \frac{\partial}{\partial x} \right) \frac{\partial^2 \psi'_0}{\partial x \partial X_1} + (\beta + Fu_0) \frac{\partial^2 \psi'_0}{\partial x \partial X_1} \]

\[ + [J(\psi'_0, \nabla^2 \psi'_0 + h') + J(\psi'_0, \nabla^2 \psi'_0)] \right] + \nabla^2 \psi'_1, \] (5b)

where the term SV denotes the slow variation of synoptic-scale eddies, the term PSI represents the planetary-to-synoptic interaction, and the term SEVC denotes the synoptic-scale eddy vorticity source, which is a simple barotropic representation of the generation of eddies via baroclinic instability.

The synoptic-scale eddies to first order are referred to as “preexisting synoptic-scale eddies” and can be described by a linear Rossby wave equation, whose amplitude is slowly varying in time and space. In contrast, the second-order solution of the synoptic-scale eddies is a forced equation because of the presence of the terms SV, PSI, and SEVC. Thus, it is possible to obtain the solution to (5b) if the term SV can be assumed to be balanced by the term SEVC. This assumption can be made because the preexisting synoptic-scale eddies are maintained by the eddy source term SEVC.

Under the balance between terms SV and SEVC, the following linear equation can be obtained:

\[ \left( \frac{\partial}{\partial t} + u_0 \frac{\partial}{\partial x} \right) \left[ \nabla^2 \psi'_0 - F(\psi'_0) \right] + 2 \left( \frac{\partial}{\partial t} + u_0 \frac{\partial}{\partial x} \right) \frac{\partial^2 \psi'_0}{\partial x \partial X_1} \]

\[ + (\beta + Fu_0) \frac{\partial^2 \psi'_0}{\partial x \partial X_1} = \nabla^2 \psi'_1, \] (6)

It is straightforward to solve (6) if we can divide \( \psi'_0 \) into \( \psi'_{01}(x, y, t, \chi) \) and \( \psi'_{02}(x, y, t, T_1, \chi) \). Here, \( \psi'_{01}(x, y, t, \chi) \) is maintained by the pre-specified synoptic-scale eddy vorticity source term, which can reflect the variation of the Atlantic storm track in strength and position. But \( \psi'_{02}(x, y, t, T_1, \chi) \) can describe the downstream development of the Atlantic storm-track eddy activity, as noted later.

In this case, one can get the \( \psi'_{01}(x, y, t, \chi) \) and \( \psi'_{02}(x, y, t, T_1, \chi) \) equations from (6). These take the form of

\[ \dot{\psi}'_{01} = f'_0(X_1) \{ \exp[i(k_x x - \omega_1 t)] + \alpha \exp[i(k_x x - \bar{\omega}_1 t)] \} \sin(my/2) + cc, \] (8)

where \( f'_0(X_1) \) is the amplitude distribution of the pre-existing synoptic-scale eddies; \( k_x = \bar{k}_x + (-1)^i \Delta k \) for \( i = 1, 2 \), where \( k_0 = 1/[6.371 \cos(\phi_0)] \) is wavenumber 1 along the zonal direction at the reference latitude \( \phi_0 \). \( \Delta k \) is the wavenumber difference between \( k_x \) and \( k_1 \), and \( n \) is an integer; \( \bar{\omega}_1 = u_0 \bar{k}_x - (\beta + Fu_0) \bar{k}_x/|\bar{k}_x|^2 + m^2/4 + F \); \( \alpha = \pm 1; m = \pm 2 \pi/L_{*}; \) and \( cc \) denotes the complex conjugate of its preceding term.

Although the observed downstream development of the storm track is a relatively complicated process (Lee
and Held 1993), it can be crudely described by the difference between the group velocity and phase velocity of each eddy in a storm track denoted by \( \psi_0(x, y, t, T_1, X_1) \). Because the group velocity of each eddy is the propagation speed of its envelope amplitude, the envelope amplitude of synoptic-scale waves may be assumed to be slowly varying in time and in the zonal direction. Thus, we seek the solution to (7b) of the form

\[
\hat{\psi}_0 = \{ \hat{f}_m (T_1, X_1) \exp[i(\tilde{k}_m x - \omega_m t)] + \hat{f}_m (T_1, X_1) \times \exp[i(\tilde{k}_m x - \omega_m t)] \} \sin(my/2) + cc, \tag{9}
\]

where \( \tilde{k}_m = \bar{n}_m k_0 + (-1)^i \Delta k \), \( \bar{n}_m \) is an integer, \( \omega_m = u_0 \bar{k}_m - (\beta + F \bar{u}_0) k_m / (k_m^2 + m^2/4 + F) \), and \( \hat{f}_m (T_1, X_1) \) is the amplitude of each component of \( \hat{\psi}_0' (x, y, t, T_1, X_1) \).

Substitution of (9) into (7b) yields

\[
\frac{\partial \hat{f}_m}{\partial T_1} + \tilde{C}_{gmi} \frac{\partial \hat{f}_m}{\partial X_1} = 0 \quad \text{for } i = 1, 2, \tag{10}
\]

where \( \tilde{C}_{gmi} = \partial \omega_m / \partial \tilde{k}_m \) is the group velocity of each wave component in \( \hat{\psi}_0' (x, y, t, T_1, X_1) \).

The solution to (10) is easily obtained as

\[
\hat{f}_m (T_1, X_1) = \hat{f}_m (X_1 - \tilde{C}_{gmi} T_1). \tag{11}
\]

It is evident that the envelope amplitude of each wave component in \( \hat{\psi}_0' (x, y, t, T_1, X_1) \) propagates eastward with its group velocity. Because \( \tilde{C}_{gmi} > \tilde{C}_{pmi} = \tilde{\omega}_m / \tilde{k}_m \) in which \( \tilde{C}_{pmi} \) is the phase speed, the downstream development of synoptic-scale eddies is inevitable because of their downstream energy dispersion (Lee and Held 1993; Chang and Orlanski 1993). Thus, the downstream development of the Atlantic storm-track eddy activity can be crudely represented by the term \( \hat{\psi}_0' (x, y, t, T_1, X_1) \).

We consider two cases: one corresponds to \( \tilde{k}_m > \tilde{k}_i \) and the other to \( \tilde{k}_m > \tilde{k}_j \). This excludes the interaction between \( \tilde{k}_m \) and \( \tilde{k}_j \) that allows the NAO anomaly to be reinforced.

As noted in previous studies (Luo 2005; Luo et al. 2007b), the large-scale topography in the Northern Hemisphere can be approximated as a two-wave topography (Charney and DeVore 1979). Thus, the large-scale topography, denoted by \( h' \), can be assumed to be

\[
h' = h_0 \exp[ik(x + x_T)] \sin(my/2) + cc, \tag{12}
\]

where \( h_0 \) is the amplitude of the two-wave topography with \( k = 2k_0 \), and \( x_T \) is the zonal position of the topographic trough relative to the center of the NAO. It should be noted that \( h_0^2 > 0 \) and \( m = -2 \pi / L_y \) represent the topographic trough in the North Atlantic sector for the negative-phase NAO. The variables \( h_0^2 < 0 \) and \( m = 2 \pi / L_y \) correspond to the case of the positive-phase NAO. Such choices can allow the region \(-2.87 \leq x < 2.87 \) for \( h' \) to represent the Atlantic basin. In this paper we assume \( u_0 = \beta(k^2 + m^2) \) (\( \omega = 0 \)) so that the quasi-stationary dipole wave considered can be resonantly forced by the Atlantic storm-track eddy activity.

Following Luo et al. (2007b), the planetary-scale streamfunction solution \( \psi_P \) in a fast-variable form during a NAO life cycle can be obtained as

\[
\psi_P = -u_0 y + \bar{\theta} \psi_0 + \epsilon^2 \psi_1 = \psi_n + \psi_m, \tag{13a}
\]

\[
\psi_n = -u_0 y + \psi_A + \psi_C, \tag{13b}
\]

\[
\psi_A = B \sqrt{\frac{2}{L_y}} \exp(ikx) \sin(my) + cc, \tag{13c}
\]

\[
\psi_C = h_A h_0 \exp(ik(x + x_T)) \sin(my/2) + cc, \tag{13d}
\]

\[
\psi_m = \psi_{m1} + \psi_{m2}, \tag{13e}
\]

\[
\psi_{m1} = -B^2 \sum_{n=1}^{\infty} q_n g_n \cos(n + 1/2)my, \tag{13f}
\]

\[
\psi_{m2} = -h_0 h_A \sqrt{\frac{2}{L_y}} (Be^{-ikx} + B \epsilon e^{ikx}) \times \sum_{n=1}^{\infty} \bar{q}_n (3a_n - b_n) \cos(nmy), \tag{13g}
\]

where \( B(t, x) = eA(et, e^2t, ex, e^2x), h_0 = \epsilon h_0, h_A = -1/[\beta + u_0 - (k^2 + m^2/4)] \), and the other coefficients and notation can be found in Luo et al. (2007b). It should be noted that these solutions are completely identical to those derived by Luo et al. (2007b), but the equation that describes the amplitude evolution of the NAO anomaly \( B(t, x) \) is different.

In Luo et al. (2007b), \( \psi_A \) is quasi-stationary and is referred to as an NAO anomaly, which is driven by synoptic-scale eddies in the Atlantic storm track. However, \( \psi_C \) is called a “climatological stationary wave” because it is driven by the large-scale topography. Here, \( m = -2 \pi / L_y (m = 2 \pi / L_y) \) represents the negative (positive) phase of the NAO event.

At the same time, the synoptic-scale \( (\psi') \) streamfunction solutions in fast variable form can be expressed as

\[
\psi' = e^{3/2} (\psi'_0 + \epsilon \psi'_1) = \psi'_1 + \psi'_2 = \psi'_{11} + \psi'_{12} + \psi'_{22}, \tag{14a}
\]
\( \psi' = \frac{3}{2} \tilde{\psi}'_{\omega} = f_0(x) \{ \exp[i(\kappa_k x - \omega_m t)] + \alpha \exp[i(\tilde{\kappa}_k x - \tilde{\omega}_m t)] \} \sin\left( \frac{m}{2} y \right) + \text{cc}, \) (14b)

\( \psi'_{\omega} = \frac{3}{2} \tilde{\psi}'_{\omega} = \{ f_{m1}(x - \tilde{C}_{gm1}) \exp[i(\tilde{\kappa}_{m1} x - \tilde{\omega}_{m1} t)] + f_{m2}(x - \tilde{C}_{gm2}) \exp[i(\tilde{\kappa}_{m2} x - \tilde{\omega}_{m2} t)] \} \sin\left( \frac{m}{2} y \right) + \text{cc}, \) (14c)

\( \psi'_e = \frac{m}{4} f_0 h_0 \sigma_1 \exp[i(\tilde{\kappa}_k x + k x_T - \tilde{\omega}_m t)] \sin\left( \frac{m}{2} y \right) = \frac{m}{4} f_0 h_0 \sigma_2 \exp[i(\tilde{\kappa}_k x + k x_T - \tilde{\omega}_m t)] \sin\left( \frac{m}{2} y \right) \times \exp[i(\tilde{\kappa}_k x - k x_T - \tilde{\omega}_m t)] \sin\left( \frac{m}{2} y \right) + \text{cc}, \) (14d)

\( \psi'_e = \frac{m}{4} f_0 h_0 \sigma_1 \exp[i(\tilde{\kappa}_k x + k x_T - \tilde{\omega}_m t)] \sin\left( \frac{m}{2} y \right) = \frac{m}{4} f_0 h_0 \sigma_2 \exp[i(\tilde{\kappa}_k x + k x_T - \tilde{\omega}_m t)] \sin\left( \frac{m}{2} y \right) \times \exp[i(\tilde{\kappa}_k x - k x_T - \tilde{\omega}_m t)] \sin\left( \frac{m}{2} y \right) + \text{cc}, \) (14e)

\( \psi'_e = \frac{m}{4} f_0 h_0 \sigma_1 \exp[i(\tilde{\kappa}_k x + k x_T - \tilde{\omega}_m t)] \sin\left( \frac{m}{2} y \right) = \frac{m}{4} f_0 h_0 \sigma_2 \exp[i(\tilde{\kappa}_k x + k x_T - \tilde{\omega}_m t)] \sin\left( \frac{m}{2} y \right) \times \exp[i(\tilde{\kappa}_k x - k x_T - \tilde{\omega}_m t)] \sin\left( \frac{m}{2} y \right) + \text{cc}, \) (14f)

\( \psi'_e = \frac{m}{4} f_0 h_0 \sigma_1 \exp[i(\tilde{\kappa}_k x + k x_T - \tilde{\omega}_m t)] \sin\left( \frac{m}{2} y \right) = \frac{m}{4} f_0 h_0 \sigma_2 \exp[i(\tilde{\kappa}_k x + k x_T - \tilde{\omega}_m t)] \sin\left( \frac{m}{2} y \right) \times \exp[i(\tilde{\kappa}_k x - k x_T - \tilde{\omega}_m t)] \sin\left( \frac{m}{2} y \right) + \text{cc}, \) (14g)

\( \psi'_e = \frac{m}{4} f_0 h_0 \sigma_1 \exp[i(\tilde{\kappa}_k x + k x_T - \tilde{\omega}_m t)] \sin\left( \frac{m}{2} y \right) = \frac{m}{4} f_0 h_0 \sigma_2 \exp[i(\tilde{\kappa}_k x + k x_T - \tilde{\omega}_m t)] \sin\left( \frac{m}{2} y \right) \times \exp[i(\tilde{\kappa}_k x - k x_T - \tilde{\omega}_m t)] \sin\left( \frac{m}{2} y \right) + \text{cc}, \) (14h)

where \( f_0(x) = e^{\lambda x} f_0(X) \) and \( f_{m1}(x - \tilde{C}_{gm1}) = e^{\lambda x} f_{m1} \) \((X - \tilde{C}_{gm1} T)\); \( \psi'_1 \) denotes the preexisting Atlantic storm-track eddies with amplitude distribution of \( f_0(x) = a_0 \exp[\mu \epsilon(x + x_0)^2] \), where \( a_0 \) is the amplitude of the upstream eddies, \( x_0 \) is the position of the maximum eddy intensity, and the other coefficients and notations can be found in Luo et al. (2007b). Note that \( \alpha = -1 \) (\( \alpha = 1 \)) in (14b) is required for the negative (positive) phase NAO to allow the eddy forcing \( J(\psi'_{\omega}, \nabla^2 \psi'_{1}) \) to excite a negative (positive) phase NAO event. The forms of \( Q_{mi}, p_{mi}, r_{mi}, s_{mi}, h_{mi}, \sigma_{mi}, \) and \( \sigma_{mi} \) with \( i = 1, 2 \) in (14g) and (14h) are the same as \( Q, p, r, s, h, \sigma, \) and \( \sigma \) with \( i = 1, 2 \) in (14e) and (14f) when \( \tilde{k} \) and \( \tilde{\omega} \) in (14e) and (14f) are replaced by \( \tilde{k}_m \) and \( \tilde{\omega}_m \).

It should be pointed out that (14b) represents the storm-track eddy activity maintained by the eddy source.
Similar to Luo et al. (2007a,b), it is straightforward to obtain the evolution equation of the amplitude $B$ of the NAO anomaly forced by both synoptic-scale eddies and the large-scale topography.

If we assume $k_{mi} = k_i$ with $i = 1, 2$, then the forced nonlinear Schrödinger equation that governs the evolution of the NAO anomaly in a fast variable form can be obtained as

$$
i \left( \frac{\partial B}{\partial t} + C_x \frac{\partial B}{\partial x} \right) + \lambda \frac{\partial^2 B}{\partial x^2} + \delta |B|^2 B + \alpha \hbar_0(t) B + G_{f_0} \exp[-i(\Delta k x + \Delta \omega t)] + C_{em1} f_{m1} f_{m2} \exp[-i(\Delta k x + \Delta \omega t)] = 0,$$

(15)

where $C_x = u_0 - (\beta + F u_0)(m^2 + F - k^2)/(k^2 + m^2 + F)^2$, $C_{em1} = \alpha/\sqrt{L/2}[(k_1 + k_2)^2(k_1 - k_2)/[4(k^2 + m^2 + F)]$, and $C_{em2} = C_e = \sqrt{L/2}[(k_1 + k_2)^2(k_1 - k_2)/[4(k^2 + m^2 + F)]$; the other notations and coefficients can be found in Luo et al. (2007b).

For $k_{mi} \neq k_i$ and $k_{m1} \neq k_i$ the B equation of the NAO anomaly forced by both synoptic-scale eddies and the large-scale topography differs from (15) and can be described by the following nonlinear Schrödinger (NLS) equation:

$$
i \left( \frac{\partial B}{\partial t} + C_x \frac{\partial B}{\partial x} \right) + \lambda \frac{\partial^2 B}{\partial x^2} + \delta |B|^2 B + \alpha \hbar_0(t) B + G_{f_0} \exp[-i(\Delta k x + \Delta \omega t)] + C_e f_{m1} f_{m2} \exp[-i(\Delta k x + \Delta \omega t) + \Delta \omega (t) = 0,$$

(16)

where $\Delta k = k - (k_2 - k_1), \Delta \omega = \alpha - \alpha_1 - \omega, \Delta \omega_m = k - (k_{m2} - k_{m1}), \Delta \omega_{m2} = \omega_{m2} - \omega_{m1} - \omega$, and $C_e = \sqrt{L/\lambda_2} [(k_{m1} + k_{m2})(k_{m2} - k_{m1})/\lambda_id_m]$, Other coefficients can be found in Luo et al. (2007b).

Equations (15) and (16) are a generalization of the NAO evolution equation of Luo et al. (2007b). When either $f_{m1}$ or $f_{m2}$ vanishes, (15) and (16) reduce to the amplitude equation of the NAO anomaly derived in Luo et al. (2007b).

Here, $f_{m1}(x - \tilde{C}_{gmi}) = f_{01} e^{-\mu_i(x - \tilde{C}_{gmi}t_x)}$ and $f_{m2}(x - \tilde{C}_{gmi}) = f_{02} e^{-\mu_i(x - \tilde{C}_{gmi}t_x)}$ can be assumed because $f_{m1}$ and $f_{m2}$ are the envelope amplitudes of two components of the synoptic-scale eddies, denoted by $\psi_{g_b}(x, y, t, \tau_1, X_1)$, where $f_{01}$ and $f_{02}$ are the strength of downstream developing eddies and two constants, $\mu_i > 0$ with $i = 1, 2$, and $x_1$ is the initial position of the downstream developing eddy.

In this study, the Atlantic storm track, organized by $\psi_{11}$ in (14b), is defined to be a storm track with no downstream development (referred to as an NDD storm track) because the amplitude of $\psi_{11}$ is fixed at a specified position in the $x$ direction and is time independent. However, the Atlantic storm track organized jointly by $\psi_{11}$ and $\psi_{12}$ can be crudely defined as a “downstream developed storm track” (or DD storm track) because $\psi_{12}$ can reflect the downstream development corresponding to the group speed being larger than the phase speed (i.e., $C_{gmi} > C_{pmi}$; Lee and Held 1993; Chang and Orlanski 1993). In this sense, it can crudely reflect the variability of the Atlantic storm track even though such a definition is not precise.

In the following sections we will focus on examining how the two types of storm tracks defined above affect the spatial structure of the eddy-driven NAO anomaly. A finite difference scheme similar to that used in Luo (2005) is utilized to solve (15) and (16) if the initial value of the NAO anomaly is given.

4. Phase of the North Atlantic Oscillation and its link with the variation of the Atlantic storm-track strength

In this section, the NDD storm track will be considered in order to investigate the impact of a change in the Atlantic storm-track strength on the variability of the NAO. Here, the fixed parameters are listed in Table 1, but $d_0$ is allowed to vary to represent the variation of the Atlantic storm-track intensity. Without loss of generality we may choose $B(x, 0) = 0.4$ as the initial amplitude of the eddy-driven NAO anomaly although the phase of the NAO anomaly is dominated by the sign of $\alpha$ and $m$ in (13) and (14). For this reason, $\alpha = -1$ and $m = -2\pi/\lambda_1$ ($\alpha = 1$ and $m = 2\pi/\lambda_2$) represent the negative (positive) phase of the NAO event (Luo et al. 2007a). Since NAO variability can be understood as a nonlinear initial-value problem (Benedict et al. 2004; Franzke et al. 2004), the initial value with a low-over-high (high over
low) anomaly can be chosen as the initial condition of the positive (negative) phase NAO event. On the other hand, positive-phase NAO events are dominant during 1978–2009 since there is a northward excursion of the Atlantic westerly jet (Luo et al. 2008). To assure that the Atlantic basin corresponds to the region of the negative topographic height from \( x = 2.87 \) to \( x = 2.87 \), \( h_0 = 0.4 \) must be required for the negative (positive) phase (Luo et al. 2007b). In this study, we define the Atlantic basin to be confined to the region of \( 2 \leq x \leq 2.0 \) and the European continent to the region of \( 2 \leq x \leq 5.74 \).

For the NDD storm track corresponding to \( \psi_1 \psi_2 = 0 \), its maximum intensity is assumed to be located at \( x_0 = 2.87 \) as in Luo et al. (2007a,b). However, for the DD storm track, we may fix the parameters of \( \psi_1 \psi_2 \) but vary those of \( \psi_1 \psi_2 \). The initial streamfunction anomalies of the NDD storm track are shown in Figs. 4a and 4b for \( a_0 = 0.17 \) and \( a_0 = 0.23 \), respectively. Figure 4c shows the initial streamfunction anomaly of the DD storm track for \( a_0 = 0.17 \), \( f_0 = 0.07 \), \( \mu_1 = \mu_2 = 1.2 \), and \( x_e = 2.87 \). It is seen that for the NDD storm track, the initial synoptic-scale eddies are evidently stronger for \( a_0 = 0.23 \) than for \( a_0 = 0.17 \). In contrast, for a DD storm track, the initial synoptic-scale eddies, shown in Fig. 4c, are also much stronger than those in Fig. 4a, but slightly weaker than those in Fig. 4b. Although under the same condition the eddies in an initial DD storm track are stronger than those in an initial NDD storm track, the initial eddies in the DD storm track can spread their energy downstream and finally cause the downstream intensification and extension of the Atlantic storm track (not shown). If the value of \( f_0 \) with \( i = 1, 2 \) is smaller, the downstream development of the Atlantic storm track is less distinct (not shown). For this case, the intensity of the DD storm track can be slightly stronger than that of the NDD storm track.

Figure 5 shows the long time evolution of the total streamfunction field \( \psi_T = \psi_T + \psi' \) of an eddy-driven NAO event with \( a_0 = 0.17 \) for its two phases in a NDD storm track. It is found that in a barotropic model, eddy-driven negative-phase NAO (NAO−) events can reappear after the decay of a NAO− event for a given NAO− initial condition in the Atlantic basin (Fig. 5a), which looks identical to blocking flows occurring in the Atlantic sector (Luo et al. 2007a; Woollings et al. 2008).

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Topographic height ( h_0 )</td>
<td>0.4</td>
</tr>
<tr>
<td>Position of positive topographic wave anomaly ( x_T )</td>
<td>-2.0</td>
</tr>
<tr>
<td>Eddy strength</td>
<td>0.17–0.23</td>
</tr>
<tr>
<td>Wavenumbers of NDD storm-track eddies ( k_i ) (( i = 1, 2 ))</td>
<td>As in Luo et al. (2007a)</td>
</tr>
<tr>
<td>Zonal distribution of NDD storm-track eddies ( \mu )</td>
<td>1.2</td>
</tr>
<tr>
<td>Position of the maximum eddy strength ( x_0 )</td>
<td>2.87/2</td>
</tr>
<tr>
<td>Positive small parameter ( \varepsilon )</td>
<td>0.24</td>
</tr>
<tr>
<td>Width of ( \beta ) plane channel ( L_\beta )</td>
<td>5.0</td>
</tr>
<tr>
<td>Referenced latitude ( \phi_0 )</td>
<td>55°N</td>
</tr>
<tr>
<td>Square of the ratio of the length scale and deformation radius ( F )</td>
<td>1.0</td>
</tr>
<tr>
<td>Zonal distribution of downstream developing storm-track eddies ( \mu_i ) (( i = 1, 2 ))</td>
<td>( \mu_1 = \mu_2 = 1.2 )</td>
</tr>
<tr>
<td>Strength of developing eddies ( f_0 )</td>
<td>0–0.07</td>
</tr>
</tbody>
</table>
FIG. 5. Long time evolution of the total streamfunction field of an NAO event obtained from the weakly nonlinear NAO model with the eddy forcing induced by eddies in a NDD storm track for $a_0 = 0.17$ and $x_0 = 2.87/2$ (CI = 0.3): (a) negative phase and (b) positive phase.
Fig. 6. Long time evolution of a positive-phase NAO event obtained from the weakly nonlinear NAO model with the eddy forcing induced by the enhanced eddies in a NDD storm track for $\alpha_0 = 0.23$ and $x_0 = 2.87/2$: (a) total field (CI = 0.3) and (b) planetary-scale field (CI = 0.15).
Fig. 7. Long time evolution of the total streamfunction field of a positive-phase NAO event obtained from the weakly nonlinear NAO model with the eddy forcing induced by eddies in a DD storm track for $a_0 = 0.17$ and $x_0 = 2.87/2$, and for the different position of the downstream development with $f_01 = f_02 = 0.07$ (CI = 0.3): (a) $x_e = 2.87$, (b) $x_e = 2.87/2$, and (c) $x_e = 0$. 
Within the 51-day period considered, there are three separate blocking events in Fig. 5a. The first event begins at about day 3 and is located slightly upstream of the Atlantic basin, which has a lifetime of about 13 days. The second blocking event begins at day 21 and its lifetime is about 13 days, and the third event begins at day 45 and then undergoes a life cycle that lasts for about 6 days. The latter two events are mainly located in the western Atlantic basin. Of course, they can exist in the central and eastern Atlantic if the initial NAO event is assumed to be located in the eastern Atlantic because most observed NAO$^+$ events initiate from the eastern Atlantic and European continent (not shown). At the same time, cyclonic wave breaking (CWB) is evident during the life cycle of each blocking event (Fig. 5a).

In contrast, a positive-phase (NAO$^+$) event can occur for a given NAO$^+$ initial condition (Fig. 5b). This event does not exhibit CWB (Benedict et al. 2004), and Franzke et al. (2004) noted that NAO$^+$ events correspond to anticyclonic wave breaking (AWB). However, Woollings et al. (2008) described an NAO$^+$ event as corresponding to an absence of CWB (i.e., an absence of blocking). During the period from day 0 to day 15, an NAO$^+$ event occurs for a given NAO$^+$ initial value in the Atlantic basin (Fig. 5b). The strongest zonal flow prevails at about day 9, which corresponds to the strongest stage of the NAO$^+$ event. An interesting feature we find is that a blocking event can occur over Europe ($2 \leq x \leq 5.74$) due to the energy dispersion of Rossby waves through the decay of the NAO$^+$ event, which is dominant from day 39 to day 45. This blocking event starts on day 27, decays, and then resumes again on day 39, undergoing further decay after day 45. This blocking flow resembles a standing wave and is mainly located over Europe. This case corresponds to a reduced number of blocking days over the Atlantic.

The long time evolution of an NAO$^+$ event is shown in Fig. 6 for the same parameters as in Fig. 5b but with $a_0 = 0.23$. It is seen that at day 0 there are several intense ridges and troughs in the NAO region and its upstream side. These ridges and troughs correspond to intense synoptic-scale eddies at the beginning of the NAO event. When these eddies interact with the preexisting planetary wave, they are absorbed by the mean flow by the weakening of eddies, resulting in a strengthened zonal flow that dominates over the North Atlantic (Fig. 6 at day 9). In this case, a typical NAO$^+$ event can form in the Atlantic basin. After day 9 the NAO$^+$ event tends to decay, and subsequent intense ridges and troughs appear in the NAO region or its upstream side. It is also evident that there is blocking beginning near day 24, followed by a retrograde movement of the block. Such a westward movement is more distinct during the period.

**Fig. 7.** (Continued)
from day 27 to day 42. Also, the retrograde behavior of the blocking can be clearly seen in the corresponding planetary-scale field (Fig. 6b). The explanation for why the European block can retrograde into the Atlantic region is easily explained in terms of the phase speed formula of the blocking flow obtained in the weakly nonlinear framework, which takes the form of

\[ C_P = u_0 - (\beta + F_{\mathcal{Q}})(k_1^2 + m_1^2 + F) - \delta M_0^2 \Delta \frac{k_1}{2k} \] (Luo 2000),

where \( \delta > 0 \) and \( M_0 \) is the blocking amplitude. It is clear that when the eddy forcing arising from the upstream storm-track eddy activity is rather strong, the amplitude of the resulting European block, due to the decay of the NAO \(^+\) event, becomes so large that \( C_P \) takes on a large negative value. In this case, the large-amplitude block over Europe will undergo a distinct westward movement. The retrograde shift of the European block can increase the number of days of the NAO \(^-\) event in the Atlantic even though the blocking flow as shown in Fig. 6 cannot get far enough upstream to reach the NAO \(^-\) region as in Fig. 5a. Thus, it is more likely that during the NAO \(^+\) winter the number of blocking days in the Atlantic is much larger for a stronger storm track than for a weaker storm track. That is to say, when the Atlantic storm-track eddies prior to the NAO are rather strong, the blocking flow occurring over Europe due to the decay of an NAO \(^+\) event can retrograde into the Atlantic basin and finally can become a long-lived Atlantic block.

The above conclusion can also be verified to be tenable for other parameters by performing sensitivity experiments (not shown).

5. The variability of the North Atlantic Oscillation and its link with the downstream development of an Atlantic storm track

In section 4, we have neglected the impact of the downstream development of the Atlantic storm-track activity on the NAO variability. In fact, there is stronger downstream development of the Atlantic storm-track eddy activity during P2 even though such downstream development is not statistically significant (Fig. 3c). In this section, we will consider the effect of the downstream development in a DD storm track \( (\psi'_{02} \neq 0) \) in order to examine whether the storm-track intensity or downstream development is more important for the variability of the NAO. Here, we only consider the case of \( k_{im} \rightarrow k_1 \) and use (15) to examine how the DD storm track affects the variability of the NAO. Because there are more NAO \(^+\) than NAO \(^-\) events during both P1 and P2, we can choose the initial structure of an NAO \(^+\) event as an initial condition of the planetary wave during the two periods. Although for some of winters the winter-mean NAO index is very small, much smaller than that associated with an NAO \(^+\) event, the initial flow with an NAO \(^+\) event can still be chosen because our attention is mainly focused on explaining why the NAO index decays from a value that is positive to a value that is close to zero or even negative within P2.

Without loss of generality, the long-time evolution of an eddy-driven NAO \(^+\) event under the eddy forcing for a DD storm track is shown in Fig. 7 for three cases of \( x_c = 2.87, x_c = 2.87/2, \) and \( x_c = 0. \) It is seen that when the Atlantic storm track exhibits distinct downstream development, an NAO \(^+\) event can form in the Atlantic sector and reach its mature stage with a strong low-overhigh pattern at about day 9. After this NAO \(^+\) event decays, a high-over-low pattern resembling a dipole blocking flow appears over the Atlantic and Europe at day 30. This block experiences further intensification at days 33 and 36, followed by retrogression and cyclonic wave breaking, leading to the occurrence of an NAO \(^-\) event in the Atlantic basin at day 39. This NAO \(^-\) event is able to continue until day 48 (Fig. 7a). Thus, a relatively long-lived NAO \(^-\) event can follow an NAO \(^+\) event through the retrograde shift of the enhanced blocking flow over Europe when the initial Atlantic storm track undergoes distinct downstream development. In particular, it is seen from a comparison between Figs. 7a and 7c that when the initial downstream development part is located more westward (Fig. 7a), the excited NAO \(^-\) event that follows an NAO \(^+\) event appears to be more prominent. This is because the initial upstream eddies over the Atlantic storm track are stronger for \( x_c = 2.87 \) than for \( x_c = 0 \) and in this case they can exert a more persistent forcing of a NAO anomaly downstream. As a result, a relatively marked increase in the number of days of the NAO \(^-\) event can be seen following a NAO \(^+\) event for \( x_c = 2.87. \) However, the transition of the NAO event from the positive to the negative phase is hardly observed once the downstream development is weaker (not shown). In addition, it must be pointed out that if there is only a pure downstream developed storm track \( (\psi'_{02} \neq 0 \text{ and } \psi'_{01} = 0) \), an NAO event for its two phases cannot be excited (not shown).

The above results indicate that NAO \(^+\) events can evolve into relatively long-lived NAO \(^-\) events through the occurrence and westward movement of more blocking events over Europe if the Atlantic storm track is sufficiently strong or if it can exhibit prominent downstream development. Although the transition from NAO \(^+\) to NAO \(^-\) takes place on intraseasonal scales, the interannual time scale variability of the NAO is closely connected to the intraseasonal scale of NAO events. This is because the slow interannual variation of the Atlantic storm track can affect the intraseasonal time
scale transition between NAO\(^+\) and NAO\(^-\) events, which in turn influences the winter-mean NAO and thus interannual NAO variability. Thus, the interannual trend of the NAO index can be reflected in part by the intraseasonal variability of the NAO associated with the variation of the Atlantic storm-track eddy activity.

The above results are based on the consideration that the initial Atlantic storm track and NAO anomaly are prespecified in our theoretical model. The two scale modes are coupled once they interact with each other. The NAO-related variability of the Atlantic storm track has been widely discussed by Lau (1988), Rogers (1997), and Wettstein and Wallace (2010), who found that the wintertime Atlantic storm-track variability and teleconnection patterns are dependent on each other. So it does seem difficult to precisely discern cause and effect between the Atlantic storm-track variability and the NAO pattern using only the observational data and diagnostic techniques. This problem may be addressed using our theoretical model (Luo et al. 2007a,b). This is because our model can reflect the coupling between the Atlantic storm-track variability and NAO anomaly. In addition, one particular merit of this model is that the different choice of the initial Atlantic storm track can crudely reflect the different variability of the Atlantic storm track during different periods. In this case, using our model might be helpful to examine the connection between intraseasonal and interannual scales.

6. Observational results

a. Relationship between the blocking action over the Atlantic–Europe sector and the phase of the NAO index

To test if the theoretical findings above are tenable, we first define a strong NAO\(^+\) (NAO\(^-\)) winter as that with a normalized NAO index that is equal to or greater (less) than \(+1.0\) \((-1.0\) standard deviation). The positive (negative) phase winters during 1950–2009 are 1980, 1982, 1983, 1987, 1988, 1990, 1992, 1993, 1994, 1999, and 2004 (1954, 1959, 1962, 1963, 1968, 1970, 1976, 1977, 1978, and 1984). It is clear that there are no strong NAO\(^-\) winters after 1984. Based on the two-dimensional blocking index proposed by Diao et al. (2006), the spatial distributions of blocking days during the NAO\(^+\) and NAO\(^-\) winters and their difference can be obtained (see Fig. 8). It is found that the center of the blocking action leans toward the Atlantic (Europe) for the negative (positive) phase winter. For the eastern Atlantic and European continent, the difference of blocking days between the NAO\(^+\) and NAO\(^-\) winters is positive and statistically significant at the 90% confidence level for a two-sided Student’s \(t\) test. This suggests a possibility that in winter the NAO\(^+\) occurrence is often followed by an increase in the number of blocking days in the eastern Atlantic and the European continent. This result is in agreement with the theoretical prediction shown in
FIG. 9. Long time evolution of a NAO event with the transition from the positive to the negative phase observed during the period from 1 Jan to 3 Feb 1996: (a) total field and (b) planetary-scale field, which is defined by the superposition of zonal wavenumbers 0–5.
FIG. 9. (Continued)
Fig. 9. (Continued)
FIG. 9. (Continued)
As further revealed in Fig. 6, blocking events occurring over the European continent due to the genesis of the NAO$^+$ event can retrograde into the eastern Atlantic when the Atlantic storm-track eddy activity is sufficiently strong. This provides a possible explanation for why more blocking days can be observed in the eastern Atlantic during the NAO$^+$ winter. However, we should look at the difference of blocking days between P2 and P1 because there are frequent NAO$^-$ events within both P1 and P2 and because the NAO$^-$ trend is different in the two intervals. If blocking days in the eastern Atlantic are more frequent during P2 than during P1, then the number of days of NAO$^-$ events should be larger for P2 than for P1. Thus, it is conjectured that the downward trend of the winter-mean NAO index within P2 should be larger than that within P1. Because the blocking index proposed by Diao et al. (2003) noted that the block must retrograde before the negative NAO forms. Woollings and Hoskins (2008) found that the dynamical link of the northern annular mode (NAM) between the Atlantic and Pacific

**b. NAO phase transition and different occurrence of NAO$^-$ events**

To see whether the transition of an NAO event from the positive to negative phase, as obtained from our weakly nonlinear NAO model, can be observed in the real atmosphere, we show in Fig. 9a an NAO$^+$ event that evolves into an NAO$^-$ event during the period from 1 January to 3 February 1996. It is found that an NAO$^+$ event takes place from 1 January to 20 January 1996, reaching its peak strength from 15 January to 18 January. This is followed by an NAO$^-$ event that occurs in the Atlantic basin through the intensification and retrograde movement of a European block, which is particularly evident at 21–29 January 1996, where cyclonic wave breaking is very clear. The blocking intensification and its retrograde behavior are more evident in the planetary-scale field, as shown in Fig. 9b. This process is in line with our theoretical finding in Fig. 6b. Figure 10 shows the corresponding daily NAO index. It is seen that this NAO index can change from a positive to a negative value, thus showing transition from an NAO$^+$ to an NAO$^-$ event.

If the European blocking that occurs due to the breakdown of an NAO$^+$ event is sufficiently strong, it will retrograde into the Atlantic basin and become a relatively strong, long-lived NAO$^-$ event. That is to say, the mean number of days of winter NAO$^-$ events within P2 should be larger than that within P1. Because the NAO$^+$ winters during P1 have fewer NAO$^-$ days than do NAO$^+$ winters during P2, the winter-mean NAO index during P2 will decrease. To confirm this point, it would be helpful to find the difference in the number of winter-mean days of NAO$^-$ events between P1 and P2. Before performing this calculation we first look at the numbers of days of NAO$^-$ and NAO$^+$ events during P1 and P2. According to the daily NAO index available from the CPC, the numbers of days of NAO$^-$ and NAO$^+$ events during P1 and P2 can be counted respectively and are shown in Table 2. It is seen that for NAO events with long periods, especially for those events with a minimum period of 10 days, the number of and mean duration of NAO$^-$ (NAO$^+$) events is greater within P2 (P1). This shows that an increase (a decrease) of NAO$^-$ (NAO$^+$) days during P2 relative to during P1 may be the main cause for the decline of the winter-mean NAO index during P2.

Because NAO$^-$ events are identical to blocking events in the Atlantic (Luo et al. 2007a; Woollings et al. 2008), one can recognize the frequency of NAO$^-$ days if the frequency of blocking days in the Atlantic can be identified by the blocking index.

Since the blocking index proposed by Diao et al. (2006) is two-dimensional, it can show the spatial distribution of blocking events over the Atlantic–Europe sector during both P1 and P2. The blocking days during P1 and P2, and their difference, are shown in Fig. 11. It is seen that the maximum center of blocking days for P1 is located over western Europe, but it is over the northeastern Atlantic for P2 (Figs. 11a,b). The difference between P2 and P1, as shown in Fig. 11c, is statistically significant at the 90% confidence level for a two-sided Student’s t test, clearly indicating that blocking days in the Atlantic are relatively more frequent during P2 than during P1. Thus, it is concluded that the mean number of NAO$^-$ days per winter for P2 is larger than that for P1. Consequently one would observe the downward trend of the winter-mean positive NAO index during P2.
sectors is due to the westward shift of the high-latitude blocking (Nakamura and Wallace 1993). However, in this paper, we propose a conjecture that the transition of the NAO event from the positive to negative phase is attributed to the retrograde shift of enhanced blocking over Europe under a strong Atlantic storm-track background. A tempting question here is whether most of the observed NAO$^-$ events during P2 arise from the retrograde movement of European blocks accompanying the decay of NAO$^+$ events. If so, our hypothesis will be at work in the generation of NAO$^-$ events.

Figure 12 shows the day and event numbers for NAO$^-$ events lasting 7 days or more that originate from retrograding blocking anomalies over Europe. Here blocking events in the Atlantic basin are defined as NAO$^-$ events, which are selected according to the criteria for blocking presented by Diao et al. (2006). Moreover, a retrograde blocking event is defined to be a block with an onset date in the domain 0°–50°E, with a westward movement into the domain 90°–10°W and a maintenance time of at least 4 days. A local developing blocking event is defined as a block with the onset and maintenance in the domain 90°–10°W.

It is found from this figure that within P2 the day and event frequencies for NAO$^-$ (Atlantic blocking) events arising from retrograde blockings over Europe are respectively 2.65 and 1.83 times greater than those for local developing events. In contrast, within P1, the corresponding values are 1.66 and 1.25 times greater than those of local developing events. This difference shows strong supporting evidence that within P2 NAO$^-$ days are relatively more frequent, which may be attributed to the retrograde shift of European blocking anomalies occurring frequently during the decay of NAO$^+$ events. This is more likely when the Atlantic storm track exhibits higher intensity. This can also be confirmed by performing a lagged composite of daily NAO indices according to the Atlantic storm-track strength index presented in section 2. The correlation between NAO$^-$ events and the downstream development may be insignificant because the change in the downstream development from P1 to P2 as shown in Fig. 3c is not statistically

<table>
<thead>
<tr>
<th>Minimal period (days)</th>
<th>P1</th>
<th>P2</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>5</td>
<td>7</td>
</tr>
<tr>
<td>Winter-mean day/event number of NAO$^-$ events</td>
<td>14.6/1.5</td>
<td>12.0/1.0</td>
</tr>
<tr>
<td>Winter-mean day/event number of NAO$^+$ events</td>
<td>19.5/1.6</td>
<td>17.8/1.3</td>
</tr>
</tbody>
</table>

**TABLE 2.** Winter-mean day and event numbers of NAO$^-$ and NAO$^+$ events with different minimal periods during P1 and P2.

Fig. 11. Distributions of the winter-mean blocking days in the Atlantic–Europe sector during (a) 1978–90 (P1) and (b) 1991–2009 (P2) and (c) the difference between (b) and (a) (P2 minus P1), in which the shading denotes the region above the 90% confidence level for a two-sided Student’s $t$ test.
significant. However, this correlation can become significant if the Atlantic storm track can undergo marked downstream development (Figs. 7a–c).

c. Composite daily NAO index and its link with the Atlantic storm-track strength

To understand how the strength of the Atlantic storm track affects the variation of the daily NAO index, a lagged composite of daily NAO indices is performed according to the mean Atlantic storm-track strength index shown in Fig. 2. In this figure, we define storm-track strength anomalies with standard deviations equal to or greater (less) than $+0.8$ and $1.0$ ($-0.8$ and $-1.0$) as corresponding to strong (weak) storm tracks.

Here NAO$^+$ and NAO$^-$ events during both P1 and P2 are selected based on the standard deviations $\pm 0.8$ and $\pm 1.0$ from the daily NAO index obtained from the CPC.
The lagged composite of the daily NAO index is performed according to our definition of strong and weak Atlantic storm tracks and shown in Fig. 13, in which the solid (dashed) line represents the weak (strong) storm-track intensity. The variation of the composite NAO index shown in Figs. 13c and 13f is statistically significant at the 90% level with a Monte Carlo test. It is found that when the Atlantic storm track is particularly strong, the composite NAO index is more likely to decay from a positive to a negative value, which seems to be insensitive to the sampling of NAO events (Figs. 13c,f). In contrast, this trend is less distinct when the Atlantic storm track is relatively weak. For this case, the second peak of the composite NAO index can follow the first peak. This indicates that under a strong Atlantic storm-track background, NAO$^+$ events can be more frequently followed by NAO$^-$ events. This seems to cause the downward trend of the winter-mean NAO index in P2 and further supports our theoretical results.

7. Conclusions and discussion

The winter-mean (DJF) NAO index has been observed to undergo a change from a linear upward trend during 1978–1990 (P1) to a linear downward trend during 1991–2009 (P2). The cause for this change in the NAO index trend is the subject of this study. We propose a hypothesis that the decrease in the positive NAO index during P2 is possibly related to the marked intensification of the Atlantic storm-track eddy activity throughout P2. In this paper, a generalized weakly nonlinear NAO model (Luo et al. 2007b) is used to investigate this hypothesis and to test if downstream development of the Atlantic storm track can affect the NAO variability. In addition, some observational evidence is provided to support our hypothesis.

Since blocking events in the Atlantic basin are identical to NAO$^-$ events (Luo et al. 2007a; Woollings et al. 2008), the number of days for NAO$^-$ events must be
greater during P2 than during P1. This is shown by calculating the difference between the number of blocking days between P2 and P1. Using the two-dimensional blocking index proposed by Diao et al. (2006), it is found that there is a marked increase in blocking days in the eastern Atlantic during P2, which corresponds to an increased frequency of NAO$^-$ days.

In this study, we have proposed that the reversal in the interannual trend of the NAO index may be due to the more frequent retrograde drifting of European blocking events under the interannual intensification of the Atlantic storm track. For a strong Atlantic storm track, the decay of the NAO$^+$ event can excite a retrograding European block, which results in more NAO$^-$ days over the North Atlantic, and thus a decline of the winter-mean NAO index during P2. This mechanism can provide a dynamical link between intraseasonal and interannual variability of the NAO.

Various studies have suggested two possibilities for changes in the North Atlantic storm track: 1) an interannual Indian Ocean convection trend (Selten et al. 2004) and 2) global warming (Hall et al. 1994; Ulbrich and Christoph 1999). The role of Indian Ocean convection in the change in the Atlantic storm track is not widely discussed, while the effect of global warming on the Northern Hemisphere storm track has been examined in detail in some numerical studies (e.g., Ulbrich et al. 2008). In fact, as shown in Selten et al. (2004), Indian Ocean convection can excite Rossby wave trains that propagate into the North Atlantic and form a traveling storm-track train that resembles a DD storm track. Thus, it is plausible that this wave train can affect the intraseasonal variability of the NAO event by strengthening the downstream development of the Atlantic storm-track eddies. Therefore, the interannual trend in the NAO index can be related to a long-term trend in Indian Ocean convection due to the intraseasonal variability of NAO events.

In addition, it should be pointed out that the theoretical model used here is highly idealized because of the exclusion of the background baroclinicity and the setting of a weak uniform background westerly wind. Thus, it cannot be used to present a quantitative comparison with observations although the theoretical result bears a reasonably close resemblance to the observations. If the winter-mean NAO is positive, it means that positive-phase events are more intense and frequent in winter compared to negative-phase events even though there may be a transition between positive and negative phases. In contrast, negative-phase events can be more intense and frequent in the winter if the winter-mean NAO index is negative. Although the theoretical model is highly idealized, it may be applicable for interpreting why the southern annular mode index has declined since 2000 because the storm-track activity in the Southern Hemisphere has exhibited a marked intensification during the past decade.

Furthermore, it should be mentioned that the impact of other external forcings such as sea surface temperature anomalies were not considered in our model (Josey et al. 2001). These problems need to be investigated in future research.

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