The Corresponding Tropospheric Environments during Downward-Extending and Nondownward-Extending Events of Stratospheric Northern Annular Mode Anomalies

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ABSTRACT

Using the NCEP–NCAR reanalysis dataset, this study classifies stratospheric northern annular mode (NAM) anomalies during the negative or positive phase into two categories—anomalies extending into the troposphere [trop event (TE); referred to as negative or positive TEs] and those not extending into the troposphere [nontrop event (NTE); referred to as negative or positive NTEs], and the corresponding tropospheric environments during the TEs and NTEs are identified. Compared with that for the negative NTEs, the upward wave fluxes entering the stratosphere are stronger and more persistent during the negative TEs. Furthermore, the stronger and more persistent upward wave fluxes during the negative TEs are due to more favorable conditions for upward wave propagation, which is manifested by fewer occurrences of negative refractive index squared in the mid- to high-latitude troposphere and stronger wave intensity in the mid- to high-latitude troposphere. However, the tropospheric wave intensity plays a more important role than the tropospheric conditions of planetary wave propagation in modulating the upward wave fluxes into the stratosphere. Stronger and more persistent upward wave fluxes in the negative TEs, particularly wave-1 fluxes, are closely related to the negative geopotential height anomalies over the North Pacific and positive geopotential height anomalies over the Euro-Atlantic sectors. These negative (positive) geopotential height anomalies over the North Pacific (Euro-Atlantic) are related to the positive (negative) diabatic heating anomalies and the decreased (increased) blocking activities in the mid- to high latitudes. The subtropical diabatic heating could also impact the strength of the mid- to high-latitude geopotential height anomalies through modulating horizontal wave fluxes. For positive NAM events, the results are roughly similar to those for negative NAM events, but with opposite signal.

1. Introduction

Thompson and Wallace (1998) found that the leading mode of the empirical orthogonal function (EOF) of the Northern Hemisphere (NH) sea level pressure (SLP) in winter is characterized by an oscillating variation between the Arctic and the surrounding midlatitudes; they referred to this feature as the Arctic Oscillation (AO). Subsequently, Limpasuvan and Hartmann (1999) named this deep and nearly barotropic leading mode the northern annular mode (NAM). The NAM exists not only in the troposphere but also in the stratosphere and is closely associated with the stratospheric polar vortex (e.g., Baldwin and Dunkerton 1999, 2001; Mukougawa and Hirooka 2004; Luo et al. 2012; Mitchell et al. 2013; Kidston et al. 2015; Zhang et al. 2016, 2018; Huang et al. 2017). Previous studies have provided convincing evidence that the AO/NAM plays an important role in...
affecting the weather system and climate variability throughout the NH (e.g., Thompson et al. 2000, 2002; Thompson and Wallace 2001; Black 2002; Cohen and Barlow 2005; McAfee and Russell 2008; Tachibana et al. 2010; Park et al. 2010, 2011). For instance, during a stratospheric sudden warming (SSW) event, a negative phase of the NAM is likely to occur near the surface a few weeks after the first warming signal appears in the upper stratosphere (e.g., Baldwin and Dunkerton 2001; Nakagawa and Yamazaki 2006; Huang et al. 2017). The stratospheric NAM events can also impact the tropospheric precipitation, blocking activities, cold-air outbreaks, and tropospheric jet streams (e.g., Luo et al. 2012; Davini et al. 2014; Kidston et al. 2015).

Although there is an overall likelihood for stratospheric NAM signals to extend downward (Thompson et al. 2003), in some cases, the stratospheric NAM signals are unable to extend downward into the troposphere; that is, some NAM signals are confined to the stratosphere (e.g., Baldwin and Dunkerton 1999, 2001; Zhou et al. 2002; Black and McDaniel 2004; Martineau and Son 2013; Karpechko et al. 2017). It is a challenge to determine why some stratospheric NAM anomalies can extend into the troposphere while others cannot. Baldwin and Dunkerton (2001) found that only strong stratospheric NAM signals tend to connect with Earth’s surface, while weaker NAM signals are limited within the stratosphere. Zhou et al. (2002) found that a stratospheric warm anomaly can extend downward into the troposphere when the initial wave force is very large and the polar westerly wind is reversed. Black and McDaniel (2004) noted that whether a stratospheric NAM signal could extend downward into the troposphere depends on two aspects: 1) whether stratospheric polar vortex anomalies can descend to sufficiently low altitudes and 2) whether preexisting NAM anomalies are present in the troposphere. Hitchcock et al. (2013) found that long-lasting negative NAM signals (a weak polar vortex) have a stronger tropospheric response. Mitchell et al. (2013) noted that the tropospheric response to the stratospheric NAM events is related to the status of the polar vortex. Karpechko et al. (2017) reported that SSWs with a stronger negative NAM index around 150 hPa have a greater probability to extend downward to the troposphere. Clearly, the conditions under which stratospheric NAM signals can extend downward into the troposphere remain under debate.

Although the stratospheric NAM signals first occur in the mid- to upper stratosphere (Matsumo 1970; Baldwin et al. 1994), they are closely connected to the upward fluxes of tropospheric planetary waves (e.g., Nakagawa and Yamazaki 2006; Garfinkel et al. 2010; Zhang et al. 2016). For example, the tropospheric blockings can impact the upward fluxes of planetary waves and further influence the development of stratospheric NAM events (e.g., Martius et al. 2009; Castanheira and Barriopedro 2010; Davini et al. 2014; Huang et al. 2017, 2018). Other factors, such as the Aleutian low, Siberian high, and Eurasian snow cover, can also influence the tropospheric wave sources and stratospheric NAM events (e.g., Cohen and Entekhabi 1999; Cohen et al. 2007; Garfinkel and Hartmann 2008; Garfinkel et al. 2010). In addition, previous studies noted that stratospheric NAM events could be modulated by the tropospheric wave propagation environment (e.g., Li et al. 2007; Hu and Guan 2018). Thus, additional issues worthy of exploration are to what extent and how tropospheric processes and the tropospheric background environment impact the downward extension of stratospheric NAM signals into the troposphere.

In this study, we attempt to identify the major factors responsible for the downward extension of NAM signals into the troposphere. We also analyze the roles of tropospheric processes and the tropospheric environment in the downward extension of stratospheric NAM signals into the troposphere. This study is structured as follows: Section 2 presents the data and methods. Section 3 gives the tropospheric wave–propagating environment during negative NAM. The tropospheric wave intensity and the reasons for wave-intensity variation during negative NAM are analyzed in section 4. Tropospheric wave–propagating environment and wave intensity during positive NAM are given in section 5, and section 6 provides a summary of the findings.

2. Data and methods

The data used in this study are derived from the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis dataset for the period 1958–2016. Isobaric fields are archived as daily averages on 17 pressure levels (from 1000 to 10 hPa) with a horizontal resolution of 2.5° × 2.5°, including geopotential height, temperature, and horizontal velocities. The anomalies are calculated by subtracting the daily climatological values of a given variable during 1958–2016.

The daily NAM index is computed from the zonal-mean geopotential anomalies field following the method used in previous studies (e.g., Baldwin and Thompson 2009; Mitchell et al. 2013). The anomalies are obtained by removing the daily climatological values, which is smoothed with a 90-day low-pass filter. After weighting the anomalies by the square root of the cosine of latitude, the daily zonal-mean anomalies at poleward of

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20°N during November–April are used for EOF analysis. The zonally varying NAM patterns are obtained by regressing geopotential height anomalies onto the leading principal component (PC) time series and daily NAM index is calculated by projecting geopotential height anomalies onto the NAM patterns (Baldwin and Thompson 2009). Finally, the time series of the NAM index is normalized at each pressure level for the period 1958–2016 to ensure that the time series has unit variance.

Positive and negative NAM indices represent positive NAM events and negative NAM events, respectively. In this study, we only focus on cases with strong NAM signals in the upper stratosphere. We select the NAM events in which a 10-hPa normalized NAM index exceeding ±1.5 lasts for at least 5 consecutive days. Notably, two events with the same sign at 10 hPa are considered as different cases if they are separated by at least 10 days (Black and McDaniel 2004). Then, four categories of NAM events classified based on the normalized NAM index are analyzed in this study. If the normalized NAM index of a selected case is greater than ±1.0 at 100 hPa and followed by a normalized NAM index exceeding ±1.0 for more than 2 consecutive days at 500 hPa, this case is regarded as a stratospheric NAM event that can propagate into the lower troposphere [trop event (TE)]; otherwise, it is regarded as a stratospheric NAM event that cannot propagate into the lower troposphere [nontrop event (NTE)]. According to the above criteria, 55 negative NAM events with 22 TEs and 33 NTEs and 39 positive NAM events with 20 TEs and 19 NTEs were noted during the period 1958–2016.

We use composite analysis to analyze the general features of these events. To understand the temporal evolution of the downward extension of upper-stratospheric NAM signals, the composite evolution is constructed with respect to the day of the peak magnitude in the 100-hPa NAM index (Black and McDaniel 2004). Because the composite NAM signals extend to the surface approximately 5 days before the peak magnitude in the 100-hPa NAM index, we refer to 5 days before the peak magnitude as day 0 and the day of the peak magnitude as day +5. The analysis results are not sensitive to the adjustment of day 0. All of the selected NAM events and the dates of the peak magnitude of the 100-hPa NAM index (day +5) are listed in Table 1. One NAM event is divided into three 10-day-average stages, that is, stage 1 (from day −30 to −21), stage 2 (from day −20 to −11), and stage 3 (from day −10 to −1). Figure 1 shows the time evolution of NAM signals during TEs and NTEs. Downward extension of NAM signals is evident in TEs (Figs. 1a,d), which is consistent with previous studies (e.g., Baldwin and Dunkerton 2001; Karpechko et al. 2017), while the NAM signals in NTEs (Figs. 1b,e) are confined to the stratosphere. The differences in NAM signals between TEs and NTEs are statistically significant in the lower stratosphere and troposphere (Figs. 1c,f), suggesting that our classification of NAM events is reliable. We use the two-sided Student’s t test to assess statistical significance in the composite analysis, and the equivalent sample sizes for the two-sided Student’s t test are estimated using Eq. (4) in Zwiers and von Storch (1995). The joint hypotheses test is used to assess the statistical significance of the regression equation, and an asterisk indicates that the regression is statistically significant at the 95% confidence level.

The Eliassen–Palm (EP) flux (Andrews et al. 1987) and Plumb wave activity fluxes (Plumb 1985) are used to diagnose the propagation of waves in the two- and three-dimensional space, respectively. The meridional and vertical components of EP flux in spherical geometry are

<table>
<thead>
<tr>
<th>Table 1. The dates of the peak magnitude of the 100-hPa NAM index.</th>
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<tbody>
<tr>
<td>Negative TEs</td>
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<tr>
<td>18 Jan 1960</td>
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<td>18 Dec 1960</td>
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\[ F_r = \rho_0 a \cos \phi (\pi \frac{u' \theta}{u} - \bar{u}' \bar{u}) \],

\[ F_z = \rho_0 a \cos \phi (\theta - a \cos \phi) \frac{u' \theta}{u} - \bar{w}' \bar{u} \],

and

\[ \text{div} F = \nabla \cdot F_r(\rho_0 a \cos \phi) \].

Here, \( \rho_0 \) is the density; \( z \) is the altitude; \( a \) is the radius of Earth; \( \phi \) is the latitude; \( f \) is the Coriolis parameter; \( \theta \) is the potential temperature; \( u \) and \( v \) represent the zonal and meridional winds, respectively; and \( w \) is the vertical velocity. The overbars represent the zonal average, and primes represent zonal deviation. We ignore the term \( \bar{w}' \bar{u} \) because it is small relative to other terms.

The square of the refractive index is calculated as given in Andrews et al. (1987):

\[ n^2_k = \frac{\pi v}{a} - \left( \frac{k}{a \cos \phi} \right)^2 - \left( \frac{f}{2NH} \right)^2 \],

where

\[ \pi = \frac{2\Omega}{a \cos \phi} - \frac{1}{a^2} \left( \frac{\pi \cos \phi}{\cos \phi} \right) - \frac{f^2}{\rho_0} \left( \frac{\pi}{\rho_0 N^2} \right) \].

Expansion of the third term on the right-hand side of Eq. (5) yields

\[ \frac{f^2}{\rho_0} \left( \frac{\pi}{\rho_0 N^2} \right) = \left( \frac{f^2}{H N^2} + \frac{f^2}{N^4} \frac{dN^2}{dz} \right) \pi - \frac{f^2}{H N^2} \pi' \].

Here, \( \pi, H, N, \Omega, \) and \( k \) denote the meridional gradient of the zonal-mean potential vorticity (PV gradient), scale height, buoyancy frequency, Earth-rotation frequency, and zonal wavenumber, respectively. The planetary waves can propagate in regions of the positive refractive index squared and avoid negative values. Because of the overlapping of positive and negative refractive indices, the composite refractive index squared (not shown) is difficult to interpret, as indicated by Li et al. (2007). To address this issue, we use the frequency
of the negative refractive index squared (referred to as $Fn^2$) described by Li et al. (2007) to diagnose the tropospheric environment of wave propagation. Some regions with positive $n^2$ have easterly winds; however, planetary waves cannot propagate when the mean zonal winds are easterly; thus, these regions where zonal wind is easterly are regarded as regions with negative $n^2$. To calculate $ Fn^2$, the day with a negative $n^2$ is defined as 1; otherwise, it is defined as 0. Dividing by the total number of days yields the frequency of the days defined as 1 in each grid. Then, we can obtain the climatological-mean $ Fn^2$. The $ Fn^2$ anomalies are calculated by subtracting the daily climatological values. Accordin to Li et al. (2007), planetary waves are more likely to propagate in regions with a large $ Fn^2$. The blocking index is based on the 500-hPa geopotential height (Tibaldi and Molteni 1990; Davini et al. 2012, 2014). The equatorward and poleward geopotential height gradients are as follows:

$$\Delta_{eqw} = \frac{Z_{500}(\lambda, \varphi_0) - Z_{500}(\lambda, \varphi_s)}{\varphi_0 - \varphi_s}$$

and

$$\Delta_{pw} = \frac{Z_{500}(\lambda, \varphi_0) - Z_{500}(\lambda, \varphi_n)}{\varphi_n - \varphi_0},$$

where $\varphi_0, \lambda_0$, and $Z_{500}$ denote the latitude, longitude, and 500-hPa geopotential height, respectively. The range of $\varphi_0$ is from 30° to 75°N; $\varphi_n = \varphi_0 + 15^\circ$ and $\varphi_s = \varphi_0 - 15^\circ$. An instantaneous blocking event is identified when

$$\Delta_{eqw} < 0 \text{ and } \Delta_{pw} > -10 \text{ gpm } (1^\circ \text{ lat})^{-1}.$$  

(9)

To ensure the spatial and temporal scales of blocking, the same constraints are applied to the instantaneous blocking index as in Davini et al. (2012, 2014).

In this study, the vertically integrated total diabatic heating $\dot{Q}$ can be obtained following Jin et al. (2013):

$$\dot{Q} = c_p/g \sum_{P_i} QdP,$$  

(10)

where $Q$ is the diabatic heating rate at different pressure levels; $g$ and $c_p$ are the gravitational acceleration and specific heat, respectively. The value of $Q$ is computed using daily variables at 28 sigma levels from the NCEP–NCAR reanalysis dataset (Colucci 2010), including the large-scale condensation heating rate (LCHR), solar radiative heating rate (SRHR), longwave radiative heating rate (LRHR), vertical diffusion heating rate (VDHR), deep convective heating rate (DCHR), and shallow convective heating rate (SCHR). The diabatic heating rate at each level can be calculated as follows:

$$Q = DCHR + SCHR + LRHR + SRHR + LCHR + VDHR.$$  

(11)

3. Tropospheric wave–propagating environment during negative NAM

Previous studies have shown that the strength of stratospheric NAM events is largely modulated by upward fluxes of planetary waves originating from the troposphere (e.g., Matsumo 1970; Limpasuvan et al. 2004; Chen et al. 2005). Figure 2 shows the EP flux and its divergence anomalies during the negative TEs (top) and NTEs (middle) and their differences (bottom). During stage 1 of negative TEs (Fig. 2a), there exist enhanced upward wave fluxes from the troposphere into the stratosphere. Then, the tropospheric waves propagate poleward and upward into the stratosphere at mid- to high latitudes during stages 2 and 3 (Figs. 2b,c). Upward wave fluxes converge in the stratosphere, forcing easterly anomalies and a negative NAM pattern. For the TEs, there are also enhanced wave fluxes entering the stratosphere at the mid- to high latitudes, and the upward wave flux anomalies during stage 3 (Fig. 2f) are stronger than those during stages 1 and 2 (Figs. 2d,e). By contrast, the magnitude of the upward flux anomalies in the TEs is stronger than those in the NTEs (Figs. 2g,h), corresponding to the negative NAM signals in the TEs extending downward to lower altitudes than those in the NTEs (Figs. 1a–c). During stage 3, the upward wave flux differences between negative TEs and NTEs are not significant, and the differences of wave fluxes between negative TEs and NTEs are characteristic of the poleward propagation (Fig. 2f).

It is well recognized that wavenumbers 1–3 are the predominant waves that propagate into the winter stratosphere and subsequently disturb the polar vortex (e.g., Charney and Drazin 1961; Polvani and Waugh 2004; Charlton and Polvani 2007; Shi et al. 2017). Refractive index squared is an important factor to modulate the upward propagation of tropospheric waves (e.g., Limpasuvan and Hartmann 2000; Hu and Tung 2002; Hu et al. 2014, 2015, 2018). Next, we will analyze the differences in the frequencies of negative refractive index squared $Fn^2$ between negative TEs and NTEs. Figure 3 shows the tropospheric $Fn^2$ for wave 1 for the negative TEs (top) and NTEs (middle) and the $Fn^2$ differences between negative TEs and NTEs (bottom). For negative TEs (Figs. 3a–c), $Fn^2$ of wave 1 in the lower troposphere is larger than that in the upper troposphere, particularly in the high latitudes. The $Fn^2$ of wave 1 in the polar region is larger than that in the middle latitudes, indicating that it is
more difficult for planetary waves to propagate upward in the polar region than in the middle latitudes. The spatial distributions of $F_n^2$ in negative NTEs (Figs. 3d–f) are similar to those in negative TEs, while the magnitude of $F_n^2$ is different in the two types of NAM events. The $F_n^2$ differences between negative TEs and NTEs are not significant in the midlatitude upper troposphere during stages 1 and 2. However, in the polar region (60°–75°N), the values of $F_n^2$ during negative TEs are smaller than those during negative NTEs in stages 1 and 2 (Figs. 3g,h), indicating that the patterns of $F_n^2$ during negative TEs are more favorable for the upward propagation of wave 1. These $F_n^2$ differences in the polar region are mainly contributed by the PV gradient differences (not shown). In stage 3, $F_n^2$ during negative TEs is larger than that during negative NTEs (Fig. 3i). The spatial distributions of tropospheric $F_n^2$ in waves 2 and 3 are similar to those in wave 1, but the magnitude of $F_n^2$ for waves 2 and 3 is larger than that of wave 1 (not shown). The above analysis reveals that the tropospheric environment in the TEs is more favorable for upward propagation of planetary waves into the stratosphere than in the NTEs, which may lead to a more persistent downward extension of stratospheric negative NAM signals in the TEs.

4. Tropospheric wave intensity and the reasons for wave-intensity variation during negative NAM

In addition to the tropospheric wave propagation environment, other tropospheric wave properties, such as tropospheric wave intensity and wave duration, should also have large impacts on the upward propagation of planetary waves into the stratosphere. Previous studies have linked the variations in upward wave fluxes to the geopotential height anomalies in the troposphere (e.g., Garfinkel and Hartmann 2008; Garfinkel et al. 2010; Seo and Son 2012; Wang et al. 2018; Hu et al. 2018). To understand differences in the tropospheric wave sources between the TEs and NTEs, the geopotential height...
anomalies are compared during the TEs and NTEs in this section.

**Figure 4** shows the composites of 500-hPa geopotential height anomalies during the negative TEs (top) and NTEs (bottom) overlapped with the corresponding climatological-mean wave-1 and wave-2 height fields for reference. During the negative TEs (Figs. 4a–c), negative height anomalies and positive height anomalies are observed at the mid- to high latitudes over the North Pacific and eastern Europe, respectively. In addition, positive height anomalies are detected at the mid- to high latitudes over the North Atlantic during stages 2 and 3 (Figs. 4b,c). These anomalies are almost in phase with the climatological mean of wave 1 (Figs. 4a–c, black lines), leading to a persistently enhanced amplitude of wave 1. During the NTEs, negative height anomalies are also observed at the mid- to high latitudes over the North Pacific but are weaker than those in the TEs. No evident positive geopotential height anomalies are observed over eastern Europe in stages 1 and 2 (Figs. 4d,e). Therefore, the tropospheric wave 1 in the NTEs is not as strong and persistent as those in the TEs. In addition, the wave 2 during stage 3 of negative NTEs (Fig. 4f) is enhanced because of the positive height anomalies over Europe and the negative height anomalies over the North Atlantic in phase with the climatological-mean wave 2 (Figs. 4d–f, black lines), indicating that wave 2 likely also contributes the upward propagation of planetary waves.

The time series of $F_z$ anomalies at 100 hPa and their time-integrated values averaged from 40° to 75°N are shown in Figs. 5a and 5c. During the TEs (Fig. 5a), enhanced upward wave fluxes from the troposphere into stratosphere can be noted from day −40 to day 5 (black solid line), while they can only be observed from day −25 to day 5 during negative NTEs (Fig. 5c). The enhancement and persistence of upward wave fluxes is mainly due to wave 1 from day −40 to day 5 (Figs. 5a,c, red line), which is consistent with the above analysis (Fig. 4). Wave 2 (3) in the TEs is weaker (stronger) than that in the NTEs. However, the fluxes of waves 2 and 3 are not as significant as that of wave 1, suggesting that wave 1 plays a dominant role in the downward extension of negative NAM signals. The strength and duration of upward wave flux anomalies in the TEs are greater than those in the NTEs, causing

**Fig. 3.** Latitude–pressure cross sections of $Fn^2$ for wave 1 during negative (a)–(c) TEs and (d)–(f) NTEs and (g)–(i) their differences (TEs minus NTEs). The differences over the dotted regions are statistically significant at the 95% confidence level according to the Student’s $t$ test. Thick black lines are the tropopause.
more wave fluxes to enter the stratosphere (cf. dashed lines in Figs. 5a,c) and lower-extending altitudes of negative NAM signals in the TEs (Fig. 1a).

To further clarify the relative importance of the propagation environment and the intensity of tropospheric waves to the 100-hPa $F_z$ anomalies, we apply a multiple linear regression analysis to estimate the contribution of the tropospheric wave–propagating environment and its intensity to variations in 100-hPa wave fluxes (Figs. 5b,d). The proxies are the averaged $Fn^2$ anomalies of waves 1–3 (600–300 hPa, 50°–75°N) and 500-hPa area-averaged height anomalies over the North Pacific ($H_{Pac}$; 40°–75°N, 120°E–120°W), Europe ($H_{Eu}$; 40°–75°N, 0°–60°W) and the North Atlantic ($H_{At}$; 40°–75°N, 0°–60°W) based on the above analyses (Figs. 3–4). The $Fn^2$ anomalies represent the tropospheric wave–propagating environment, and 500-hPa height anomalies represent wave intensity. Because variations in the stratosphere lag behind those in the troposphere (e.g., Garfinkel et al. 2010), the lag time is also considered in the multiple linear regression based on the lag correlation coefficient between proxy variable and $F_z$ anomalies; that is, the days that $F_z$ anomalies lag behind the variables are the days that the lag correlation coefficient between the $F_z$ anomalies and proxy variable are maximal and have the same sign as the 0-day lag correlation coefficient. The regression is computed using the $F_z$ anomalies from day −40 to day 5, that is, between the day when anomalous waves begin to propagate upward and the day when anomalously upward waves disappear. The multiple regression model is written as follows:

$$F_z = aH_{Pac} + bH_{Eu} + cH_{At} + dF_{n^2} + \text{const},$$

where $a$, $b$, $c$, and $d$ are regression coefficients. The percentages that variables contribute to integral regression curve of $F_z$ anomalies from day −40 to day 0 are as follows:

$$\sum_{-40}^0 aH_{Pac} / \sum_{-40}^0 F_z, \sum_{-40}^0 bH_{Eu} / \sum_{-40}^0 F_z, \ldots$$

For the negative TEs (Fig. 5b), the negative height anomalies over the North Pacific and the positive height anomalies over Europe account for 52% and 14% of the
The contribution of $F_n^2$ anomalies is only 5%, which is much less than those of height anomalies. In other words, the variation in tropospheric wave intensity makes a greater contribution than wave propagation environment to the wave fluxes entering the stratosphere. In the negative NTEs (Fig. 5c), tropospheric wave intensity also plays a dominant role in controlling upward wave fluxes. The contributions of height anomalies to the integral wave fluxes are 105% and 41% over the North Pacific and the Europe, respectively, while $F_n^2$ anomalies contribute negatively ($-31\%$) to the integral wave fluxes (dashed line at day 0), further supporting that the tropospheric $F_n^2$ anomalies in negative NTEs are more unfavorable for the upward propagation of tropospheric waves than those in negative TEs (Fig. 3).

The differences in tropospheric wave source between negative TEs and NTEs are mainly associated with the variations in geopotential height over the North Pacific and the Euro-Atlantic sector. Next, we will further discuss the reasons for wave pattern differences between negative TEs and NTEs. Planetary wave sources are largely forced by topography and diabatic heating (e.g., Chen and Robinson 1992; Cohen et al. 2007; Yu et al. 2009; Seo and Son 2012; Simpson et al. 2016; Yoo and Son 2016). Here, we first focus on the differences in diabatic heating. Figure 6 shows the vertically integrated total diabatic heating anomalies for the negative TEs (top) and NTEs (bottom). Positive $\tilde{Q}$ anomalies are observed at the middle latitudes over the North Pacific and negative $\tilde{Q}$ anomalies at the mid- to high latitudes over the North Atlantic during stage 1 of the negative TEs (Fig. 6a), corresponding to the negative height anomalies over the North Pacific and positive height anomalies over the North Atlantic, respectively (Fig. 4a). During stages 2 and 3 (Figs. 6b,c), positive $\tilde{Q}$ anomalies still exist in the middle latitudes over the North Pacific, but negative $\tilde{Q}$ anomalies over the North Atlantic shift northward relative to those during stage 1, which also match with the height anomalies (Figs. 4b,c). According to the geopotential tendency equation and the vorticity equation (Liu et al. 2004; Vigh and Schubert 2009), positive and negative $\tilde{Q}$ anomalies favor the maintenance of negative and positive height anomalies (Figs. 4a–c), respectively. Compared with TEs, no significant and persistent positive $\tilde{Q}$ anomalies over the North Pacific are observed at the middle latitudes in the NTEs, which is in accordance with the weaker negative height anomalies throughout the NTEs (Figs. 6d–f). In addition, positive $\tilde{Q}$ anomalies over the North Atlantic are observed at the middle latitudes.

Figure 7 shows the vertical profile of the diabatic heating rate anomalies and their six terms (see section 2 for more details) over the North Pacific and the North Atlantic. Over the North Pacific, the VDHR (Figs. 7a–c, deep red lines) dominates the $\tilde{Q}$ anomalies below 850 hPa, while
the DCHR (deep blue lines) plays the most important role in the Q anomalies from 850 to 500 hPa. The positive Q anomalies above 500 hPa are evidently smaller than those below 500 hPa, and the LRHR (orange lines) mainly contributes these positive Q anomalies. Note that the positive Q anomalies can extend to the middle troposphere and decrease with altitude (Figs. 4a–c), favoring the maintenance of cyclonic anomalies (negative height anomalies). Over the Atlantic (Figs. 7d–f), the Q anomalies are negative, opposite to those over the North Pacific. The VDHR, DCHR, and LRHR contribute the majority of Q anomalies at the levels below 850 hPa, from 850 to 500 hPa, and above 500 hPa, respectively. These negative Q anomalies with magnitude decreased with altitude favor the maintenance of anticyclonic anomalies (positive height anomalies) over the North Atlantic. During negative NTEs, the positive Q anomalies and their vertical gradients (Figs. 7g–i) over the North Pacific are not as significant as those during negative TEs, in accordance with the weaker cyclonic anomalies during the negative NTEs. These weaker positive Q anomalies in negative NTEs are related to the weaker VDHR and DCHR compared with those in negative TEs. Over the North Atlantic (Figs. 7j–l), positive Q anomalies, whose magnitude decreases with altitude, are observed in negative NTEs, while negative Q anomalies are present in negative TEs (Figs. 7d–f). These positive Q anomalies restrain the formation of anticyclonic anomalies in negative NTEs. The above analysis results show that the diabatic heating at middle latitudes is more favorable for the enhancement of planetary waves during negative TEs than those during negative NTEs.

In addition to the diabatic heating at mid- to high latitudes, the diabatic heating related to tropical or subtropical meteorological processes could also exert a remote impact on the tropospheric planetary waves entering the stratosphere (Yoo et al. 2011, 2012; Seo and Son 2012; Wang et al. 2018). The tropical divergence induced by diabatic heating can lead to a Rossby wave source in the subtropical westerlies (Sardeshmukh and Hoskins 1988; Lin 2009). Seo and Son (2012) reported that Rossby waves generated over the subtropics can propagate northward to the northern Pacific and North America, which is also reported by Wang et al. (2018).
Figure 8 shows horizontal wave flux and zonal wind anomalies for the negative TEs (Figs. 8a–c) and NTEs (Figs. 8d–f). During stage 1 of negative TEs, the negative $Q$ anomalies over the subtropical Pacific (Fig. 6a) force an anticyclonic anomaly ($20^\circ$–$40^\circ$N, $180^\circ$–$150^\circ$W; Fig. 4a) with easterly anomalies on its north side (Fig. 8a, colors). These subtropical easterly anomalies can lead to waves anomalously propagating equatorward, which constitute a wave train from the subtropical Pacific to the North Atlantic (Fig. 8a). The anomalously equatorward waves diverged in the midlatitude Pacific favor the maintenance of westerly anomalies and cyclonic anomalies in the mid- to high latitudes. Further, the waves entering the Atlantic decrease and anomalous waves propagate poleward over the Atlantic. This wave train can also be seen in the geopotential height anomalies, with positive anomalies over the subtropical Pacific and North America and negative anomalies over the midlatitude Pacific (Fig. 4a). The Pacific wave train in stage 2 (Fig. 8b) is weaker than that in stage 1, related to the weakening of subtropical negative $Q$ anomalies (Fig. 6b). In stage 3 (Fig. 8c), a wave train opposite those in stages 1 and 2 is observed from the subtropical Pacific to the North Atlantic and weakens the amplitude of cyclonic anomalies in the mid- to high latitudes over the North Pacific. However, the anomalous convergence of poleward wave fluxes favors the easterly anomalies related to anticyclonic anomalies over the Euro-Atlantic sectors. During negative NTEs, the anomalously equatorward wave fluxes are observed over the North Pacific in stages 1–3 (Figs. 8d–f). However, the equatorward wave fluxes related to the subtropical diabatic heating are weaker in stages 1 and 2 than those during negative TEs, corresponding to the weaker cyclonic anomalies in the high latitudes over the North Pacific. As a whole, the mid- to high-latitude wave pattern (Fig. 4) can also be modulated by the horizontal wave fluxes related to the subtropical diabatic heating.

In addition to the heat source force, previous studies have noted that anomalous upward planetary waves are closely related to atmospheric processes, such as tropospheric blockings (e.g., Martius et al. 2009; Nishii et al. 2009, 2011; Huang et al. 2017, 2018). Figure 9 shows the blocking-frequency anomalies for the TEs (top) and
During the TEs (Figs. 9a–c), blocking frequencies consistently increase over the northern Euro-Atlantic sector and decrease over the Bering Sea. These anomalous distributions of blocking frequency correspond to the height anomalies (Figs. 4a–c) and are associated with the enhanced wave-1 flux. This finding is consistent with previous results (e.g., Castanheira and Barriopedro 2010; Huang et al. 2017, 2018). For the NTEs (Figs. 9d–f), decreased (increased) blocking frequencies are not significant over the Bering Sea (Euro-Atlantic) sectors compared with those in negative TEs, leading to weaker negative height anomalies and a weaker flux of wave 1 than those in the TEs. The above analysis shows that tropospheric blocking variations may also contribute to the wave source differences in Fig. 4.

5. Tropospheric wave–propagating environment and wave intensity during positive NAM

As shown in Fig. 1, the positive stratospheric NAM signals also show downward extension into the troposphere; therefore, the positive NAM events are also analyzed here. Similar to the process of negative NAM, the formation and downward extension of positive stratospheric NAM are also related to upward wave activities. Figure 10 shows the time series of the 100-hPa $F_z$ during positive TEs and NTEs. Upward wave fluxes into the stratosphere decreased during both positive TEs and NTEs, mainly because of the reduced wave 1. The duration of wave flux weakening in positive TEs persists longer than those in positive NTEs, leading to fewer wave fluxes entering the stratosphere than those in positive NTEs (cf. Figs. 10a and 10c). The $F_n^2$ anomalies contribute positively (9%) to the integral regressive $F_z$ anomalies (Fig. 10b), while $F_n^2$ anomalies contribute negatively (−14%) to the integral regressive $F_z$ anomalies in positive NTEs (Fig. 10d). Additionally, contributions of $F_n^2$ anomalies to the integral $F_z$ anomalies are weaker than those of tropospheric wave intensity during both positive TEs and positive NTEs. By contrast, the distributions of $F_n^2$ in positive NAM are similar to those in the negative NAM. The $F_n^2$ of positive TEs is larger than those of positive NTEs in the polar region during stage 1 and in the midlatitudes during stages 2 and 3 (not shown), indicating a less favorable...
environment for the upward propagation of planetary waves during positive TEs. The height anomaly differences between positive TEs and NTEs are mainly located over the North Pacific, that is, more persistent positive height anomalies are observed during positive TEs than during positive NTEs, causing a persistent weakening of wave 1 in the positive TEs. The differences in height anomalies between positive TEs and NTEs are also related to the diabatic heating, blocking activities, horizontal wave activities, etc. The results for positive NAM are roughly similar to those for negative NAM except for opposite signals, thus, the detailed analysis about positive NAM cases are not discussed here.

6. Summary and discussion

The main purpose of this study is to analyze and clarify the corresponding tropospheric environments during downward-extending and nondownward-extending events of stratospheric NAM signals. In this study, NAM events are divided into two types using the NCEP–NCAR reanalysis dataset according to whether the NAM signals (including negative and positive) can propagate into the lower troposphere. More waves enter the stratosphere during the negative TEs than during the negative NTEs. Furthermore, the wave flux differences in the stratosphere between the negative TEs and NTEs are mainly related to the tropospheric wave–propagating environment and tropospheric wave intensity.

The tropospheric environment can be characterized by the frequency of negative refractive index squared. The $F_{n}^2$ values during negative TEs in the polar region are lower than those during negative NTEs in most stages, implying that $F_{n}^2$ in the TEs is more favorable for upward propagation of planetary waves into the stratosphere than that in the NTEs. In addition to the wave-propagating environment, wave intensity is another nonnegligible factor influencing stratospheric wave fluxes and plays a more important role than the wave-propagating environment in the stratospheric wave anomalies. The upward tropospheric wave fluxes in negative TEs are stronger than those in negative NTEs, particularly the wave flux of wave 1. Stronger wave fluxes in the negative TEs are

![Composite anomalies of the blocking frequency during negative TEs and NTEs. The colors over the dotted regions are statistically significant at the 95% confidence level according to the Student’s t test.](http://journals.ametsoc.org/jcli/article-pdf/32/6/1857/4886373/jcli-d-18-0574_1.pdf)
caused by larger negative height anomalies over the North Pacific and positive height anomalies over the Euro-Atlantic sector related to diabatic heating, horizontal wave propagation, and blocking frequencies. For positive NAM anomalies, the results are roughly similar to those for negative NAM anomalies, but with opposite signal.

In this study, we mainly investigate the corresponding tropospheric environments during TEs and NTEs of the NAM signal. The differences in diabatic heating, blocking activities, and horizontal waves between TEs and NTEs are analyzed, and their influences on the stratospheric planetary wave fluxes are also discussed. It should be pointed out that the diabatic heating may be related to multiple factors, such as the Madden–Julian oscillation (Seo and Son 2012; Yoo and Son 2016; Son et al. 2017), El Niño–Southern Oscillation (Garfinkel and Hartmann 2008; Garfinkel et al. 2010; Xie et al. 2012), and the North Pacific sea surface temperature (Tian et al. 2017; Hu and Guan 2018; Li et al. 2018). Which factors and how these factors dominate the diabatic heating differences between TEs and NTEs are beyond the scope of this study but deserve further investigation.

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REFERENCES
Castanheira, J. M., and D. Barriopedro, 2010: Dynamical connection between tropospheric blockings and stratospheric


