Influence of the solar cycle on the Polar-night Jet Oscillation in the Southern Hemisphere

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Abstract The Polar-night Jet Oscillation (PJO) is the dominant mode of stratospheric variability in the Southern Hemisphere and persists from midwinter to spring. The influence of the 11 year solar cycle on modulation of the PJO from late winter to spring is examined using observations and three 42 year simulations from a chemistry-climate model. The only variation applied to model boundary conditions was the strength of ultraviolet (UV) radiation. This is set at 2 times larger than observations to enhance the strength of the solar signal. Simulations show a downward propagation of the stratospheric signal into the troposphere from late winter to spring, which tends to be enhanced as UV strength increases. This result is similar to observations but with a 1–2 month lag. The behavior of the PJO with respect to wave-mean flow interactions is examined using a newly developed momentum budget analysis as well as wave energy analysis. We suggest that UV modulation of the interactions between planetary waves and zonal-mean flow in the stratosphere, rather than direct diabatic processes as suggested in a previous study, is the source of solar cycle modulation of the PJO.

1. Introduction

There is increasing evidence that the 11 year solar cycle plays an important role in climate variability [e.g., Gray et al., 2010], including via the modulation of annular modes [Kodera, 2002; Ogi et al., 2003; Kuroda and Kodera, 2005], which are the dominant modes of high-latitude variability in both hemispheres [Thompson and Wallace, 1998; Limpasuvan and Hartmann, 1999, 2000]. A long-lived mode of variability, the Polar-night Jet Oscillation (PJO), occurs in the Southern Hemisphere (SH) stratosphere from winter to spring [Kuroda and Kodera, 1998, 2001]. Previous studies have shown that the Southern Annular Mode (SAM) is strongly influenced by the PJO from October to December, when stratosphere-troposphere coupling is strong [Kuroda, 2002]. Considering the close relationship between the SAM and the PJO, and between solar activity and the PJO [Kodera and Kuroda, 2002; Kuroda and Kodera, 2002], a possible solar cycle modulation of the SAM could be closely related to the PJO. In previous studies, it was shown that the Meteorological Research Institute-Chemistry-Climate Model (MRI-CCM) can reproduce solar cycle modulation of stratosphere-troposphere coupling in spring in the SH [Kuroda and Shibata, 2006; Kuroda et al., 2007]. Thus, the model also has the potential to reproduce the solar cycle modulation of the PJO. Because solar cycle-SAM interactions are highly stochastic, a longer model experiment would be useful in helping draw more robust conclusions. In addition, although the PJO and SAM are created by different types of waves and operate on very different timescales, they both originate from wave-mean flow interactions [Limpasuvan and Hartmann, 2000; Kuroda and Kodera, 2001; Kuroda and Mukougawa, 2011]. Therefore, analyses of wave-mean flow interactions are useful in understanding solar modulation of the PJO and SAM. Analytical tools developed in previous studies for examining momentum and energy budgets between waves and the zonal field [Kuroda and Mukougawa, 2011, 2013] can also be applied to provide insight into the solar modulation of the PJO and SAM. For example, our momentum analysis tool can accurately diagnose the effect of area-specific waves on zonal wind acceleration, including the Coriolis force, which was insufficiently handled by traditional Eulerian or transformed Eulerian mean momentum analysis.

The purpose of the present study is to examine the effect of the solar cycle on the relationship between the PJO and the SAM. Extended chemistry-climate model simulations and new analytical tools are used to investigate wave-mean flow interactions to help identify the key processes underlying solar modulation of the PJO and SAM.

The remainder of this paper is organized as follows. Section 2 describes the data and principal analytical methods. After the presentation of the results in section 3, including the momentum and energy analysis, section 4 provides a discussion and concluding remarks.
2. Data and Analysis Methods

2.1. Data and Statistical Methods

Observations used in the study are from the European Centre for Medium-range Weather Forecasts (ECMWF) 35 year reanalysis data set (ERA-Interim) from 1 January 1979 to 31 December 2013 [Dee et al., 2011]. Data are output on a 1.5° × 1.5° longitude-latitude grid with 37 vertical pressure levels from 1000 to 1 hPa. Our analysis is based on monthly mean data, but the Eliassen-Palm (E-P) flux was first calculated at daily intervals and then averaged monthly.

The SAM is obtained from the 850 hPa geopotential height south of 20°S as the first month-to-month empirical orthogonal function mode [Thompson and Wallace, 1998]. The PJO is extracted as a dominant mode of year-to-year multiple singular value decomposition (SVD) analysis [e.g., Wallace et al., 1992] between the normalized zonal-mean zonal wind and the vertical component of the E-P flux as follows. Two vectors, \( X_i \) and \( Y_i \), are defined as

\[
X_i = (U_i(l), U_i(l + 1) \ldots U_i(l + T - 1)), \\
Y_i = (E_i(l), E_i(l + 1) \ldots E_i(l + T - 1)),
\]

and then the SVD between these vectors is calculated under the assumption that data from each year are independent. Here \( U_i(l) \) \( (E_i(l)) \) is the standardized anomalous mean monthly zonal-mean zonal wind (vertical component of the E-P flux) at each height and latitude for the \( l \)th month of the \( l \)th year. The PJO index is defined as the standardized time series of the vertical component of the E-P flux associated with the leading mode \( (b) \). The pattern vector (called the singular vector in SVD terminology) of the zonal wind is proportional to the regression of zonal wind data \( X_i \) against the time series of the E-P flux \((i.e., < bX_i > \) where the bracket indicates time averaging). Similarly, the pattern vector of the E-P flux is proportional to \( < aY_i > \) if the standardized time series of zonal wind is defined by \( a \). Thus, pattern vectors are sometimes termed heterogeneous regressions [e.g., Wallace et al., 1992]. As the two time series of \( a \) and \( b \) are commonly highly correlated, heterogeneous regression tends to be similar to a simple regression against the PJO index. However, following the conventions of SVD analysis, this paper presents basic heterogeneous regression maps. The detailed theory and calculation methods of SVD analysis can be found in Wallace et al. [1992] and Kuroda [1998].

The chemistry-climate model (CCM) used in this study was developed by the Meteorological Research Institute (MRI) [Shibata et al., 2005] and was used by Kuroda and Shibata [2006] and Kuroda et al. [2007]. We integrated the model over 42 years with seasonally varying climatological sea surface temperatures and a perpetual solar condition. Three runs were performed under the same initial and boundary conditions, except for the strength of the solar ultraviolet (UV) radiation, which was obtained from a spectral table [Lean et al., 2005]. The three simulations were as per Kuroda et al. [2007], comprising stronger (solar maximum (SMX)), normal (solar normal (SN)), and weaker (solar minimum (SMN)) UV strengths. The SN UV strength was determined from the average of the solar maximum month (November 1989) and the solar minimum month (September 1986). The SMX UV strength was obtained from \( S_0(1 + d) \), where \( S_0 \) is the SN UV strength and \( d \) is the ratio of the increase in UV strength at the solar maximum compared with the solar minimum. Note that both \( S_0 \) and \( d \) depend on wavelength. Similarly, the SMN UV strength was obtained from \( S_0(1 - d) \). As a result, the ratio of UV strengths between SMX and SMN is about twice as large as the observed 11 year solar variability. We used a larger UV variation in the numerical experiments to enhance the effect of UV on climate.

Simulation runs covering the first 21 years were the same as those described by Kuroda et al. [2007]. We continued the integration over an additional 21 years to achieve higher statistical significance and performed the statistical analysis in the present study using data from 41 winter-to-summer seasons.

Three-dimensional daily data are used as the basis for analysis of simulation results. Although most of the analysis is based on monthly means, second-order or higher-order quantities (e.g., Eulerian wave forcings) are calculated first at daily intervals and then averaged monthly. Frictional forcing (diabatic heating) is calculated as residuals of the three-dimensional momentum (thermodynamic) equation using daily data.
2.2. Momentum Equations

Our momentum diagnosis is based on the zonal-mean primitive equations as follows:

\[
\frac{\partial \bar{\mathbf{u}}}{\partial t} = 2\Omega \sin \phi + \vec{F}_e + \bar{F}_n + \vec{X},
\]

(2a)

\[
2\Omega \sin \phi + \frac{1}{a} \frac{\partial \bar{\mathbf{u}}}{\partial \phi} = J,
\]

(2b)

\[
\frac{\partial \bar{T}}{\partial t} = -\frac{RT}{p},
\]

(2c)

\[
\frac{\partial \bar{\mathbf{v}}}{\partial t} = \Gamma + \vec{Q}_e + \bar{Q}_n + \vec{S},
\]

(2d)

\[
\frac{1}{a \cos \phi} \frac{\partial}{\partial \phi} (v \cos \phi) + \frac{\partial \bar{\mathbf{v}}}{\partial p} = 0,
\]

(2e)

where overbars represent zonal means and the terms

\[
\bar{F}_e = \frac{1}{a \cos^2 \phi \sin \phi} \left( \frac{\partial}{\partial \phi} (u'v' \cos^2 \phi) - \frac{\partial}{\partial p} (u'v') \right),
\]

(3)

\[
\bar{Q}_e = \frac{1}{a \cos \phi} \left( \frac{\partial}{\partial \phi} (\Gamma' \cos \phi) - \frac{\partial}{\partial p} (\bar{F}_e') + \frac{\kappa}{p} \bar{T}' \right),
\]

(4)

are eddy mechanical forcings and thermal forcings with the prime being the eddy component (departure from zonal mean); \(\vec{X}\) and \(\vec{S}\) are zonal-mean zonal frictional forcing and diabatic heating, respectively; \(\Gamma(p) = -\partial T_0/p + \kappa T_0/p\) is the stability of a reference atmosphere at temperature \(T_0(p)\); \(\Omega\) is the angular velocity of the Earth; \(p\) is pressure; \(\phi\) is latitude; \(a\) is the radius of the Earth; and other terms follow the usual conventions [e.g., Andrews et al., 1987]. The nonlinear advection terms \(\bar{F}_n\) and \(\bar{Q}_n\) are defined as follows:

\[
\bar{F}_n = -\frac{v}{a} \frac{\partial \bar{\mathbf{u}}}{\partial \phi} - \frac{\partial \bar{\mathbf{u}}}{\partial p} + \frac{\partial}{\partial \phi} \kappa \tan \phi,
\]

\[
\bar{Q}_n = -\frac{v}{a} \frac{\partial \bar{T}}{\partial \phi} + \left( \frac{\partial \bar{T}}{\partial p} + \frac{\kappa}{p} \bar{T} \right) \bar{\mathbf{v}}.
\]

(4)

The meridional wind equation has been replaced by a balanced wind relationship with a violation of the balanced wind \(J\) defined by equation (2b). This is directly evaluated from the data. Note that equation (2d) does not depend on \(\Gamma\) if all terms are summed.

Following the calculation steps outlined in Appendix A, equations (2a) to (2e) can be rearranged to give an elliptical differential equation of \(\bar{\mathbf{u}}\), as follows:

\[
\frac{1}{\cos \phi} \frac{\partial}{\partial \phi} \left( \cos \phi \frac{\partial \bar{\mathbf{u}}}{\partial \phi} \right) + \frac{4\Omega^2 a^2 p}{RT} \frac{\partial^2 \bar{\mathbf{u}}}{\partial p^2} = \frac{2\Omega a p}{RT \cos \phi} \frac{\partial}{\partial \phi} \left( \cos \phi \frac{\partial \bar{\mathbf{u}}}{\partial \phi} \left( \bar{F}_e + \bar{F}_n + \vec{X} - \frac{J}{2\Omega \sin \phi} \right) \right) - \frac{1}{\cos \phi} \frac{\partial}{\partial \phi} \bar{\mathbf{u}} \left( \bar{Q}_e + \bar{Q}_n + \vec{S} \right).
\]

(5)

The boundary conditions should be an approximate form of \(\partial \bar{\Phi}/\partial t = 0\) at the 1000 hPa level, \(\bar{\mathbf{u}} = 0\) at the uppermost level, and \(\bar{\mathbf{v}} = 0\) at \(\phi = \pm \pi/2\).

Equation (5) with boundary conditions can be symbolically represented as

\[
\mathbf{L} \bar{\mathbf{u}} = \mathbf{A}_1 \left( \bar{F}_e + \bar{F}_n + \vec{X} \right) + \mathbf{A}_2 \bar{J} + \mathbf{A}_3 \left( \bar{Q}_e + \bar{Q}_n + \vec{S} \right),
\]

(6)

where \(\mathbf{L}\) and \(\mathbf{A}_i (i = 1, 2, 3)\) are linear operators. If \(\bar{\mathbf{v}}\) is represented as \(\bar{\mathbf{v}} = \mathbf{M} \bar{\mathbf{u}}\) with a linear operator by equation (2e), the nonlinear advection terms \(\bar{F}_n\) and \(\bar{Q}_n\) in equation (4) can also be represented by

\[
\bar{F}_n = \mathbf{B} \bar{\mathbf{u}},
\]

\[
\bar{Q}_n = \mathbf{C} \bar{\mathbf{u}}.
\]
Here $\mathbf{B}$ and $\mathbf{C}$ are linear operators represented by

$$
\mathbf{B} = \frac{\partial}{\partial \phi} \tan \phi - \frac{1}{\partial \phi} \frac{\partial \mathbf{M}}{\partial \phi} + \frac{\partial \mathbf{I}}{\partial \phi},
$$

$$
\mathbf{C} = \frac{1}{\partial \phi} \frac{\partial \mathbf{M}}{\partial \phi} + \left( - \frac{\partial \mathbf{M}}{\partial \phi} + \frac{\kappa \mathbf{T}}{p} - \mathbf{I} \right),
$$

with $\mathbf{I}$ as the identity operator. For the evaluation of equation (7), data and their spatial derivatives at the time of evaluation are used for $\mathbf{B}$ and $\mathbf{T}$. Thus, the vertical velocity $\mathbf{W}$ can be represented as

$$
\mathbf{W} = (\mathbf{L} - \mathbf{A}_1 \mathbf{B} - \mathbf{A}_2 \mathbf{C})^{-1} \left( \mathbf{A}_1 (\mathbf{\bar{F}}_y + \mathbf{\bar{X}}) + \mathbf{A}_2 \mathbf{\bar{J}} + \mathbf{A}_3 (\mathbf{\bar{Q}}_y + \mathbf{\bar{J}}) \right).
$$

Similarly, acceleration of zonal wind can be represented from equation (2a) as

$$
\begin{align*}
\frac{\partial \mathbf{U}}{\partial \phi} & = \left( \{ (2 \Omega \sin \phi \mathbf{M} + \mathbf{B})(\mathbf{L} - \mathbf{A}_1 \mathbf{B} - \mathbf{A}_2 \mathbf{C})^{-1} \mathbf{A}_1 + \mathbf{I} \} (\mathbf{\bar{F}}_y + \mathbf{\bar{X}}) \\
+ \{ (2 \Omega \sin \phi \mathbf{M} + \mathbf{B})(\mathbf{L} - \mathbf{A}_1 \mathbf{B} - \mathbf{A}_2 \mathbf{C})^{-1} \} \mathbf{A}_2 \mathbf{\bar{J}} + \mathbf{A}_3 (\mathbf{\bar{Q}}_y + \mathbf{\bar{J}}) \right).
\end{align*}
$$

As equations (8) and (9) are completely linear with respect to $\mathbf{F}_y$, $\mathbf{Q}_y$, $\mathbf{X}$, and $\mathbf{J}$, the meridional circulation and acceleration produced by each forcing are evaluated separately (see Kuroda [2016] for details). Kuroda [2016] also reported that the present analysis method has sufficient accuracy for diagnosing meridional circulation and acceleration of zonal wind.

In the present study, the reference temperature $T_0(p)$ is defined as the average temperature south of 20°S. We evaluate acceleration of zonal wind in each simulation using the diagnostic model. This is run daily with 1001 vertical levels at intervals of 1 hPa and 121 horizontal grids of equal intervals of the sine of the latitude.

### 2.3. Wave Energy Equations

The model used to evaluate energy transfer between zonal-mean fields and atmospheric waves is based on the wave energy equation derived by Holton [1975]. Its integral form is as follows:

$$
\frac{d}{dt} (K' + P') + W = \{ \mathbf{\bar{R}}, K' \} + \{ \mathbf{\bar{P}}, P' \} + D,
$$

where $K'$ and $\mathbf{\bar{R}}$ are kinetic energies of eddies and zonal-mean fields, respectively; $P'$ and $\mathbf{\bar{P}}$ are available potential energies of eddies and zonal-mean fields, respectively; $W$ is the surface integral of the wave energy flux; $\{ \mathbf{\bar{R}}, K' \}$ is the energy conversion from $\mathbf{\bar{R}}$ to $K'$; $\{ \mathbf{\bar{P}}, P' \}$ is that from $\mathbf{\bar{P}}$ to $P'$; and $D$ is the external forcing or dissipation term. The integrands $\epsilon(\mathbf{\bar{R}}, K')$ and $\epsilon(\mathbf{\bar{P}}, P')$ of $\{ \mathbf{\bar{R}}, K' \}$ and $\{ \mathbf{\bar{P}}, P' \}$ (e.g., $\{ \mathbf{\bar{R}}, K' \} = \frac{d}{dt}\chi(\mathbf{\bar{R}}, K')$) are written as

$$
\epsilon(\mathbf{\bar{R}}, K') = -\rho_0 \left( \frac{1}{\nu^2} \frac{\partial^2}{\partial \phi^2} + \frac{1}{\nu^2 \partial z^2} + \frac{1}{\nu^2} \frac{\partial^2}{\partial \phi^2} + \frac{1}{\nu^2 \partial z^2} + \frac{\partial^2 \tan \phi}{\partial \phi^2} - \frac{\partial^2 \tan \phi}{\partial z^2} \right),
$$

and

$$
\epsilon(\mathbf{\bar{P}}, P') = -\rho_0 \frac{R^2}{K^2 H^2} \left( \frac{1}{\nu^2} \frac{\partial^2}{\partial \phi^2} + \frac{\partial^2 \mathbf{T}}{\partial z^2} \right).
$$

The integrand $\epsilon(D)$ of $D$ is written as

$$
\epsilon(D) = \rho_0 \left( \frac{\nabla^2 \mathbf{X'}}{\nu^2} + \frac{\nabla^2 \mathbf{X'}}{\nu^2 \partial z^2} + \frac{R^2 \rho_0}{H^2 K^2} \nabla^2 \mathbf{T} \right).
$$

where $\mathbf{X'}$, $\mathbf{Y'}$, and $\mathbf{S'}$ are the eddy components of zonal and meridional frictions and diabatic heating, respectively. In the present study, we use the differential form of equation (10) to evaluate the energy transfer to waves.

### 2.4. Figures

Figures 1 and 3 show heterogeneous regressions, and Figures 4–7 show regressions using the PJO index. In both cases, it should be noted that figures are relative to a positive phase of the PJO index when a winter polar vortex anomaly extends toward low latitudes in the stratosphere.
3. Results

3.1. Observations

The PJO shows prominent year-to-year variability in the stratosphere. This corresponds with the variability in the extent of the Polar-night Jet in the upper stratosphere at low latitudes [Shiotani et al., 1993] and the persistence of this jet until spring. Statistically, this mode is identified as a dominant multiple SVD mode between standardized quantities of zonal-mean zonal wind and the vertical component of the E-P flux in winter months through equation (1) [Kuroda and Kodera, 2001]. This mode can be identified because it is closely related to continuous stratosphere-troposphere coupling through upward propagating planetary waves. The patterns extracted by multiple SVD analysis from August to September are similar to those obtained by Kuroda and Kodera [1998, 2001]. The patterns are almost identical even when the months used in SVD calculations are changed, e.g., July through October (not shown). Thus, this PJO pattern is an interannual mode of variability with intraseasonal temporal evolution. A winter with a positive PJO index corresponds to an anomalous extension of the polar vortex toward low latitudes in the stratosphere [Kuroda and Kodera, 2001]. Temporal evolution of the PJO is also characterized by a slow poleward and downward propagation of the zonal-mean zonal wind anomaly with a corresponding change in the E-P flux and coupling to the troposphere in late winter to spring (October to December) [Kuroda and Kodera, 1998, 2001].

To examine the effect of solar activity on the PJO, we divided the time series according to the strength of the October–December mean $F_{10.7}$ index, which is an indicator of solar activity. High solar (HS) years are defined as those with mean $F_{10.7}$ indices greater than 130. Similarly, low solar (LS) years are those with indices lower than 100. By these definitions, we extracted 17 HS and 17 LS years from a total of 35 years. Figure 1 compares the seasonal evolutions of the zonal-mean zonal wind and the E-P flux shown by heterogeneous regression maps in HS and LS years (i.e., $< b_i^{HS}X_i^{HS} >$ for the zonal wind and $< a_i^{HS}Y_i^{HS} >$ for the vertical component of the E-P flux in case of HS years, where $a_i^{HS}$ ($b_i^{HS}$) is the standardized time series of zonal wind (vertical component of the E-P flux) for HS years, etc.). The meridional component of the E-P flux was obtained by regressing this variable against the PJO index.

Figure 1. Lagged heterogeneous regressions of the observed zonal-mean zonal wind (contours) and E-P flux (arrows) from August to October. Data were obtained using the multiple SVD analysis time series between August and September for normalized zonal-mean zonal wind and the vertical component of the E-P flux. The meridional component of the E-P flux was obtained using regression. Years corresponding to (a–c) HS and (d–f) LS activities, respectively. The contour interval is 2 m s$^{-1}$, and dashed lines indicate negative values. Shading identifies areas that are significant at 95% confidence (Student’s t test). The E-P flux is scaled by the reciprocal square root of pressure, and arrows show fluxes that are significant at 95% confidence.
Until August, the anomalous zonal wind shows a meridional dipole structure with an anomalous downward E-P flux in the stratosphere in both HS and LS years (Figures 1a and 1d). In September, however, zonal wind anomalies begin to propagate downward in both types of solar year, but the signal is stronger in HS years (Figures 1b and 1e). In October, the stratospheric signal extends downward and clearly connects to the surface SAM in HS years, but the extension is very weak in LS years (Figures 1c and 1f). In fact, the correlation coefficient between the PJO and October SAM index is 0.80 for HS years but only 0.35 for LS years. The highest correlation for LS years is in November (0.45). Note that the overall result is not sensitive to a small change in the threshold used to select HS and LS years. For example, a near-identical pattern was obtained even if HS was defined by years of mean $F_{10.7}$ larger than 150 (not shown).

In HS years, the PJO in late winter tends to propagate faster and is more strongly coupled to the tropospheric SAM, as compared to LS years. However, the present analysis is based on only 17 years of data, and observational data may include other complicating factors such as volcanic effects and oceanic conditions that can influence stratospheric variability.

### 3.2. Simulations

We analyzed 42 year simulations from the MRI-CCM to investigate the effects of solar activity on the coupling of the PJO to the SAM in late winter to spring. Kuroda et al. (2007) previously demonstrated that the model simulates solar effects on stratosphere-troposphere coupling in late winter/spring. Because simulations have no external forcings, except for the UV strength, they have potential to isolate solar effects on PJO-SAM coupling.

First, we examine climate in the simulations. Figure 2 depicts the seasonal march of the 42 year climatological zonal-mean zonal wind and the E-P flux (Figures 2a–2c) and temperature (Figures 2d–2f) for the SN simulation, as well as the temperature and the zonal wind difference between SMX and SMN simulations (Figures 2g–2i and 2j–2l, respectively) from October to December. As zonal wind and temperature are similar in all simulations, only those in the SN run are shown. Although the difference in zonal wind between the SMX and SMN simulations is not significant at the 95% confidence level (Student's $t$ test; Figures 2j–2l), there are some similarities with observations [Kuroda and Kodera, 2002] (Figure 6). The difference in temperature (and ozone) in the stratosphere is highly significant (Figures 2g–2i). The seasonal evolution of ozone and its difference between simulations are similar to those in Figure 1 of Kuroda and Shibata [2006]. However, a comparison of the seasonal march of zonal wind observations reveals that modeled climate lags observations by about 1 month due to model bias [Kuroda et al., 2007]. In association, the PJO tends to propagate into the troposphere from October to November in observations but around December in model simulations. Such a bias is a common problem in many chemistry-climate models [Butchart et al., 2011].

SVD analysis using the standardized zonal-mean zonal wind and the vertical component of the E-P flux in equation (1) is used to identify the PJO as the dominant mode of variability. The use of multiple months, such as July to October, reflects the characteristics of the PJO. However, for SVD analysis in this study, a single month (October) is selected for simplicity. Even using a single month in the analysis, the poleward and downward propagating characteristics of the PJO are captured effectively in all runs. To identify the PJO, the same analysis is performed in all runs.

Figure 3 compares responses to a positive PJO phase. It shows the temporal evolution of zonal-mean zonal wind and the E-P flux, as heterogeneous regression maps from October to December for all three simulations. The meridional component of the E-P flux was obtained using regression analysis. In October, the spatial pattern of the PJO is similar for all runs; i.e., anomalous positive zonal-mean zonal wind equatorward of $60–70^\circ$S at the stratosphere begins to shift poleward with reduced upward propagation of waves (Figures 3a, 3d, and 3g). However, the temporal evolution of the PJO in later months is different between runs. In fact, for the SMX run the anomalous positive zonal wind signal in the stratosphere begins to extend into the troposphere in November (Figure 3b) and strongly couples with the surface in December (Figure 3c). For the SMN run, the extension of zonal wind to the troposphere and surface is much weaker (Figures 3h and 3i). The SN run has characteristics that fall between those of the SMX and SMN runs (Figures 3e and 3f). The correlation coefficients for PJO and December SAM indices are 0.54 for the SMX run compared with only 0.08 for the SMN run. Thus, our results indicate that downward propagation of the PJO tends to be more pronounced with increasing solar activity, which is consistent with observations, although the magnitude of solar forcing imposed in these simulations is idealized.
3.3. Momentum and Energy Analysis

Because acceleration of zonal wind is created mainly through a change in wave forcings, we examined the difference in wave forcings between runs to determine the origin of different stratospheric variability (PJO) responses to solar activity. There are two types of Eulerian wave forcings: mechanical and thermal ones, as Figure 2.

Figure 2. The 42 year climatology of (a–c) mean zonal wind, (d–f) temperature in the solar normal (SN) simulation, and the difference in (g–i) temperature and (j–k) zonal wind between solar maximum (SMX) and solar minimum (SMN) simulations from October to December. The contour intervals are 10 m s\(^{-1}\) (Figures 2a–2c), 10 K (Figures 2d–2f), 1 K (Figures 2g–2i), and 1 m s\(^{-1}\) (Figures 2j–2l), and dashed lines represent negative values. The thin solid line indicates zero values. Arrows in Figures 2a–2c show the E-P flux scaled by the reciprocal square root of pressure. Shading indicates year-to-year variability (Figures 2a–2c and 2d–2f) or statistical significance according to Student’s t test (Figures 2g–2i and 2j–2k). The light and heavy shading indicate regions where the standard deviation is greater than 5 and 10 m s\(^{-1}\) for Figures 2a–2c, and 5 K for Figures 2d–2f, and 2 (significant at the 95% level) and 4 (significant at the 99.95% level) for Figures 2g–2i and 2j–2k, respectively.
defined in equation (3). The former is a direct source of acceleration through equation (2a). On the other hand, thermal forcing controls acceleration only through a change in meridional circulation. In this sense, mechanical forcing is more closely related to the acceleration of zonal wind. It is also identical to the meridional divergence of the E-P flux for the quasi-geostrophic limit [Andrews et al., 1987]. However, it should be noted that during formation of the SAM, contributions from “medium-scale waves” with a timescale of less than 2 days account for almost one third of total forcings and cannot be ignored [Kuroda and Mukougawa, 2011]. Medium-scale waves have spatiotemporal characteristics that fall between those of synoptic-scale waves and mesoscale disturbances [Sato et al., 1993]. Although we use daily data as the basis for our analysis, which exclude wave forcings from medium-scale waves, their mechanical forcing is included in a residual of the momentum equation. By comparison of mechanical forcing calculated from daily and 6-hourly data from the reanalysis, we included residual momentum data from the 850 to 70 hPa levels in a mechanical forcing. In this way, all wave forcing effects were calculated.

Figure 4 compares regressions of mechanical forcing ($F_e$ in equation (3)) and the E-P flux against the PJO index. Although forcing patterns are similar among the October simulations (Figures 4a, 4d, and 4g), the features differ in November. In particular, positive centers in the troposphere around 55°S and 300 hPa are most
prominent in the SMX run in both November and December (Figures 4b and 4c). Such signals can also be observed in the SN run in December (Figure 4f) and the SMN run in November (Figure 4h), but they are neither prominent nor persistent. A comparison between runs indicates that if UV forcing is enhanced, the positive stratospheric signal associated with meridional divergence of the E-P flux around the subtropical stratopause in October tends to move downward and splits into two signals in November (Figures 4b, 4e, and 4h). This is closely related to the downward propagation of zonal wind anomalies in the troposphere for the run with stronger UV forcing (Figure 3).

Although mechanical forcing is a primary driver of waves that control zonal-mean zonal wind (see equation (2a)), it does not indicate an actual acceleration due to the existence of meridional circulation driven by both types of wave forcings. In fact, meridional circulation is also produced by wave forcings through equations (5) and (2e), and Coriolis forcing of such circulation tends to largely offset direct acceleration produced by mechanical forcing. As described in section 2.2, the acceleration of zonal wind can be decomposed into a linear combination of acceleration due to each forcing. Moreover, our method enables us to precisely evaluate acceleration due to the respective forcings and to identify the acceleration of zonal wind due to waves alone.

Figure 5 shows the regression of acceleration caused by waves alone (i.e., \((2\Omega \sin \phi M + B)(L - A_1B - A_2C)^{-1} (A_1F_e + A_2\Omega_e) + F_e\) from equation (9)). Although accelerations are similar for the three runs in October and

Figure 4. Lagged regressions of Eulerian mechanical forcing due to all waves (contours) from October to December for chemistry-climate model simulations. Arrows show the E-P flux as in Figure 3. The contour interval is 0.2 m s\(^{-1}\) d\(^{-1}\), and dashed lines indicate negative values. Heavy and light shading indicate areas of statistical significance at the 99.5% and 95% levels, respectively (Student's t test). The E-P flux is scaled by the reciprocal square root of pressure, and arrows show fluxes that are significant at 95%.
(Figures 5a, 5d, and 5g), their features differ in November. In fact, a deep meridional dipole acceleration structure extends to the surface in the SMX run (Figure 5b); the meridional dipole structure is limited to the stratosphere in the SN run (Figure 5e). Conversely, the tropospheric signal almost disappears in the SN and SMN runs (Figures 5e and 5h). The meridional dipole structure that extends to the middle stratosphere is persistent in December in the SMX run (Figure 5c), acceleration almost disappears in the SN run (Figure 5f), and deceleration is more prominent in the SMN run (Figure 5i).

Figure 5 shows acceleration produced by waves alone. As this acceleration is in large part offset by near-surface frictional forcing (in the planetary boundary layer), only a small residual fraction (~1/5 near the surface) produces an actual zonal wind tendency (not shown). However, frictional forcing works passively against zonal wind or its tendency near the surface. Thus, acceleration produced by waves alone is a good qualitative indicator of the actual change in zonal wind.

Our analysis shows that differences in the downward propagation of the PJO due to solar activity arise from differences in the strength of wave forcings and not from diabatic heating or gravity waves represented by residuals of the thermodynamic and momentum equation, respectively. However, it is unclear why solar activity affects the strength of wave forcings. Generally speaking, a change in the strength of wave forcings is associated with the transfer of wave energy.

To further clarify the mechanisms underlying the effect of solar activity on the strength of wave forcings, we performed a wave energy budget analysis similar to that in Kuroda and Mukougawa [2011, 2013]. Since the

Figure 5. Lagged regressions of zonal wind acceleration due to all waves from October to December for chemistry-climate model simulations. The contour interval is 0.1 m s$^{-1}$ d$^{-1}$, and dashed lines indicate negative values. Heavy and light shading indicate areas of statistical significance at the 99.5% and 95% levels, respectively (Student’s t test).
effects of solar activity are imposed in model simulations by increasing UV energy input, it is logical that different wave forcing strengths may originate from external forcings associated with diabatic heating. Thus, we calculated the second term of equation (13), which represents energy inputs from the eddy component of diabatic heating (as heating by ozone waves). Climate analysis shows positive energy inputs below 200 hPa and negative inputs above that level in all months. In the stratosphere, a strong negative signal centered around 50°S and 200 hPa tends to extend to 60°S and 10 hPa from October to December. The difference between the SMX and SMN runs is insignificant (at 95% confidence), and a meridional dipole structure with a peaked negative signal ($-3 \times 10^{-7} \text{ W m}^{-3}$ around 60°S and 10 hPa) occurs at high latitudes in the stratosphere in October and persists until November (not shown).

Regressions of energy transfer against PJO indices are compared in Figure 6. The energy transfer from diabatic heating to waves in the stratosphere is about two times larger in the SN and SMN runs than in the SMX run, in both October (Figures 6a, 6d, and 6g) and November (Figures 6b, 6e, and 6h). For example, in November the peak energy transfer at around 65°S and 50 hPa is about $1 \times 10^{-6} \text{ W m}^{-3}$ for the SMX run (Figure 6b) but about $2 \times 10^{-6} \text{ W m}^{-3}$ for the SN and SMN runs (Figures 6e and 6h). Analysis using short- and long-wave radiative heating rates shows that the present result is similar to that of the sum of these and originates mainly from long-wave heating (not shown).

For the source term of the wave energy equation (RHS of equation (10)), the largest contribution comes from transfer terms from the zonal-mean field to waves (the sum of the first and second terms in equation (10)).
These terms are about $1 \times 10^{-4}$ W m$^{-3}$ in the lower stratosphere and $5 \times 10^{-4}$ W m$^{-3}$ in the lower troposphere in all runs.

Regressions of energy transfer from zonal fields to waves are compared in Figure 7. In October, the spatial patterns of energy transfer are similar between runs (Figures 7a, 7d, and 7g). However, in November the spatial patterns begin to differ and energy transfer in the stratosphere tends to become larger with increasing solar activity (Figures 7b, 7e, and 7h). In fact, energy transfer around 60°S and 30 hPa is $1 \times 10^{-5}$ W m$^{-3}$ in the SMX run (Figure 7b) but only $0.5 \times 10^{-5}$ W m$^{-3}$ in the SMN run (Figure 7h). The area of positive energy transfer propagates farther into the troposphere with a significant value of $1.5 \times 10^{-5}$ W m$^{-3}$ in the SMX run in December (Figure 7c) but an insignificant value of $1 \times 10^{-5}$ W m$^{-3}$ in the SMN run (Figure 7i). It should be noted that wave generation is associated with positive energy transfer. However, as planetary waves have westward momentum, the conservation of momentum requires that zonal fields have eastward momentum. This explains the strong correspondence between energy transfer and zonal wind acceleration. The formation of meridional dipole structures of these quantities and zonal wind (Figure 3c) corresponds with the SAM and is sustained through eddy-feedback processes [Lorenz and Hartmann, 2001; Kuroda and Mukougawa, 2011].

We suggest the following possible scenario regarding the solar influence on the downward propagation of the PJO. The PJO in the SH is a dominant mode of year-to-year variability in the stratosphere and is created by dynamical interaction between the zonal-mean field and planetary waves [Kuroda and Kodera, 1998, 2001]. During the season when the polar vortex breaks down, solar radiation begins to penetrate

**Figure 7.** Lagged regressions of total wave energy transfer from zonal-mean fields to waves from October to December for chemistry-climate model simulations. The contour interval is $5 \times 10^{-6}$ W m$^{-3}$, and dashed lines indicate negative values. Heavy and light shading indicate areas of statistical significance at the 99.5% and 95% levels, respectively (Student’s t test).
the polar stratosphere. The temperature, stability, and ozone density of the stratosphere tend to increase with the increase in UV. With such changes in the stratosphere, the effect of incoming waves on wave forcings in the stratosphere tends to be amplified by enhanced solar radiation, although the mechanisms underlying this phenomenon are not clear. As a result, the interaction between planetary waves and zonal-mean flow is enhanced in the stratosphere and the PJO tends to propagate farther into the troposphere. This could create the tropospheric SAM, which has the same polarity as the PJO when solar conditions are enhanced.

Thus, the key factor in the solar-PJO relationship is that wave forcings in the stratosphere tend to be modulated so that their induced anomalous zonal wind acceleration is enhanced by increased UV. Consequently, during a positive PJO winter (as in this study), wave forcings with a larger positive anomalous zonal wind acceleration (as for forcings with weaker convergence of the E-P flux) tend to appear, but during a negative PJO winter (reversed polarity compared with the present study), wave forcings with a larger negative anomalous zonal wind acceleration (as for forcings with stronger convergence of the E-P flux) tend to appear alongside an increase in UV.

4. Discussion and Remarks

Our observational analysis shows that the PJO tends to propagate farther down into the troposphere and is more strongly coupled with the tropospheric SAM in late winter to spring when solar activity is enhanced. A set of 42 year simulations produced using a chemistry-climate model was able to effectively reproduce observational data. The only variation in model boundary conditions was for UV, which was set at twice that observed and adjusted to be strong, normal, or weak. The solar UV forcing employed thus represents the different phases of the 11 year sunspot cycle: maxima mean and minima (albeit with exaggerated amplitude). Though the solar UV forcing has little effect on the mean state (see Figure 2), it has a significant effect on the PJO, especially in terms of its connection with the tropospheric SAM in spring. Analyses of zonal momentum and the wave energy budget suggest that modulation of the PJO due to changes in UV originates from modulation of interactions between planetary waves and mean flow in the stratosphere.

There are similarities between the results of the present study (Figure 1) and those of previous work [Kuroda et al., 2007, Figure 1], suggesting a significant correlation between indices of the PJO and stratospheric SAM on November. In fact, these indices have a correlation coefficient of 0.56, significant at the >99.99% level. The close relationship between the PJO and the stratospheric SAM in spring reflects the fact that a positive stratospheric SAM in spring tends to be associated with a positive PJO index in winter when the midwinter stratospheric polar vortex extends to lower latitudes. Thus, the influence of the PJO tends to persist until spring in the stratosphere.

The present results are consistent with those of Kuroda et al. [2007], who found that the correlation between the stratospheric and tropospheric SAM tends to increase with UV strength in December and originates from enhanced interactions between planetary waves and zonal-mean flow in the stratosphere in response to stronger UV. It should be noted that the correlation of the SAM index in December with that in November or the following January is highest in the SMX run. This result is consistent with Simpson et al. [2011], who found that a stronger stratospheric influence tends to prolong the duration of the SAM from late winter/spring to summer. The present analysis of the wave energy budget shows that stronger interactions between planetary waves and zonal-mean flow in response to stronger UV originate from enhanced energy transfer from zonal-mean fields rather than energy input from diabatic heating due to UV, contrary to the speculation of Kuroda et al. [2007].

The factors that underlie the modulation of wave-mean flow interactions in the stratosphere due to the change in UV remain unknown. Because stability plays an important role in equations (5) and (12), one possible mechanism is increased stability in the stratosphere when solar input is high. However, it is also possible that the combined effects of many factors, such as increased ozone and increased UV, contribute to such modulation. It should be noted that our budget analyses were based only on dynamical meteorological fields produced by model simulations and that chemistry effects are implicitly incorporated but not considered directly. More experiments and new analyses will be needed in future studies to identify key factors that underlie the modulation of interactions between planetary waves and zonal-mean flow in the stratosphere.
Appendix A

If equation (2b) is differentiated with respect to \( p \) and equation (2c) is substituted for the second term, we have

\[
2 \Omega \sin \phi \frac{\partial \sigma}{\partial p} - \frac{R}{4 \Omega^2} \frac{\partial T}{\partial \phi} = \frac{\partial J}{\partial p}. \tag{A1}
\]

If equation (A1) is differentiated with respect to time, equations (2a) and (2d) can be substituted for its first and the second terms, respectively, and it is divided by \( 4 \Omega^2 \sin^2 \phi \), giving

\[
\frac{\partial \sigma}{\partial p} + \frac{1}{2 \Omega \sin \phi} \frac{\partial (F_x + F_n + \chi)}{\partial \phi} - \frac{R}{4 \Omega^2 \sin^2 \phi} \left[ \frac{\partial \phi}{\partial \phi} + \frac{\partial}{\partial \phi} \left( \frac{Q_x + Q_n + \bar{S}}{4 \Omega^2} \right) \right] = \frac{1}{4 \Omega^2 \sin^2 \phi} \frac{\partial J}{\partial p}. \tag{A2}
\]

On the other hand, if equation (2e) is differentiated with respect to \( p \), we have

\[
\frac{1}{\cos \phi} \frac{\partial}{\partial \phi} \left( \frac{\partial \sigma}{\partial \phi} \cos \phi \right) + \frac{\partial^2 \sigma}{\partial \phi^2} = 0. \tag{A3}
\]

If equation (A2) is multiplied by \( \cos \phi \), differentiated with respect to \( \phi \), and the first term with \( \bar{\tau} \) is replaced by that with \( \bar{\sigma} \) from equation (A3), we finally obtain equation (5) after minor rearrangement.

References


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