The Combined Influences of Westerly Phase of the Quasi-Biennial Oscillation and 11-year Solar Maximum Conditions on the Northern Hemisphere Extratropical Winter Circulation

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Abstract

The combined influences of the westerly phase of the quasi-biennial oscillation (QBO-W) and solar maximum ($S_{\text{max}}$) conditions on the Northern Hemisphere extratropical winter circulation are investigated using reanalysis data and Center for Climate System Research/National Institute for Environmental Studies chemistry climate model (CCM) simulations. The composite analysis for the reanalysis data indicates strengthened polar vortex in December followed by weakened polar vortex in February–March for QBO-W during $S_{\text{max}}$ (QBO-W/$S_{\text{max}}$) conditions. This relationship need not be specific to QBO-W/$S_{\text{max}}$ conditions but may just require strengthened vortex in December, which is more likely under QBO-W/$S_{\text{max}}$. Both the reanalysis data and CCM simulations suggest that dynamical processes of planetary wave propagation and meridional circulation related to QBO-W around polar vortex in December are similar in character to those related to $S_{\text{max}}$, furthermore, both processes may work in concert to maintain stronger vortex during QBO-W/$S_{\text{max}}$. In the reanalysis data, the strengthened polar vortex in December is associated with the development of north–south dipole tropospheric anomaly in the Atlantic sector similar to the North Atlantic oscillation (NAO) during December–January. The structure of the north–south dipole anomaly has zonal wavenumber 1 (WN1) component, where the longitude of anomalous ridge overlaps with that of climatological ridge in the North Atlantic in January. This implies amplification of the WN1 wave and results in the enhancement of the upward WN1 propagation from troposphere into stratosphere in January, leading to the weakened polar vortex in February–March. Although WN2 waves do not play a direct role in forcing the
stratospheric vortex evolution, their tropospheric response to QBO-W/S\textsubscript{max} conditions appears to be related to the maintenance of the NAO-like anomaly in the high-latitude troposphere in January. These results may provide a possible explanation for the mechanisms underlying the seasonal evolution of wintertime polar vortex anomalies during QBO-W/S\textsubscript{max} conditions and the role of troposphere in this evolution.

**Keywords** quasi-biennial oscillation; 11-year solar cycle; polar vortex; planetary wave; North Atlantic Oscillation; stratosphere-troposphere coupling

1. **Introduction**

Observed variations in polar vortex intensity in Northern Hemisphere (NH) winter are related to the quasi-biennial oscillation (QBO) in the equatorial stratosphere. This relationship was first suggested by Holton and Tan (1980, 1982) and is now known as the Holton–Tan effect (see Anstey and Shepherd 2014 for a recent review). Holton and Tan found that in the composite average, the zonal wind of the NH polar vortex is anomalously strong during the westerly phase of the QBO (QBO-W) at 50 hPa, whereas the polar vortex is anomalously weak in the easterly phase of the QBO (QBO-E) at 50 hPa. The equatorial zero wind line, which represents a critical line for stationary planetary waves, lies at different latitudes for QBO-W and QBO-E at 50 hPa. This north–south shift of the critical line in the equatorial stratosphere between QBO-W and QBO-E was conventionally thought to provide a causal connection from the QBO to the NH polar vortex, affecting the propagation of planetary waves toward the equatorial region (Holton and Tan 1980, 1982). Watson and Gray (2014) have provided recent support for this mechanism by showing that the reflection of waves around the critical line in the equatorial lower stratosphere directly accounted for the deceleration of the polar vortex for at least the first eight days of their transient experiment under QBO-E conditions. However, stratosphere-resolving climate models and the extension of the upper boundary of observational reanalyses allow the inclusion of the equatorial upper and middle stratosphere in the statistical analysis of QBO influences on the polar vortex using both models and reanalyses, and a causal connection from the critical line shift in the equatorial upper/middle stratosphere to the NH polar vortex has also been suggested (e.g., Gray et al. 2001a, b; Pascoe et al. 2006; Naoe and Shibata 2010). Yamashita et al. (2011) proposed this mechanism as a possible explanation for QBO-vortex coupling, suggesting that the north–south shift of the critical line in the equatorial upper/middle stratosphere is important for the propagation of planetary waves toward the equatorial region and causes a change in meridional circulation over the whole depth of the stratosphere, which influences the NH polar vortex. Garfinkel et al. (2012) provide support for this line of argument, using transient experiments with an idealized model under perpetual QBO-E and fixed radiative conditions. Overall, there is still no agreement on the mechanisms of QBO–vortex influence, but there is no doubt about the existence of the effect and the sign of the response in the NH polar vortex.

The influence of solar activity on climate is a long-standing issue, and the influence of the 11-year solar cycle has been explained as follows. The solar irradiance change in ultraviolet (UV) radiation near 200 nm and at 240–320 nm between solar maximum (S\textsubscript{max}) and solar minimum (S\textsubscript{min}) are approximately 6 % and 4 %–8 %, respectively (e.g., Gray et al. 2010). The former is important for the changes in oxygen dissociation and ozone production, and the latter is important for the changes in the absorption of solar flux by ozone, which are associated with the observed temperature changes in the upper/middle stratosphere and mesosphere (e.g., Brasseur and Solomon 1986; Gray et al. 2010). Another temperature maximum is observed with the solar cycle in the tropical lower stratosphere, but there is no direct explanation for it, and its attribution to solar forcing remains controversial (e.g., Gray et al. 2010). In NH winter, the Sun illuminates the Southern Hemisphere, and the NH low- and mid-latitudes, whereas sunlight does not reach NH high latitudes (the polar night region). It thus makes sense that the influence of the solar UV radiation changes with the 11-year cycle exists only in these sunlit regions, and the latitudinal temperature gradient is larger around the edge of the sunlit regions during S\textsubscript{max}. The latitudinal temperature gradient in the NH mid-latitudes is related to the intensity of the zonal wind in the upper stratosphere/lower mesosphere. Thus, in those regions, S\textsubscript{max} could magnify the
wind speed around the polar vortex indirectly as well as directly through wave–mean flow interaction (e.g., Kodera and Kuroda 2002).

The combined solar–QBO effect on the NH winter vortex is a challenging topic, and it is still the subject of observational and modeling studies (Gray et al. 2010; Matthes et al. 2013). The pioneering works of Labitzke (1987) and Labitzke and van Loon (1988) demonstrated a statistically significant warming of the NH polar stratosphere in late winter under QBO-W and $S_{\text{max}}$ (QBO-W/$S_{\text{max}}$) years. Later works (e.g., Naito and Hirota 1997; Labitzke et al. 2006; Camp and Tung 2007; Anstey and Shepherd 2014) with larger datasets have confirmed the observational results of Labitzke and van Loon (1988). A number of modeling studies has simulated the combined solar–QBO effect on the NH winter vortex, and the results of Labitzke and van Loon (1988) have been reproduced by experiments under constant solar flux and QBO conditions (Matthes et al. 2004, 2010), under constant solar flux conditions with an internally generated QBO (Schmidt et al. 2010), and under imposed forcing with observed solar cycle and QBO conditions (Chiodo et al. 2012).

Kodera (1991) and Kodera et al. (1991) proposed a mechanism for the combined solar–QBO effect on the NH winter vortex based on observations and experiments with an idealized general circulation model (GCM) with a simplified solar cycle and QBO forcing. They indicated that in the early winter, both the solar activity and QBO produce initial zonal wind anomalies in the upper stratosphere, which are transported to the lower stratosphere in late winter through wave–mean flow interaction. Later work (Kuroda and Kodera 2002) confirmed the downward transport of the initial anomalies as a projection onto the polar-night jet oscillation (PJO) and analyzed its seasonal evolution. Gray et al. (2004) analyzed the output of a mechanistic model experiment and ERA-40 reanalysis data and showed that the timing of stratospheric sudden warmings (SSWs) is influenced by the solar activity and the phase of the QBO. They proposed a mechanism in which the reinforcement of the influences of the QBO-W and $S_{\text{max}}$ on the zonal wind anomalies in the upper stratosphere corresponds to the late timing of the SSWs, resulting in the high SSW frequency in late winter, following the work of Kodera et al. (1991).

Several studies have worked on the confirmation of the mechanism proposed by Kodera et al. (1991) and Gray et al. (2004). For example, Matthes et al. (2013) ran version 3.1 of the Whole Atmosphere Community Climate Model (WACCM3.1) under time-varying forcing of the solar cycle and QBO. They derived a high SSW frequency in late winter under QBO-W/$S_{\text{max}}$ conditions, corresponding to the observation of Labitzke and van Loon (1988) and the mechanism of Kodera et al. (1991) and Gray et al. (2004). As WACCM3.1 has an upper boundary in the lower thermosphere, the mesosphere was well resolved and Matthes et al. (2013) suggested that the initial zonal wind anomalies in the early winter are produced around the stratosphere and mesosphere rather than the upper stratosphere, which affects the evolution of the PJO in the stratosphere during the winter.

Several studies have noted the influence of solar variability on surface climate. Kodera (2002) analyzed reanalysis data and found a north–south dipole anomaly of sea level pressure (SLP) over the North Atlantic region, which was synchronized with the modulation of the PJO. A similar signal was derived from the UK Hadley Centre sea level pressure time-series (Gray et al. 2013), from a GCM experiment (Matthes et al. 2006), from a chemistry climate model (CCM) experiment (Chiodo et al. 2012), and from the ensemble mean of the Coupled Model Inter-comparison Project Phase 5 models, especially the high-top models (Mitchell et al. 2015). Note that Gray et al. (2004) and Matthes et al. (2013) only touched briefly on the timings and signs of the tropospheric signals during QBO-W/$S_{\text{max}}$.

In this study, we investigate the combined influences of QBO-W and $S_{\text{max}}$ on the stratosphere and troposphere mid- and high-latitudes using both CCM simulations with interactive ozone chemistry (where there are no influences from volcanic eruptions and sea surface temperature (SST) variability) and reanalysis data. We do not address the mechanism behind the initial upper stratospheric anomaly associated with solar–QBO conditions, which remains an open issue. Instead, we focus on the role of the troposphere in the subsequent evolution of the polar vortex through the winter season, which involves an interaction between stratospheric and tropospheric processes.

The reanalysis data and model experiments are described in Section 2 along with the analysis method. The results of our analysis are presented in Section 3, and a summary is provided in Section 4.

2. Data and analysis method

2.1 Data

The reanalysis data used in this study were the European Centre for Medium-Range Weather Forecasts (ECMWF) Interim Re-Analysis (ERA-Interim)
2.2 Model description and experimental setup

The CCSR/NIES CCM used for the experiments was developed from version 5.4 g of the CCSR/NIES atmospheric GCM with a T42 horizontal resolution and 34 vertical layers with the upper boundary located at approximately 80 km. See Nagashima et al. (2002) and Akiyoshi et al. (2009) for details. Yamashita et al. (2010) suggested that a simulation period of ~140 years would be sufficient for detecting the solar cycle response in a transient run. In this study, we use simulations for 138 winters from three ensemble runs of 47 years each from 1 January 1960 to 31 December 2006. Note that the experiment was performed for 56 years from 1951 to 2006, including a 9-year spin-up run with different initial conditions at 1 January 1951. The outputs from 1 January 1951 to 31 December 1959 were abandoned.

The method of QBO inclusion was the same as in Akiyoshi et al. (2009) and Yamashita et al. (2011); the zonal-mean zonal wind around the equator in the CCM was nudged toward the observational QBO wind profiles from a merged dataset of equatorial zonal wind profiles at three observation sites provided by the CCM validation (CCMVal) activity. The latitude–height range of the nudging was 20°S–20°N and 4–76 hPa.

The effects of the 11-year solar cycle were included using the same method as in Akiyoshi et al. (2009) and Yamashita et al. (2010); the range of solar flux variation in each spectral bin between maximum and minimum solar activity was determined by a solar spectral observation at the solar maximum (November 1989) and solar minimum (September 1986) reported by Lean et al. (1997). The monthly mean 10.7 cm solar radio flux (F10.7) was used as a time evolution factor of each spectral bin. Note that the scaling of the F10.7 is a crude approximation of the solar signal, and the model response might differ from that of the ERA-Interim data.

Austin et al. (2008), Eyring et al. (2010), and Morgenstern et al. (2010b) reported that the multiple regression analysis of the CCSR/NIES CCM outputs for the REF-B1 experiment of the CCMVal-2 activity of the stratospheric processes and their role in climate project produced an unrealistically large coefficient in the 11-year solar cycle term of ozone concentration variation in the tropical lower stratosphere. Yamashita et al. (2010) reported that this large coefficient was mainly because the solar term in the regression analysis erroneously includes an overestimated contribution of the volcanic eruptions due to the 9-year interval occurrence (i.e., Mount El Chichón in 1982 and Mount Pinatubo in 1991). Furthermore, Yamashita et al. (2010) suggested that decadal SST variability was also possibly affecting the solar term. To exclude the spurious effects of volcanic eruptions and interannual SST variability, we performed the CCM experiments without these effects (CNTL experiment).

In the CNTL experiment, the volcanic aerosol was fixed to the background aerosol conditions in 2000. The SST and sea ice were fixed to the calendar month climatology averaged for the 1960–2006 SST and sea ice dataset provided by the UK Met Office Hadley Centre (Rayner et al. 2003). Thus, the interannual variations included in the CNTL experiment were the 11-year solar cycle, QBO, greenhouse gases (GHGs), and ozone-depleting substances (ODSs) such as halogen gases. The GHGs and ODSs were given by scenario A1B of the Intergovernmental Panel on Climate Change (2001) Special Report on Emissions Scenarios and the World Meteorological Organization (WMO)-adjusted scenario A1 (WMO 2007), respectively.

2.3 Analysis method

The QBO and solar cycle influences on the NH winter circulation were estimated by a composite analysis with the following four categories: QBO-W/S_max, QBO-W during S_min (QBO-W/S_min), QBO-E during S_max (QBO-E/S_max), and QBO-E during S_min (QBO-E/S_min). QBO-W and QBO-E are defined as the direction of the zonal-mean zonal wind at 50 hPa averaged over 10°S–10°N for the NH winter (December–January–February; DJF), as used by Yamashita et al. (2011). The definitions of QBO-W and QBO-E winters derived from ERA-Interim for the period 1979–2006 are shown in Fig. 1a, and those derived from the CNTL experiment are shown in Fig. 1b. Although Fig. 1b shows the result of one ensemble run, similar results were obtained from the two other runs because the model zonal wind was nudged...
toward the same observational dataset at 10°S–10°N and 50 hPa. $S_{\text{max}}$ and $S_{\text{min}}$ are defined as the higher and lower values of the DJF mean $F_{10.7}$ compared to its mean value averaged for 1960–2006 (see Fig. 1). By these definitions: 6 years are in QBO-W/$S_{\text{max}}$, 7 years are in QBO-W/$S_{\text{min}}$, 7 years are in QBO-E/$S_{\text{max}}$, and 7 years are in QBO-E/$S_{\text{min}}$ for ERA-Interim. For the CNTL experiment, 39 years are in QBO-W/$S_{\text{max}}$, 42 years are in QBO-W/$S_{\text{min}}$, 24 years are in QBO-E/$S_{\text{max}}$, and 33 years are in QBO-E/$S_{\text{min}}$. We tested the sensitivity of other threshold values and found that choosing threshold values of ± 5 m s$^{-1}$ for the QBO index and ± 0.2 for the solar cycle index gave similar results.

The composite anomaly of the QBO-W/$S_{\text{max}}$ winters was calculated by subtracting the climatology from the QBO-W/$S_{\text{max}}$ composite. Considering the small sample size for ERA-Interim, we use a two-sided bootstrap method to determine the statistical significance level of the difference between the QBO-W/$S_{\text{max}}$ composite and climatology, as explained in Appendix (e.g., Fox 2008).

3. Results

3.1 Seasonal evolution of the NH polar vortex intensity

Figures 2a and 2b show the daily mean time-series of the zonal-mean zonal wind for ERA-Interim averaged over 50°N–70°N at 1 and 10 hPa, respec-
tively, which is a measure of the NH polar vortex intensity. All time-series were smoothed by 60 iterations of a 1:2:1 filter to lessen the short time-scale variations. The polar vortex is stronger in December for the QBO-W/S\textsubscript{max} years (solid line) compared to the climatology (broken line) from 10 to 1 hPa. The polar vortex intensity at 1 hPa for the QBO-W/S\textsubscript{max} years decreases rapidly in January, and a comparatively weak vortex is seen in late January, persisting through to March (Fig. 2a). A similar evolution is seen in the polar vortex intensity at 10 hPa, although the vortex weakening is less pronounced and the timing is slightly later (Fig. 2b). Thus, a relatively weak polar vortex from 10 to 1 hPa is evident in February–March. The seasonal evolution of the polar vortex for the QBO-W/S\textsubscript{max} years is similar to that identified in previous studies (e.g., Gray et al. 2004).

3.2 Combined influences of the QBO-W and S\textsubscript{max} on the stratospheric circulation in December

In this subsection, we investigate how the strong polar vortex arises in December in response to QBO-W/S\textsubscript{max} conditions. This is followed by a weak vortex and more SSWs in February–March. For the QBO, Yamashita et al. (2011) suggested dynamically consistent relationships among the zonal wind, temperature, planetary wave drag, and circulation anomalies around the NH polar vortex in response to the northward shift in latitude of the critical line around 10 hPa during QBO-W conditions defined at 50 hPa, as shown in the diagram of Fig. 3a. We do not consider this diagram to be an explanation of the response but simply a description of its manifestation. For the solar cycle, Kodera and Kuroda (2002) suggested that the strong latitudinal temperature gradient induced by S\textsubscript{max} leads to a stronger westerly wind in the mid-latitude upper stratosphere, with influence on the zonal wind anomalies in the NH high latitudes. We speculate that this process involves wave–mean flow interaction around the polar vortex in the stratosphere, as shown in Fig. 3b. The NH high-latitude anomalies of the solar maximum have the same signs as the QBO-W anomalies in early winter (Figs. 3a, b), and both processes can therefore work in concert during QBO-W/S\textsubscript{max}. This can, in principle, provide a mechanism to strengthen the stratospheric polar vortex during QBO-W/S\textsubscript{max}. Our analysis is limited up to the upper stratosphere due to the upper limit of the reanalysis data and the coarse vertical resolution in the mesosphere of our model, although Matthes et al. (2013) suggested that the QBO initially influences the anomaly of the solar cycle in the subtropical lower mesosphere.

To test the proposed mechanism, we analyzed the QBO-W/S\textsubscript{max} anomalies in the ERA-Interim dataset. Figure 4a shows the latitude–height section of the QBO-W/S\textsubscript{max} anomaly of the Eliassen–Palm (E-P) flux and its divergence. There is suppression of planetary wave propagation entering the polar vortex during QBO-W/S\textsubscript{max} conditions compared to the climatology, which is seen in the mid- and high-latitude stratosphere as a downward E-P flux anomaly. Accordingly, convergence of E-P flux in the stratosphere is suppressed, as shown in the divergence anomaly. This anomalous divergence drives an upward residual velocity anomaly in the polar region and a downward residual velocity anomaly in mid-latitudes (Fig. 4b) according to the downward control principle (Haynes et al. 1991). These velocity anomalies are consistent, through adiabatic heating, with
an anomalously low temperature in the polar region and an anomalously high temperature in the mid-latitudes. Furthermore, they are consistent, through the thermal wind relationship, with a strengthened polar vortex during the QBO-W/\textit{S}_\text{max} years (Fig. 4c). These dynamically consistent relationships suggest the maintenance of a self-maintaining system of circulation and polar vortex anomalies in response to QBO-W/\textit{S}_\text{max}. Note that such a diagnostic analysis cannot provide causation as to how the vortex is modulated in the first place (e.g., Dunkerton and Baldwin 1991; Watson and Gray 2014); the above results only describe the consistency of the anomalies as a self-maintaining system.

Although the ERA-Interim anomalies in Figs. 4a–c exhibit reasonably high significance levels, there could potentially be confounding effects from volcanic eruptions or SST variations. Thus, as confirmation, we analyzed the CCM simulations under conditions of no volcanic eruptions and fixed SST. The results from the CCM show similar anomalies of zonal wind, temperature, E-P flux, and residual mean meridional circulation to those seen in ERA-Interim around the stratospheric NH polar vortex in December, whereas the vertical extent of the zonal wind anomaly from the stratosphere to the troposphere is not simulated in the CCM (Figs. 4d–f).

3.3 Relationship between the strong polar vortex in December and the weak polar vortex in February–March

In this subsection, we investigate how the strong polar vortex in December that arises in response to QBO-W/\textit{S}_\text{max} conditions is related to the weak polar vortex found in February–March under those same conditions, which comes with more frequent SSWs. Previous studies (e.g., Kodera et al. 1991; Gray et al. 2004; Matthes et al. 2013) have focused on this seasonal evolution of the stratospheric polar vortex during QBO-W/\textit{S}_\text{max} conditions in the context of the transfer of the QBO-solar signal within the stratosphere/mesosphere, whereas we focus on the role of the troposphere in the evolution of the QBO-solar signal.

The blue shading in Figures 5a and 5b shows a negative high-latitude SLP anomaly, with evidently increasing magnitude of the anomaly around the North Pole in December (Fig. 5a) and January (Fig. 5b) during QBO-W/\textit{S}_\text{max}. The climatological center of the Icelandic Low is seen in the Atlantic sector of the polar region (indicated by the pressure contours in Figs. 5a, b); thus, the increase in magnitude of
Fig. 4. (a) The longitude–height section of composite anomalies for QBO-W/S max from the December mean E-P flux (vector) and its divergence (shading, units: m s$^{-1}$ d$^{-1}$) from ERA-Interim. (b) The anomalies of the residual mean meridional circulation (vector) and temperature (shading, units: K). The E-P flux vector is scaled by density of atmosphere and latitude as $F_y = \rho \cos\phi \left( \frac{\nabla \theta'}{\partial z} - \overline{u'v'} \right)$, $F_z = \rho \cos\phi \left( \frac{\nabla \theta'}{\partial z} - \overline{u'w'} \right)$, where $\hat{f} = f - (\cos\phi)^{-1}(\overline{\cos\phi}) \hat{\phi}$. The vertical components of the E-P flux and the residual mean meridional circulation are magnified 200 times relative to the horizontal component, and their scales for the horizontal vector are shown at the bottom right of the panel in units of kg m$^{-1}$ s$^{-2}$ and m s$^{-1}$, respectively. A nine-point smoothing was applied to the gridded data of E-P flux, its divergence, and the residual mean meridional circulation. (c) The anomalies of the zonal-mean zonal wind (shading, units: m s$^{-1}$) with the zonal wind climate (gray contour, interval is 5 m s$^{-1}$). The hatched areas of light gray, dark gray, light black, and dark black respectively denote 80 %, 85 %, 90 % and 95 % statistical significance for the anomalies in the E-P flux divergence, temperature, and zonal wind. (d–f) The same as (a–c) but for the CNTL experiment.
Fig. 5. The climatology (solid contour, interval: 4 hPa) and the QBO-W/S anomaly (dotted contour with shading, interval: 2 hPa) for ERA-Interim SLP in (a) December and (b) January. The gray and black stippling show 90 % and 95 % statistical significance, respectively. (c, d) The same as (a, b) but for the 500 hPa geopotential height (Z500). The contour interval for the climatology is 50 m, and that for the anomaly is 20 m.
in contrast to the strengthenings found in December during QBO-W/S\textsubscript{max}, is induced in a model forced only by the anomalous stratospheric zonal-mean flow. They further showed the NAO-like response in the troposphere. Thus, the zonally asymmetric NAO-like response is characteristic of the tropospheric circulation response to an anomalous stratospheric vortex, with stronger vortices inducing a positive NAO anomaly, as seen here.

Figure 7a shows the climatological and anomalous zonal wavenumber 1 (WN1) Z500 field. The climatological ridge with zonal WN1 is located in the mid-latitude North Atlantic. The longitudes of the climatological and anomalous ridges overlap each other, especially around 50°N, implying the constructive interference previously reported by an observational study by Garfinkel et al. (2010) and a modeling study by Smith et al. (2010). Because climatological WN1 usually propagates upward, constructive interference of the anomalous and climatological WN1 waves implies an upward E-P flux anomaly around 50°N, which is indeed seen through the depth of the troposphere into the stratosphere (Fig. 8b). Consequently, anomalous upward propagation is seen in the mid- and high-latitude stratosphere with an enhancement of E-P flux convergence around the polar vortex. The WN1 convergence term accounts for the total convergence term across all wavenumbers (Fig. 8a). Thus, the WN1 convergence term is consistent with the decrease in the polar vortex intensity during January–February, corresponding to the weak polar vortex in February–March (Figs. 2a, b).

These results therefore suggest that the WN1 anomaly developing in the troposphere in January in response to the strengthened early-winter stratospheric vortex during QBO-W/S\textsubscript{max} conditions leads to a weakened polar vortex in February–March through constructive interference between climatological and anomalous WN1, corresponding to an enhancement of the upward propagation of WN1 from the troposphere into the stratosphere. Figure 9 shows the latitude–height section of the QBO-W/S\textsubscript{max} anomalies during December–March as in Figs. 4a–c. The evolution of the zonal wind and temperature anomalies around the stratospheric polar vortex includes the poleward/downward movement with wave propagation/circulation changes and is similar to that of the anomalies shown by the modulation of the PJO in the stratosphere (e.g., Kuroda and Kodera 2002). The novel aspect of our study is the role of the troposphere in this evolution, although Kodera and Kuroda (2002) suggested that the modulation of the PJO may not be related to a change in the tropospheric wave sources.

Figure 7b shows the WN2 component of Z500. The longitude of the anomalous ridge overlaps with that of the climatological trough in the North Atlantic around 50°N, whereas the longitude of the anomalous trough overlaps with that of the climatological ridge around 70°N in the Barents Sea. These relationships are all out of phase, rather than in phase as with WN1. Such destructive interference between climatological and anomalous waves (i.e., a diminution of the amplitude of the climatological waves when their phases are opposite) suppresses the WN2 climatological waves and is manifested as a downward E-P flux anomaly in the troposphere, with anomalous E-P flux divergence in the upper troposphere over 50°N–75°N, 200–500 hPa (Fig. 8c).

The climatological WN2 pattern in Fig. 7b has a northwest–southeast tilt, indicating southward transport of westerly momentum from the polar region within the troposphere. Thus, the weakening of WN2 under QBO-W/S\textsubscript{max} conditions leads to less equatorward transport of westerly momentum. Accordingly, there is a southward E-P flux anomaly with anomalous divergence over 50°N–75°N, 200–500 hPa, as shown in Fig. 8c.

The effect of WN2 waves is quantitatively important relative to the effect of synoptic waves for the zonal wind anomaly during QBO-W/S\textsubscript{max} years, although the synoptic waves are important for the maintenance of the jet position (Lorenz and Hartmann 2003). The E-P flux divergence anomaly corresponds to a westerly wind anomaly in the troposphere, indicating the maintenance of the tropospheric westerly wind anomaly through interaction between the WN2
waves and the mean westerly wind. These results suggest that while the WN2 waves do not play a direct role in forcing the stratospheric vortex evolution, their tropospheric response to QBO-W/S_{max} conditions in early winter appears to be related to the maintenance of the NAO-like westerly wind anomaly in the high-latitude troposphere.

The intensification of the NAO-like response from December to January was not simulated in the CNTL experiment (not shown). Eyring et al. (2010) and Morgenstern et al. (2010a) also reported the lack of the tropospheric response to the strong polar vortex in our model. We suggest that this may be due to the coarse resolution of the model within the troposphere. Interestingly, the model does not show a weakened stratospheric vortex in February–March in relation to the small upward anomaly of E-P flux compared with the ERA-Interim. Thus, the lack of a weakened late-winter stratospheric vortex in the model under QBO-W/S_{max} conditions is consistent with the observational result that the troposphere plays a crucial role in the late-winter stratospheric response.

We analyzed the responses in QBO-E/S_{min} years and found a nearly opposite response to QBO-W/S_{max} years from December to April (not shown). The responses in the other cases (QBO-E/S_{max}, QBO-W/S_{min}) were small.

4. Summary

This study investigated the combined influences of QBO-W and S_{max} conditions on the winter NH extratropical circulation by composite analysis using the ERA-Interim reanalysis and model outputs from the CCSR/NIES CCM experiments. The composite analyses for the ERA-Interim data indicated a strengthened stratospheric polar vortex in December followed by a weakened polar vortex with more frequent SSWs in February–March in response to QBO-W/S_{max} conditions. The latter relationship need not be specific to QBO-W/S_{max} conditions, but may just require a strengthened early-winter vortex, which is more likely in QBO-W/S_{max}. Note that the responses in QBO-E/S_{min} years were found to be nearly opposite to the responses in QBO-W/S_{max} years.

The mechanism underlying the combined influence of QBO-W and S_{max} conditions on the NH polar vortex in December was explained as follows: QBO-W and S_{max} produce similar anomalies in zonal wind, temperature, E-P flux, and residual mean meridional circulation in the NH mid- and high-latitude stratosphere in December. Hence, both anomalies act in concert to induce a stronger upper stratospheric polar vortex in QBO-W/S_{max} compared to climatology. The analysis of ERA-Interim data provided evidence of these anomalies in December.
with dynamically consistent relationships. Similar results were derived from CCM simulations without the potentially confounding effects from volcanic eruptions or SST variability, further establishing the validity of the combined influence of QBO-W/Smax in early winter.

To investigate the relationship between the strengthened polar vortex in December followed by the weakened polar vortex in February–March, we used the ERA-Interim dataset, as the CCM does not show a weakened vortex in February–March. Under QBO-W/Smax, north–south dipole anomalies similar to the NAO developed from December to January. This is consistent with the development of negative NAO anomalies following SSWs, which Hitchcock and Simpson (2014) showed had a similar signal arising from a purely stratospheric, zonally symmetric perturbation. The dipole-like anomalies seen under QBO-W/Smax have a zonal WN1 component, with the locations of the climatological ridge and anomalous ridge overlapping each other. This implies constructive interference and an amplification of zonal WN1 waves, leading to an enhanced upward propagation of zonal WN1 from the troposphere to the stratosphere. This corresponds to the weakened polar vortex in February–March. Taken together, this mechanism can account for the rather complex, opposite-signed response of the polar vortex to QBO-W/Smax between early and late winter. The opposite-signed response is similar to the evolution of the PJO in the stratosphere (e.g., Kuroda and Kodera 2002). The novel contribution of our study is elucidation of the role of the troposphere in this evolution. The failure of the CCM to produce the observed late-winter response to QBO-W/Smax is plausibly explained by its inability to produce the observed tropospheric response to the early-winter vortex strengthening, which is attributed to its coarse tropospheric resolution.

We also found that WN2 waves in the troposphere, whose anomalies interfered destructively with the climatological waves (in contrast to WN1) and thus were not involved in the weakening of the stratospheric vortex, were responsible for anomalous westerly momentum deposition in the high-latitude troposphere where the strong westerly wind anomaly related to the NAO is seen. Thus, WN2 appears to

**ERA-interim QBO-W/S\textsubscript{max} anomaly of E-P flux**

(a) Jan. All  
(b) Jan. WN1  
(c) Jan. WN2  
(d) Jan. U

Fig. 8. (a) The same as Fig. 4a but for January. (b–c) The same as (a) but for the (b) WN1 and (c) WN2 components. (d) The same as (a) but for the anomalies of the zonal-mean zonal wind.
ERA-interim QBO-W/S$_{\text{max}}$ anomaly in Dec.

(a) ERA-interim DEC E-P flux anom.
(b) ERA-interim DEC (v, w), T anom.
(c) ERA-interim DEC Zonal Wind anom.

ERA-interim QBO-W/S$_{\text{max}}$ anomaly in Jan.

(d) ERA-interim JAN E-P flux anom.
(e) ERA-interim JAN (v, w), T anom.
(f) ERA-interim JAN Zonal Wind anom.

ERA-interim QBO-W/S$_{\text{max}}$ anomaly in Feb.

(g) ERA-interim FEB E-P flux anom.
(h) ERA-interim FEB (v, w), T anom.
(i) ERA-interim FEB Zonal Wind anom.

ERA-interim QBO-W/S$_{\text{max}}$ anomaly in Mar.

(j) ERA-interim MAR E-P flux anom.
(k) ERA-interim MAR (v, w), T anom.
(l) ERA-interim MAR Zonal Wind anom.

Fig. 9. (a–c) The same as Figs. 4a–c. (d–l) The same as (a–c) but for (d–f) January, (g–i) February, and (j–l) March.
play a crucial role in the tropospheric response to QBO-W/Smax.

Thus, we suggest that the strengthened early-winter stratospheric polar vortex that arises in response to QBO-W/Smax conditions influences both the development of the NAO in January and the weakened stratospheric polar vortex in February–March through wave–mean flow interaction across the troposphere and stratosphere. These results may clarify the seasonal evolution of polar vortex anomalies in NH winter during QBO-W/Smax conditions and the role of the troposphere in this evolution.

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Appendix: Estimation of statistical significance with bootstrap method

The surrogate QBO-W/Smax years and the surrogate climatological years were chosen randomly from the entire analysis period, and a surrogate composite anomaly, defined as the difference between the surrogate QBO-W/Smax composite and the surrogate climatology, was calculated. This procedure was repeated 1000 times to derive the probability distribution function of the surrogate composite anomaly. The significance level of the difference between the QBO-W/Smax composite and the climatology was determined with (100 − α)% for the magnitude of the anomaly that would be expected to occur α% of the time under the null hypothesis that the difference between the QBO-W/Smax composite and climatology is zero.

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