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Is the stratospheric quasi-biennial oscillation affected by solar wind dynamic pressure via an annual cycle modulation?

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[1] In this study, statistical evidence of a possible modulation of the equatorial stratospheric quasi-biennial oscillation (QBO) by the solar wind dynamic pressure is provided. When solar wind dynamic pressure is high, the QBO at 30–70 hPa is found to be preferably more easterly during July–October. These lower stratospheric easterly anomalies are primarily linked to the high-frequency component of solar wind dynamic pressure with periods shorter than 3 years. In annually and seasonally aggregated daily averages, the signature of solar wind dynamic pressure in the equatorial zonal wind is characterized by a vertical three-cell anomaly pattern with westerly anomalies in both the troposphere and the upper stratosphere and easterly anomalies in the lower stratosphere. This anomalous behavior in tropical winds is accompanied by a downward propagation of positive temperature anomalies from the upper stratosphere to the lower stratosphere over a period of a year. These results suggest that the solar wind dynamic pressure exerts a seasonal change of the tropical upwelling that results in a systemic modulation of the annual cycle in the lower stratospheric temperature, which in turn affects the QBO during austral late winter and spring.

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1. Introduction

[2] The quasi-biennial oscillation (QBO) of the equatorial stratosphere is characterized by alternating easterly and westerly wind regimes and dominates the variability of the tropical lower stratosphere [*Naujokat*, 1986; *Baldwin et al.*, 2001]. The winds descend from 10 to 100 hPa at an average speed of 1 km/month and repeat at intervals from 18 to 36 months, with an average periodicity of ~28 months [*Pascoe et al.*, 2005]. Our current understanding of the QBO is largely based on the theoretical augment of *Holton and Lindzen* [1972], who showed how vertically propagating equatorial waves are absorbed at a critical layer by the mean flow and in turn generate alternating acceleration of the zonal flow in either easterly or westerly directions [*Plumb and Bell*, 1982].

[3] In addition to this wave-driven mechanism, seasonal or interannual variation of the QBO may also arise from seasonal changes in tropical upwelling [*Dunkerton*, 1991, 2000; *Holton et al.*, 1995; *Kinnersley and Pawson*, 1996], shifting of latitudinal gradients in the subtropics [*Randel et al.*, 1998], or variation of the semiannual oscillation (SAO) in the tropical upper stratosphere [*Dunkerton and Delisi*, 1997]. Additional temperature change may be caused by the secondary circulation, induced by the QBO itself, which consists of an increase in the upwelling at the easterly shear zone and a suppression of the upwelling at the westerly shear zone [*Plumb and Bell*, 1982].

[4] While the QBO is primarily driven by internal wave dynamics, a possible modulation of the QBO via external forcing cannot be ruled out. There is increasing evidence that part of the variability in the stratosphere may be linked to the solar activity [*Haigh*, 2003; *Labitzke et al.*, 2006; *Lockwood et al.*, 2010a, 2010b; *Gray et al.*, 2010]. Studies have indicated that solar variability may indirectly modulate the QBO through the variation in solar UV irradiance [*Salby and Callaghan*, 2000; *Soukharev and Hood*, 2001; *McCormack*, 2003; *Mayr et al.*, 2005, 2006; *McCormack et al.*, 2007]. Absorption of solar ultraviolet (UV) radiation by ozone may cause changes in temperature and hence in the QBO near the equatorial stratosphere [*Hood*, 2004; *Crooks and Gray*, 2005; *Pascoe et al.*, 2005]. Such anomalous heating in the upper stratosphere may alter Rossby wave propagation and breaking [*Cordero and Nathan*, 2005; *McCormack et al.*, 2007] and hence causes an indirect dynamic feedback through modulation of the polar vortex and the Brewer-Dobson (BD) circulation [*Kodera and Kuroda*, 2002]. *Salby and Callaghan* [2000] examined the radiosonde equatorial zonal wind at 45 hPa for the period of 1956−1996 and found that the average duration of the westerly QBO (wQBO) around solar maxima is ~3−6 months shorter than that around solar minimum. *Soukharev and Hood* [2001] performed composite analysis using the radiosonde-derived equatorial zonal wind at 50−10 hPa from 1957 to 1999 and found that the duration of both QBO phases was shorter at solar maxima than at solar minima.

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Using European Centre for Medium Range Weather Forecasting (ECMWF) ERA-40 reanalysis, *Pascoe et al.* [2005] found that the average time required for the eQBO to descend from 15 to 44 hPa is ~2 months less at solar maximum than at solar minimum. By using a fully interactive 2-D chemical-dynamical model, *McCormack* [2003] and *McCormack et al.* [2007] found that the duration of the wQBO can be 1–3 months shorter at solar maximum than solar minimum. They suggested that the ozone heating induced by solar UV in the upper stratosphere reduces the tendency for a westerly wind and hence produces an anomalously stronger easterly wind in the tropical upper stratosphere. By generating the QBO internally using a high-resolution general circulation model (GCM), *Palmer and Gray* [2005] showed that the durations of both QBO phases may be reduced by up to 2 months at solar maximum. *Cordero and Nathan* [2005] showed that prescribing higher solar UV inputs produces a QBO with shorter duration and larger amplitude. In addition to the direct ozone heating, wave-ozone feedback has been shown to modify the refraction of tropical quasi-stationary Rossby waves, reducing the tropical upwelling and resulting in faster descent of eQBO [*Nathan and Cordero*, 2007]. *Mayr et al.* [2006] prescribed a synthetic period of 10 years with varying amplitude of radiative forcing at three different heights (0.2%, surface; 2%, 50 km; 20%, 100 km), and their model simulation also resulted in a modulation of the QBO in the stratospheric equatorial region which appeared to be generated by this synthetic solar cycle (SC) modulation. *Mayr et al.* [2005] suggested that an annual oscillation can originate near 60 km, through which the prescribed 10 year modulation could be transferred downward and amplified by tapping the momentum from the upward propagating gravity waves.

[5] In contrast, other studies have been unable to find a significant link between solar UV and QBO variability. *Hamilton* [2002] used data from 1950 to 2000 and found a quasi-decadal variation in the duration length of the wQBO at 40–50 hPa, but the connection with the 11 year solar cycle noted by *Salby and Callaghan* [2000] did not always hold, and the correlation coefficient was only −0.1 when computed over a longer period that included 22 westerly phases. *Fischer and Tung* [2008] used radiosonde measurements for 1953–2005 and showed that the correlations between the duration of the QBO and the 11 year SC for either phase of the QBO were close to zero. In addition, oppositely signed correlations were found to exist in the pre-1990 and post-1990 periods. More recently, *Kuai et al.* [2009a, 2009b] also dismiss the solar cycle modulation of the QBO period. With exceedingly long model runs, those authors found that a strong synchronization of the QBO period with integer multiples of the semiannual oscillation in the upper stratosphere may generate “decadal” variability in the QBO period, causing a waxing and waning of the correlation between solar forcing and the QBO period. They therefore suggested that the SC-QBO duration relationship reported previously reflects only a chance behavior resulting from the use of relatively short observational records.

[6] In addition to solar UV, signature of the solar wind streaming out from the Sun has also been found in various climate records [*Lu et al.*, 2008a; *Lockwood et al.*, 2010a, 2010b; *Woollings et al.*, 2010]. The solar wind variability manifests itself at Earth in a number of ways but most prominently through geomagnetic activity, thermospheric heating, and the aurorae. Observational studies have shown that the solar wind–induced geomagnetic activity may alter stratospheric chemistry indirectly through particle precipitation [*Solomon et al.*, 1982; *Randall et al.*, 2005, 2007; *Seppälä et al.*, 2007; *Siskind et al.*, 2007]. Chemical-dynamical coupled general circulation models (GCMs) have indicated that odd nitrogen (NO_x) induced by energetic charged particle precipitation during geomagnetic storms may cause temperature changes in both the polar and equatorial regions and in the stratosphere and the troposphere [*Langematz et al.*, 2005; *Rozanov et al.*, 2005; *Seppälä et al.*, 2009].

[7] Other studies suggested that solar wind–induced geomagnetic activity may perturb atmospheric circulation through a change in planetary wave reflection conditions [*Arnold and Robinson*, 2001; *Lu et al.*, 2008b]. Model simulations by *Arnold and Robinson* [2001] have indicated that a change in planetary wave reflection conditions at the lower thermospheric boundary caused by thermsopheric heating induced by solar wind–driven geomagnetic activity can reduce planetary wave propagation into the stratosphere, leading to a strengthening of the stratospheric polar vortex and a weakening of the BD circulation. Significant correlations have also been established between geomagnetic activity and lower stratospheric temperature by *Lu et al.* [2007]. They found that the temperature response to the 11 year SC and geomagnetic *Ap* index tend to enhance each other in the tropical upper troposphere and lower stratosphere, the magnitude of the response to geomagnetic activity being slightly larger than that associated with the 11 year SC. *De Artigas and Elias* [2005] have shown that, when solar flux is high, the eQBO (at 15–50 hPa) is more likely to be associated with high geomagnetic activity and the wQBO is more likely to be associated with low geomagnetic activity; at solar minimum, this relationship is reversed. *Lu et al.* [2008a] have shown that there is a robust relationship between solar wind dynamic pressure and the zonal wind and temperature in the northern polar winter. Stratospheric wind and temperature variations are positively projected onto the Northern Annular Mode (NAM) when the 11 year SC is at its maximum phase and negatively projected onto the NAM during the 11 year SC minimum phase. This implies a weakening of the BD circulation with reduced upwelling into the lower stratosphere at low latitude under high solar wind forcing, consistent with the tropical warming signals detected by *Lu et al.* [2007].

[8] While variations in solar radiation are most strongly seen in the 11 year solar cycle, the solar wind–induced geomagnetic activity exhibits a strong variability at shorter time scales. *Lu et al.* [2008a] shows that the solar wind dynamic pressure and $F_{10.7}$ solar flux are poorly correlated, largely because of those shorter time variations. Their weak statistical dependence allows the signals associated with the 11 year solar irradiance and the solar wind dynamic pressure to be separated. However, their mechanisms may not be mutually exclusive and they could be acting together to either enhance or cancel the overall atmospheric responses.

[9] Motivated by these results and rather strong extra-tropical signals in Northern Hemisphere (NH) winter and spring observed by *Lu et al.* [2008a], a statistical analysis is

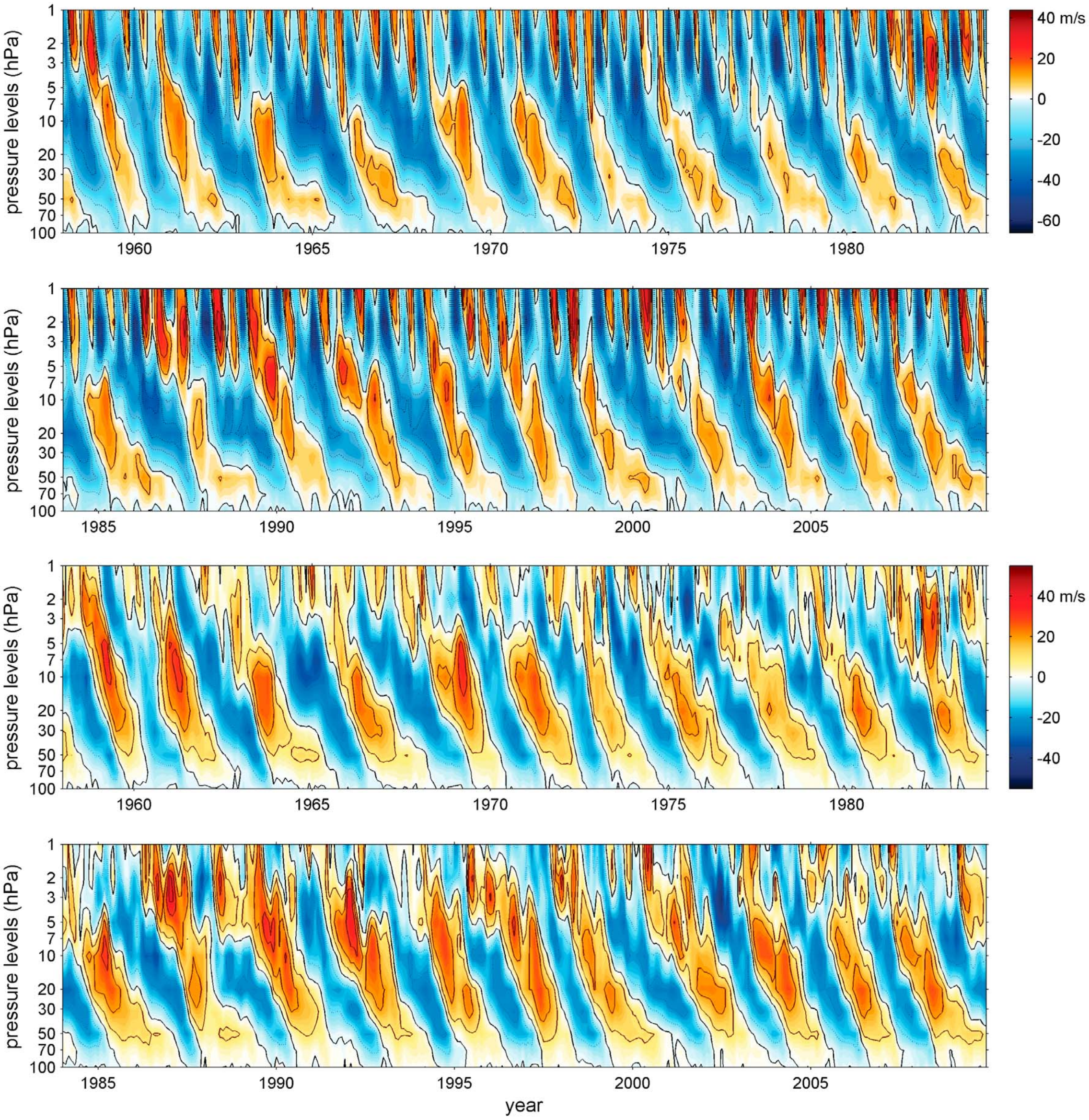

Figure 1. Height-time cross section of the monthly mean zonal wind (first and second panels) and the QBO time series (i.e., deseasonalized monthly mean zonal mean zonal wind averaged over 5°S–5°N) (third and fourth panels) for the period of 1958–2009. Red and blue represent westerly and easterly winds, respectively.

conducted here to study possible solar wind dynamic pressure perturbations near the tropics, with a primary focus on detecting possible solar wind perturbation of the QBO. We use solar wind dynamic pressure, instead of geomagnetic indices, because geomagnetic indices are measured through the interaction of the terrestrial environment with the solar wind and are therefore not completely independent of internal Earth atmospheric processes. For instance, *Sugiura and Poros* [1977] identified a QBO-like periodicity in the disturbed storm time current (*Dst*) and hence in geomagnetic activity, and *Jarvis* [1996] has shown a QBO periodicity in geomagnetic daily range induced by the semidiurnal tide. Conversely, the solar wind dynamic pressure is measured well outside of Earth's magnetosphere and is therefore wholly independent of the internal variability of the Earth's atmosphere. Nevertheless, it can be shown that similar but slightly weaker results are obtainable by using the geomagnetic *Ap* index.

2. Data and Methods

[10] In this study, daily QBO is defined as the deseasonalized daily zonal mean zonal wind averaged over the latitude band of 5°S–5°N for five pressure levels at 70, 50, 30, 20, and 10 hPa. Zonal mean zonal winds are extracted from the ECMWF ERA40 reanalysis (September 1957 to August 2002) and ECMWF operational analyses (September 2002 to December 2009). The ERA-40 reanalysis was assimilated using direct radiosonde and satellite measurements [*Uppala et al.*, 2005]. It has a spectral resolution of

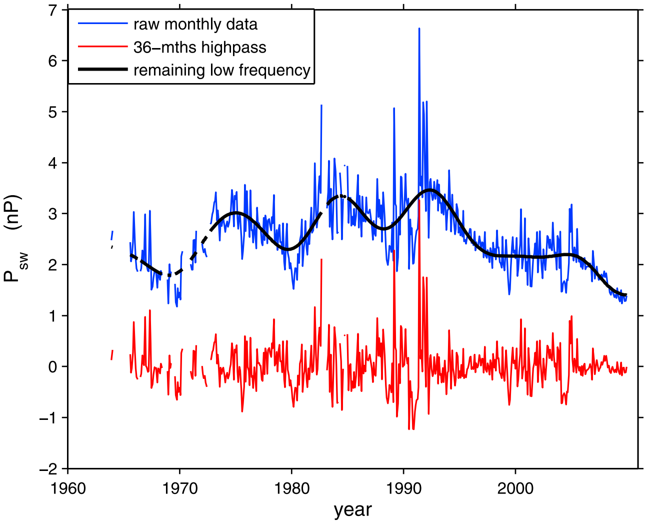

Figure 2. Time series of monthly solar wind dynamic pressure P_{sw} (blue line), with its high- and low-frequency components shown as the red and solid black lines. The frequency separation is done by treating the original monthly P_{sw} with an order 11 Butterworth high-pass filter with a cutoff period of 36 months.

T159, corresponding to a 1.125° horizontal resolution in latitude and longitude and is available at 23 standard pressure surfaces from 1000 to 1 hPa. The ECMWF operational data were output from the ongoing analyses produced by the most recent ECMWF integrated forecasting system (IFS) model. Data from September 2002 to the present day are available on the same 1.125° grid and on 21 pressure levels (before 11 July 2007) and 25 pressure levels (since 11 July 2007). The ERA-40 and the operational data sets share 21 pressure levels; the exceptions are four levels in the lower troposphere (i.e., 600, 775, 900, and 950 hPa) that are excluded from this analysis. Nevertheless, the two data sets share the same five pressure heights where the QBOs are estimated.

[11] It has been established that before the satellite era (i.e., before September 1978), the scarcity of Southern Hemisphere (SH) radiosonde measurements and lack of direct measurement at altitudes above 10 hPa resulted in unreliable estimations in ERA-40, particularly in the Southern Hemisphere and in the upper stratosphere. In the tropics, however, *Baldwin and Gray* [2005] showed that the QBO extracted from ERA-40 is consistent with rocketsonde winds (which were not assimilated by ERA-40) measured at Ascension and Kwajalein up to the 2–3 hPa altitude level, even for those years before the satellite era. *Pascoe et al.* [2005] also found that the ERA-40 accurately describes the QBO and the tropical stratospheric circulation. We also compared the monthly averaged QBO with those available at the Free University of Berlin [*Naujokat*, 1986] for the five pressure levels used here. The two data sets are in good agreement with a correlation coefficient greater than 0.9 for all five pressure levels. In addition, the response to solar wind dynamic pressure reported in this paper is found primarily below 10 hPa. Hence, for these reasons, the results reported here make use of all available data since 1963. Figure 1 shows latitudinally integrated zonal mean zonal wind at 5°S–5°N as a function of height (1–100 hPa) for the period of 1958–2009.

[12] Figure 1 (first and second panels) displays the original monthly mean data, while Figure 1 (third and fourth panels) shows its deseasonalized anomalies. A prominent semiannual oscillation exists near the stratopause level when the data are not deseasonalized. Once deseasonalization is applied, the descending alternating easterly and westerly winds of the QBO are the dominant feature, with maximum amplitude of ~40 m s^{-1} at ~3–5 hPa, where the QBO is initialized. In the upper stratosphere, discernible enhancement of easterly anomalies occurred in the 1970s, while an enhancement of westerly anomalies occurred from the mid-1980s to the late 1990s. From 2000 onward, it becomes comparable to the earlier pattern again. It has been suggested that the enhanced westerly anomalies seen in the upper stratosphere during the 1980s and 1990s might be a result of data assimilation presatellite and postsatellite eras [*Punge and Giorgetta*, 2007]. The extended data seem to suggest it is more likely to be a real phenomenon in which a multidecadal variation is superimposed upon the upper stratospheric SAO.

[13] Solar wind dynamic pressure (P_{sw}) measured in geocentric Earth magnetic coordinates is obtained from the NASA-OMNI Web site (http://omniweb.gsfc.nasa.gov/). This data set is produced from solar wind and interplanetary magnetic field measurements from 15 geocentric satellites and 3 spacecraft in orbit around the L1 Sun-Earth Lagrange point and has been carefully compiled through cross calibration. P_{sw} has been calculated by NASA-OMNI as $P_{\mathrm{sw}} = 1.6726 \times 10^{-6}\ N_{\mathrm{sw}} V_{\mathrm{sw}}^2$, where N_{sw} is the flow density in number of particles cm^{-3} and V_{sw} is the solar wind speed in km s^{-1}. In physical terms, P_{sw} represents the momentum flux of the solar wind and has a unit of nPa. Daily averages of solar wind P_{sw} from January 1963 to December 2009, covering ~4.5 solar cycles, together with their monthly mean, are used here. The main advantage of using the daily data over the monthly mean is that the downward decent of the signals can be studied in more detail.

[14] There are a few missing data periods of P_{sw} due to inappropriate positioning of satellites or instrument failure. For instance, before August 1965, and also between September 1982 and October 1994, the data availability is below 50% at hourly resolution, with 8–15 complete days showing as missing data in each month [*King and Papitashvili*, 2005]. For up to 3 days of missing values, a simple interpolation is used to fill the gap. For longer missing data periods, no treatment is applied and those days are ignored by the analysis. For the months with more than 15 days of missing values, monthly P_{sw} is treated as missing data. The effect of missing data for running composite or regression is that different months or seasons under investigation may involve different sample lengths. While the monthly P_{sw} varies from 1.2 to 6.7, the daily P_{sw} varies from 0.1 to 27 nPa with an arithmetic mean value of 2.5 nPa and standard deviation of 1.5 nPa (not shown).

[15] Figure 2 shows monthly mean P_{sw} and its high- and low-frequency components, where the separation cutoff period is 36 months. The low-frequency P_{sw} shows a quasi-decadal variation superimposed on a slow varying trend. Its quasi-decadal variation peaks at the minimum phase of the 11 year solar cycle, suggesting an 11 year solar cycle

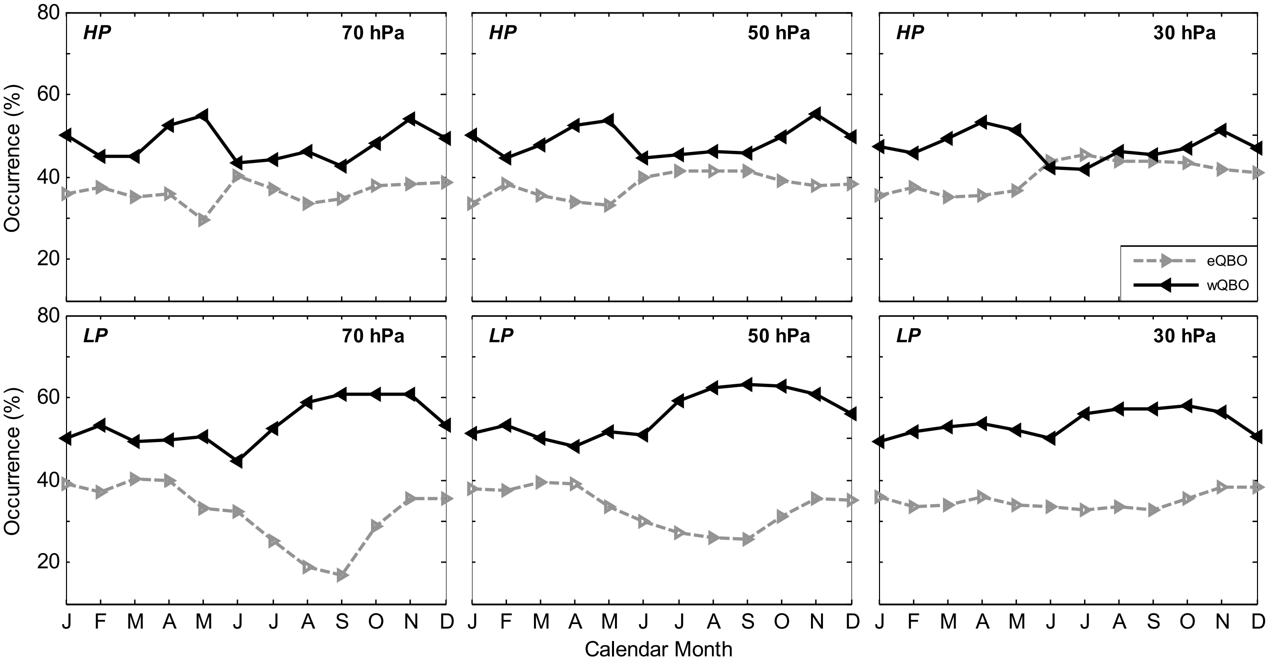

Figure 3. The QBO phase occurrences (top) under HP and (bottom) under LP at 70–30 hPa pressure levels, calculated on the basis of daily data. Westerly winds (wQBO) are denoted as solid lines, and easterly winds (eQBO) are denoted as dashed gray lines. The wQBO and eQBO are defined as >0.25 and <−0.25 of the normalized monthly mean of the QBO.

modulation on P_{sw}. There is clear evidence of slow-varying 11 year cycle in its high-frequency component. Nevertheless, over the period 1963–2009, the correlation coefficient between P_{sw} and sunspot number or $F_{10.7}$ solar flux is below 0.1 on a daily time scale and remains below 0.2 for the monthly mean. This is largely due to the high-frequency variation of P_{sw}. Thus, any significant P_{sw} signals, especially those associated with the high-frequency component, should be fairly statistically independent of the signature associated with the 11 year solar cycle.

[16] For simplicity, we use $\overline{P}_{sw} < -0.25$ and $\overline{P}_{sw} > 0.25$ to define low and high solar wind dynamic pressure for either the raw P_{sw} or its high-frequency component, where $\overline{P}_{sw}$ stands for the median-normalized values of P_{sw}; transition periods where $|\overline{P}_{sw}| \leq 0.25$ are excluded. The same rule is applied to differentiate the westerly and easterly QBO phases. Qualitatively similar results can be obtained if other definitions, such as mean instead of median for $\overline{P}_{sw}$, are used. Hereinafter, high and low solar wind dynamic pressure are abbreviated as HP and LP.

3. Results

3.1. Perturbations Caused by Solar Wind Dynamic Pressure

[17] Figure 3 shows the QBO phase occurrence by month as a percentage separated into high (Figure 3, top) and low (Figure 3, bottom) solar wind dynamic pressure for 70 hPa (Figure 3, left), 50 hPa (Figure 3, middle), and 30 hPa (Figure 3, right) estimated using daily data. Under HP, the difference in occurrence between eQBO and wQBO is noticeably smaller than it is under LP at all three pressure levels, and the difference even reverses at 30 hPa. The largest difference occurs under LP conditions during the austral winter and spring and reduces with increasing altitude. The departure between the QBO phase occurrence under HP and LP begins in June–July. By assuming that both the QBO and P_{sw} follow a first-order autoregressive process, we use a Monte Carlo trial-based test [*Wang et al.*, 2006] to determine the significance of the differences. It is found that the differences under LP at 50 and 70 hPa are significant at the 95% confidence level only for July–October, while the differences at 30 hPa are significant at the 95% for almost the entire year (except December and January). As a whole, the behavior of the QBO phase occurrence differs from January–June, when the occurrence of wQBO hardly changes from LP to HP, to July–December, when the occurrence of wQBO increases and the occurrence of eQBO decreases significantly under LP. This suggests a higher eQBO to wQBO occurrence ratio in the lower stratosphere under HP than under LP conditions. It suggests that under HP conditions there is either anomalously more eQBO descent down from higher altitudes or a slower descent rate of the eQBO at 30–70 hPa. The difference becomes weaker and not significant at higher altitudes.

[18] The same conclusion can be reached if the QBO phase occurrence is grouped according to QBO phase and then the occurrence percentages are calculated for HP and LP conditions (see Figure 4). At 70 hPa during September, nearly 50% of eQBO occurrence takes place under HP compared with only 20% under LP; the difference becomes considerably smaller for wQBO. The differences under eQBO for August–October for all three pressure levels from 30 to 70 hPa are significant at the 95% confidence level or above and at the 90% confidence level for July. This confirms that significantly more eQBO occurred in the lower

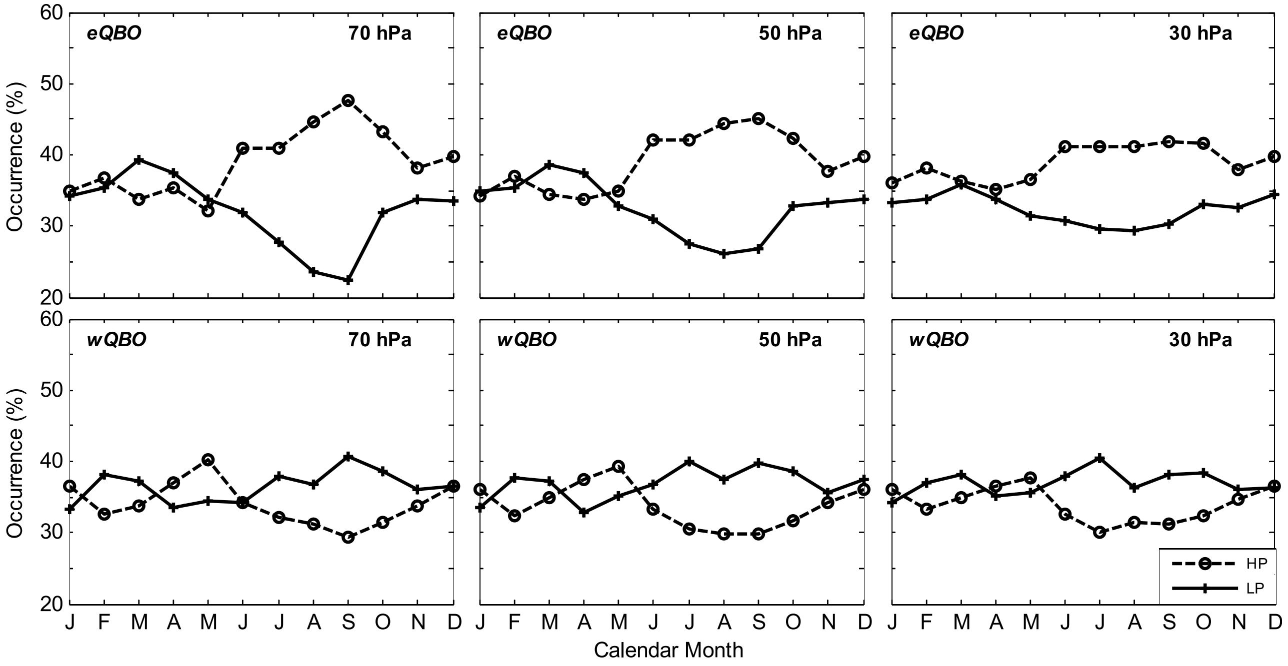

Figure 4. The QBO phase occurrences in relation to high (HP, solid line with pluses) and low (LP, dashed line with circles) solar wind dynamic pressure for (top) eQBO and (bottom) under wQBO phases at 70–30 hPa pressure levels. HP and LP are defined as >0.25 and <−0.25 of the normalized monthly P_{sw} median.

stratosphere during the austral late winter and spring and a positive relationship exists between P_{sw} and increased eQBO occurrence.

[19] Figure 5a shows the vertical profile of annual mean equatorial zonal wind (solid line) ±1 standard deviation. Greater variability exists in the stratosphere than in the troposphere largely because of the QBO. Figures 5b, 5c, and 5d show the vertical profile of tropical zonal wind anomalies for separated HP and LP conditions under three different sampling conditions, i.e., for 1963–2009 (Figure 5b), for 1979–2009 (Figure 5c), and for 1963–2009 (Figure 5d) but with volcanic eruption affected data excluded. All the vertical profiles of zonal wind anomalies are calculated by using daily deviation from the long-term mean and then aggregating to the annual average. It shows that the modulation of solar wind dynamic pressure on daily equatorial zonal wind is marked by a vertically three-cell structure with westerly anomalies (~0.3–0.5 m s^{-1}) in the troposphere, easterly anomalies (~2 m s^{-1}) at 20–70 hPa, and westerly anomalies (~3 m s^{-1}) at 3–10 hPa under HP-LP. As a whole, this vertical structure accounts for ~5–10% of 1 standard deviation of the tropical zonal wind for each associated vertical region, and these wind differences are significant at the 95% confidence level. Changing the data period from 1963–2009 (Figure 5b) to 1979–2009 (Figure 5c) or excluding the years affected by major volcanic eruptions (Figure 5d) does not alter its general pattern or significant levels. Similar results can also be obtained by excluding the years affected by the major Niño events.

[20] Figure 6 shows that the same vertical profile of equatorial zonal wind anomalies also holds on a seasonal time scale and the general vertical structure remains across all four seasons (i.e., December–February, March–May, June–August, and September–November). In the lower stratosphere, the largest and most significant departure from zero wind occurs during austral winter (June–August). A closer examination also suggests that there is a noticeable seasonal variation in its vertical structure. From December to February, an upward shift is obtained for the bottom cell of the stratosphere; from September to November, a downward shift is apparent for both top and bottom cells in the stratosphere. The largest differences between HP and LP conditions in the upper stratosphere occur in September-October-November and December-January-February, while in the lower stratosphere they occur in March-April-May and June-July-August. Similar results can also be obtained by excluding the years affected by either the major El Niño events or the major volcanic eruptions.

[21] Figure 7a shows the 31 day running composite difference (HP − LP) of the equatorial zonal mean zonal wind from June to October. Significant easterly anomalies (~8 m s^{-1}) are found in 30–70 hPa pressure levels, while westerly anomalies (~1 m s^{-1}) are simultaneously detected in the upper troposphere. It is apparent that the stratospheric easterly anomalies descend from 10 hPa or above from June and become significant from late July to early October. Figure 7b shows the same running composite difference except that a high-pass filter with a cutoff period of 1095 days (~3 years) is applied to the daily P_{sw}. Note that both the strength of the composite difference and the statistical confidence level are enhanced in the stratosphere, suggesting that not the decadal-scale variability of P_{sw} but the variation of P_{sw} at periods below 3 years is responsible for the easterly anomalies at 30–50 hPa. Longer-term variation of P_{sw} appears to be responsible for the anomalies at and below 70 hPa.

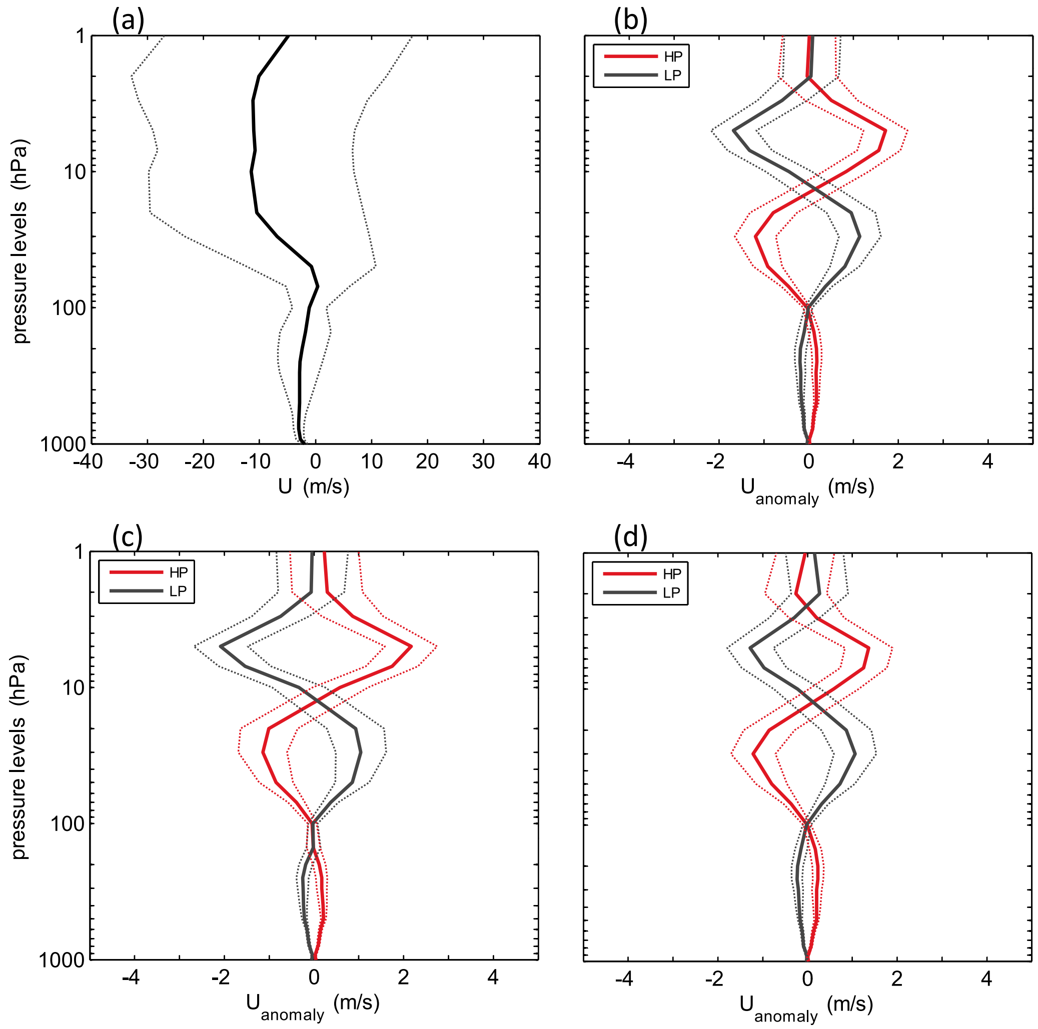

Figure 5. (a) Vertical profile of the annual mean tropical wind (solid line) and ±1 standard deviation (dotted lines). (b) Departure from the annual mean for HP and LP conditions (solid lines) for the period of 1963–2009. (c) Same as Figure 5b but based on data from 1979 to 2009. (d) Same as Figure 5b but based on data when 2 years after each major volcanic eruption are excluded. In Figures 5b, 5c, and 5d, 95% confidence intervals are shown as dotted lines for both HP and LP conditions. When the dotted lines do not overlap, it indicates that the average differences between the HP and LP groups are significant at or above the 95% confidence level. All the departures are calculated as deviations from the monthly mean on the basis of daily data which are then aggregated to give the annual mean departure.

[22] The radiative heating induced by volcanic aerosols may also influence the lower stratospheric temperature, which in turn may lead to stalling of the downward propagation of the QBO [*Dunkerton*, 1983]. To examine possible contamination by the temporary heating caused by volcanic aerosols, Figure 7c shows the same composite difference as Figure 7b but with the major volcanic affected data (i.e., Agung in March 1963, El Chichón in March 1982, and Pinatubo in June 1991) excluded. A quantitatively similar result is obtained, suggesting that those easterly anomalies in the lower stratosphere are not strongly biased by the major volcanic induced warming in the lower stratosphere. Note that there are quite a lot of missing data in the daily P_{sw} during 1963, 1964, 1982, and 1983 (see Figure 2), and the contamination of those years is already excluded in the analysis of Figures 7a and 7b, meaning that the data affected by Pinatubo volcanic eruptions are the only difference between Figures 7b and 7c.

[23] It has been observed that ozone concentrations and temperature in the tropical lower stratosphere (~70 hPa) are anomalously low during the El Niño phase of the El Nino–Southern Oscillation (ENSO) [*Randel et al.*, 2009]. To examine possible bias due to the large cooling effect caused by the major El Niño events, Figure 7d shows the same analysis as Figure 7b but with the major El Niño–affected years (i.e., 1972, 1973, 1982, 1983, 1997, and 1998) excluded. Note that, because of missing P_{sw} in 1982 and 1983, the data affected by 1972, 1973, 1997, and 1998 represent the only difference between Figures 7b and 7d. Again a quantitatively similar result is obtained, suggesting that the major El Niño events do not alter the QBO-P_{sw} relationship significantly.

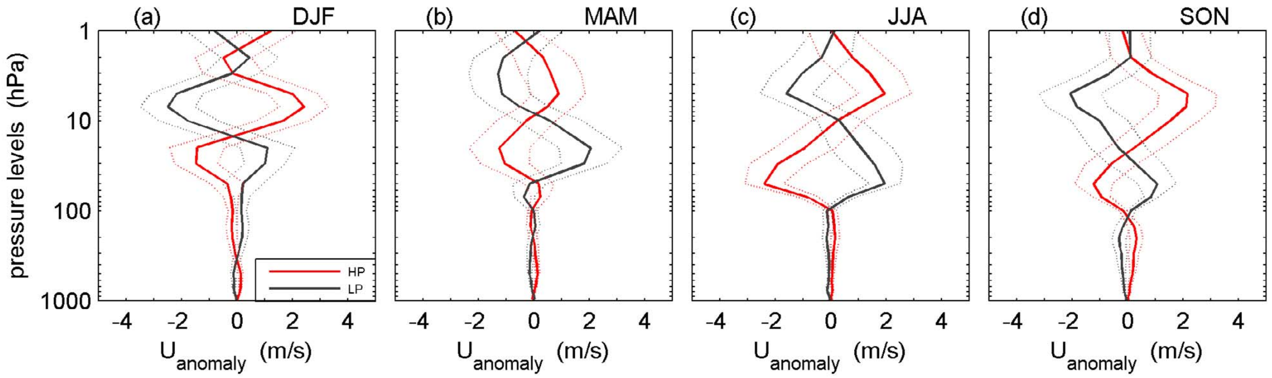

Figure 6. Same as Figure 5b but for (a) December–February, (b) March–May, (c) June–August, and (d) September–November means. The departures from the seasonal means are calculated on the basis of daily data aggregated into a seasonal mean departure. The usage of the lines and the definition of significant levels are the same as in Figure 5.

[24] Figure 8 summarizes the effect of the higher-frequency (<3 years, the red line in Figure 2) components of P_{sw} on the QBO at 50 and 30 hPa (Figures 8a and 8d), where both P_{sw} and the QBO are averaged from July to October. These pressure levels and months are chosen because the results shown in Figures 3, 4, and 7 suggest the largest effect of P_{sw} on the QBO at those levels and during those months. At both pressure levels, there are significant negative correlations between mean P_{sw} and the mean QBO. Neither volcanic eruption nor El Niño–affected years are found to dominate or alter the correlations significantly. This helps to rule out a possible linear contamination of the QBO-P_{sw} relationship by volcanic eruption and the major El Niño events. However, it does not rule out the possibility that the ENSO might modulate the QBO-P_{sw} relationship through nonlinear processes and a possibility of a modulation in other calendar months or at different frequency. This is beyond the scope of the current study.

[25] Figure 9 shows the running composite HP − LP difference of the equatorial zonal mean temperature from January to December. The largest positive anomalies (~2 K) are found from August to November at 30–50 hPa. Positive anomalies (~1 K) are also found at 70–150 hPa (becoming significant between 100 and 150 hPa) during the December–June period, while significant negative anomalies (~0.5 K) are detected at 300–700 hPa between July and October. The two most striking and significant features of Figure 9 are (1) downward propagation of positive temperature anomalies in the upper stratosphere from December of the previous year to the lower stratosphere in December of the following year and (2) the anomalous warming at ~100 hPa during the boreal winter and spring and anomalous cooling at 700–300 hPa during the austral winter. The combined effect of these temperature anomalies is a systematic annual modulation of the annual oscillation of temperature in the lower stratosphere related to solar wind dynamic pressure.

[26] Such annual cycle modulation can be compared to the annual oscillation normally present in the stratosphere. Figure 10a shows the climatological mean annual pattern of the tropical zonal mean zonal wind. The strongest variation appears in the upper stratosphere (1–5 hPa), with a semi-annual cycle clearly visible. In the lower stratosphere (50–100 hPa), a weaker semiannual cycle also exists, and this is roughly in phase of that in the upper stratosphere. In addition, there is a pronounced annual cycle both in the upper and lower stratospheres. In the upper stratosphere, the annual cycle is marked by nearly 3 times stronger easterly winds in the boreal winter (~40 m s^{-1}) than in the austral winter (~15 m s^{-1}). In the lower stratosphere, the situation reverses, with weaker easterlies occurring in the boreal winter (~2 m s^{-1}) and stronger easterlies in the austral winter (~5 m s^{-1}). In the 10–50 hPa region where the QBO prevails, the magnitude of zonal mean wind is primarily easterly (~15 m s^{-1}). It also shows a modulation by an annual cycle, albeit weak, with stronger easterly winds occurring in the austral winter.

[27] Figure 10b shows the temperature anomaly (shaded contours) compared to the climatological altitude-dependent mean value of tropical zonal mean temperature (thick solid lines). In the upper stratosphere, similar to the winds, the SAO dominates. The amplitude of temperature there during the boreal winter and spring (4–5 K) is about twice as large as it is in austral winter and spring (2–3 K), implying an additional annual cycle influence. An annual cycle dominates at 50–100 hPa, with a magnitude of −3 K during the boreal winter and 4 K in austral winter, a peak-to-peak value of ~8 K at 70 hPa. The semiannual cycle in the upper and annual cycles in the lower stratosphere oppose each other during the austral winter, resulting in a close to flat temperature climatology at ~20 hPa. The climatological tropical mean wind and temperature compare well with previous studies [*Baldwin et al.*, 2001; *Pascoe et al.*, 2005; *Fueglistaler et al.*, 2009].

[28] Thus, the annual oscillation linked to solar wind dynamic pressure (Figure 9) works in opposition to the normal annual oscillation in the lower stratosphere (i.e., 50–100 hPa), reducing its amplitude. It is possible that, because the eQBO is most sensitive to diabatic change in the lower stratosphere and particularly to temperature changes near the tropopause [*Kinnersley and Pawson*, 1996], the modulation effect of solar wind dynamic pressure is stronger for eQBO than for wQBO.

[29] The seasonal variations of the equatorial wind and temperature anomalies in relation to the variation of solar

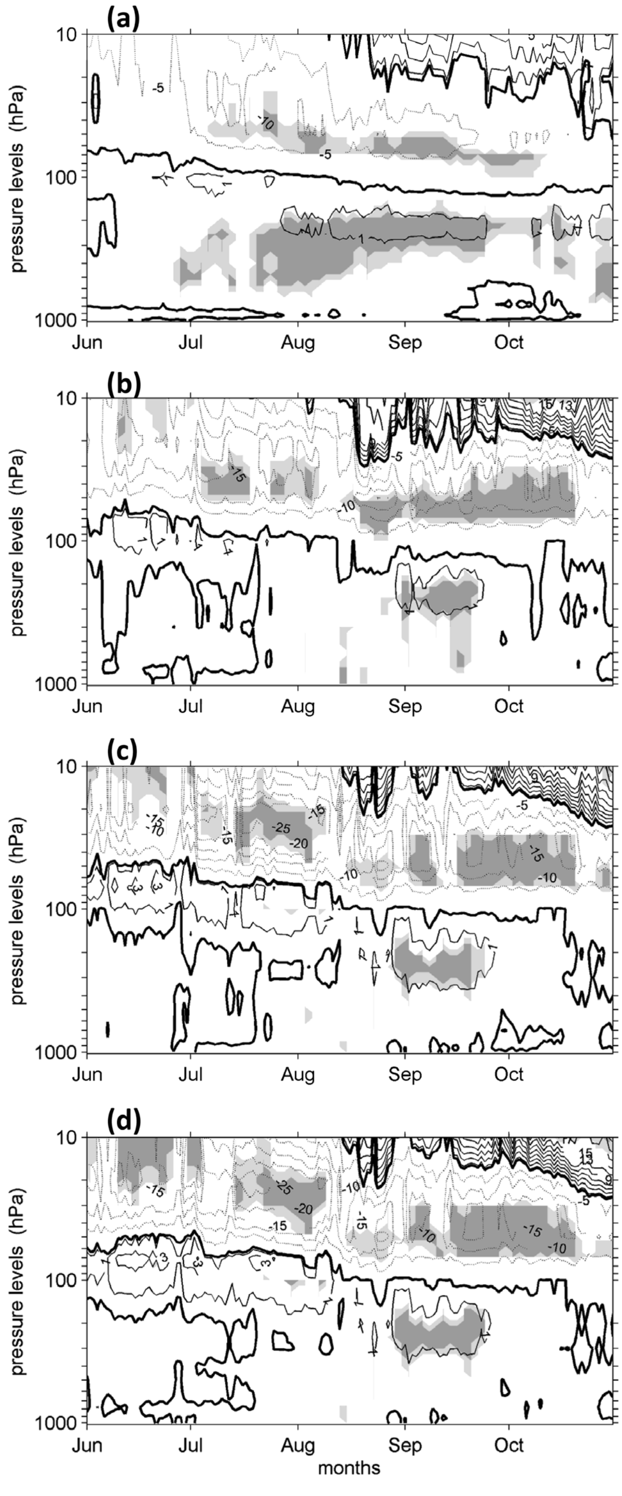

Figure 7. Height-time cross section of running composite difference of the daily equatorial zonal wind (in m s^{-1}) from 1 June to 30 October for HP-LP condition. A 31 day running window is applied to both the wind and P_{sw} without any time lag. (a) Linear regression based on raw P_{sw}. (b) High-pass filter with cutoff period of 1095 days (~3 years). (c) Same as Figure 7b, but data affected by the major volcanic eruptions are excluded. (d) Same as Figure 7b, but data affected by the major El Niño (1972, 1973, 1982, 1983, 1997, and 1998) are excluded.

wind dynamic pressure can also be compared with the seasonal statistics of QBO phase transitions. Figure 11 shows every major phase transition of the QBO between 1953 and 2009 at five different pressure levels grouped by months into histograms, where the transitions before 1958 were determined by using deseasonalized monthly radiosonde data and after transitions after 1958 were based on daily data. The months where the transitions occurred may differ by a month or two when compared to the earlier findings [*Dunkerton*, 1990; *Baldwin et al.*, 2001; *Hampson and Haynes*, 2004; *Christiansen*, 2010] as our deseasonalization is based on extended daily (instead of monthly) data from 1958 to 2009. In terms of the general distribution of the transitions, they remain similar to the previous studies based on radiosonde estimates, especially at 50 hPa. At 30 hPa, the peak of the phase transition tends to occur about 2 months earlier than those based on monthly NCEP reanalysis [see *Hampson and Haynes*, 2004, Figure 1].

[30] The histograms of transitions at 30–70 hPa show a noticeable annual cycle variation for both QBO phases. It is clear from Figure 11 that at 30 and 50 hPa the transitions occur less often from July to November than at other months. The months and pressure levels with reduced QBO phase transition coincide with the period (height) when (where) the effect of solar wind dynamic pressure on both equatorial zonal wind and temperature is most significant (see Figures 7 and 9). Similarly, the transitions of both QBO phases at 10 hPa occurred far less often during NH winter and spring than during other months. Again, this coincides with the period when solar wind dynamic pressure–related positive temperature anomalies were found to be present at that pressure level and above (see Figure 9).

4. Discussion

[31] The variability of the stratospheric QBO is known to be largely characterized by the "stalling" of easterly QBO (eQBO) near 30–50 hPa as the easterly shear zone tends to descend more irregularly than the westerly shear zone [*Naujokat*, 1986]. Modeling simulations have suggested that the "stalling" of the QBO depends critically on whether deposition of easterly momentum at the equator is sufficient to overcome the tendency for the easterly shear zone to be advected upward or held static by the combination of the BD circulation and the QBO-induced residual circulation [*Dunkerton*, 1991, 2000; *Kinnersley and Pawson*, 1996]. The tropical upwelling of the BD circulation acts to slow down the descent rate of the QBO and consequently extends the QBO period or increases the occurrence frequency for a given QBO phase. Because of the QBO-induced residual circulation, the effect of the tropical upwelling on the descent of the shear zone depends on the QBO phase. Upward motion is associated with eQBO [*Plumb and Bell*, 1982]: it strengthens the upwelling so that the descent of the easterly shear zone is slowed further. Conversely, downward motion is associated with wQBO: it causes cancellation of the tropical upwelling so that the BD circulation imposes a smaller slowing effect on the descent of the westerly shear zone. As a whole, the descent rate of the wQBO is less affected by variations associated with the BD circulation than that of the eQBO [*Kinnersley and Pawson*, 1996; *Dunkerton*, 2000; *Hampson and Haynes*, 2004].

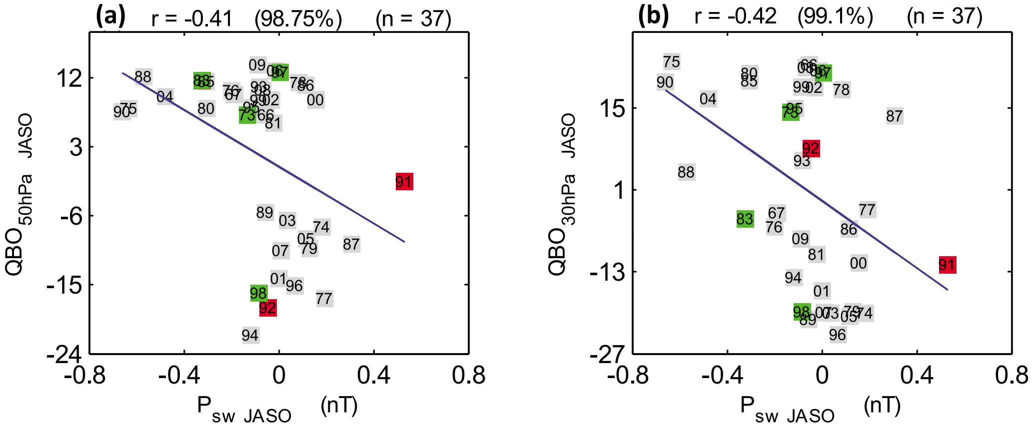

Figure 8. Correlations between the July–October averaged high-frequency P_{sw} (i.e., with period shorter than 3 years, denoted as $P_{\mathrm{sw\ JASO}}$) and the averaged QBO for the same calendar months (denoted as $\mathrm{QBO_{JASO}}$) at (a) 50 hPa and (b) 30 hPa. Individual years are shown as two digital numbers. Years affected by major volcanic eruptions and major El Niño events are highlighted as red and green squares, respectively. The correlation coefficient, statistical confidence level (in parentheses), and number of samples used (also in parentheses) are given at the top of each plot.

[32] The annual variation of the QBO may arise from a seasonal variation of the BD circulation, which is explained by the "extratropical stratospheric pump" mechanism [*Holton et al.*, 1995]. This is dynamically driven by momentum dissipation of extratropical waves so that air is drawn upward from the tropical troposphere and then poleward and downward at high latitudes, causing a seasonal variation in the BD circulation [*Yulaeva et al.*, 1994]. The periods with stronger overturning require stronger diabatic heating at the ascending branch and cooling at the descending branch, which cools the stratosphere in the tropics and warms it in the polar region. Because the planetary wave drag in the stratosphere is stronger in the northern hemisphere winter than in the southern hemisphere winter, it drives stronger upwelling and lower temperatures in the tropical lower stratosphere during NH winter. As a result, the descent of the eQBO is preferentially stronger during the months of May to July, when the tropical upwelling in the lower stratosphere is weakest [*Baldwin et al.*, 2001]. Here we show that, on top of the modulation due to the extratropical wave-driven BD circulation, solar wind dynamic pressure may also perturb the tropical zonal wind in the lower stratosphere, and the largest effect is detected during July–October.

[33] It is known that the annual cycle in lower stratospheric temperature plays an important role in the QBO phase transition and occurrence [*Fueglistaler et al.*, 2009]. Previous studies have suggested that the annual synchronization of the QBO phase transition is linked to seasonally varying tropical upwelling [*Dunkerton*, 1990; *Hampson and Haynes* 2004], and the annual cycle is a consequence of faster (slower) tropical upwelling in the boreal (austral) winter [*Yulaeva et al.*, 1994; *Holton et al.*, 1995; *Rosenlof*, 1995]. More recent studies have, however, challenged the "extratropical pump mechanism" and have resulted in some debate regarding the processes driving upwelling in the tropical BD branch [*Fueglistaler et al.*, 2009]. It has been shown by both observation and numerical calculation that tropical wave forcing produced by convective motion may also be partially responsible for the variation in the lower stratospheric temperature and the QBO [*Norton*, 2006; *Randel et al.*, 2008; *Lu et al.*, 2009; *Taguchi*, 2009]. Here we present some statistical evidence of a possible modulation of the lower stratospheric annual cycle via solar wind dynamic pressure P_{sw} and the descent pattern of the QBO and tropical temperature anomalies. We show that there is a significant perturbation of tropical zonal wind, apparently related to P_{sw}, in both the stratosphere and the troposphere. On the basis of these results, we suggest that, in addition to atmospheric internal variability, solar wind–driven modulation may also play a role in causing interannual variation in the strength and descent pattern of the QBO.

[34] More recently, *Lu et al.* [2009] found that the processes that synchronize the QBO exert different effects on different QBO phases from season to season. During July and August, the phase speed is significantly slowed when westerlies prevail through the pressure levels and becomes more irregular when easterlies are dominating there. They found that tropical upwelling alone is insufficient to explain

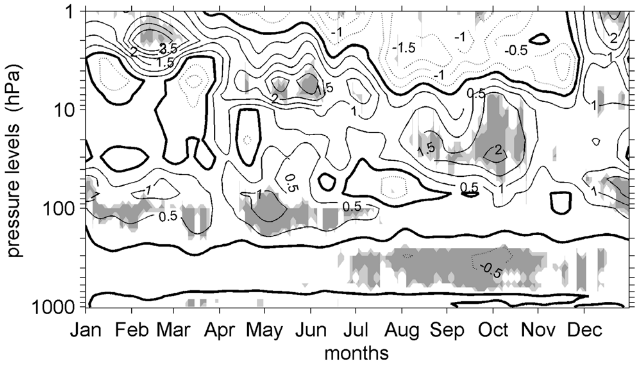

Figure 9. Same as Figure 7a except that the equatorial zonal mean zonal wind is replaced by equatorial zonal mean temperature, the *x* axis is extended from 1 January to 31 December, and the *y* axis is from 1000 to 1 hPa.

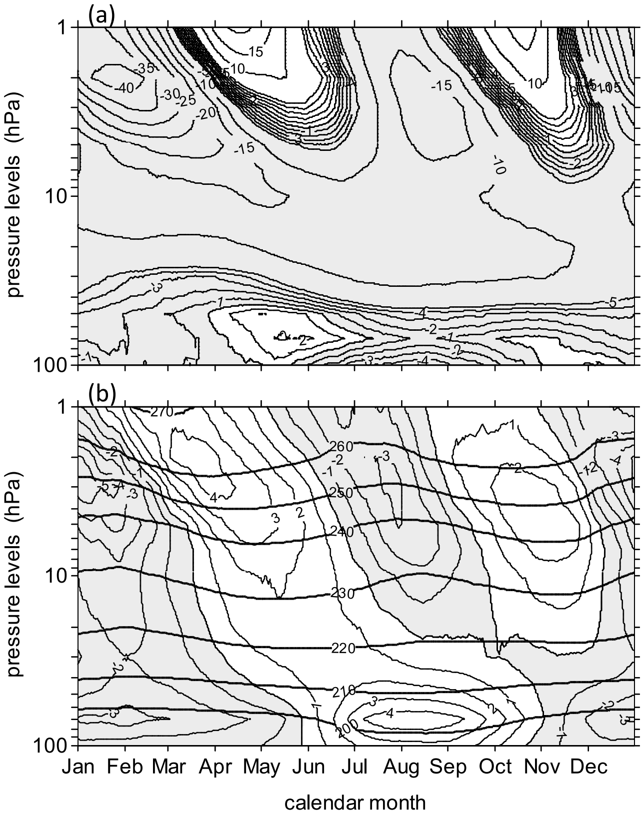

Figure 10. (a) Vertical profile of tropical (5°S–5°N) climatological mean annual pattern of zonal mean zonal wind in m s^{-1}. Easterly winds are shaded, and westerly winds are not shaded. (b) Vertical profile of tropical climatological mean annual variation of temperature (thick black lines) and temperature anomalies from annual mean profile in K. Negative anomalies are gray shaded.

the more irregular behavior of the eQBO shear zone during SH winter and suggested that changes in tropical waves could be involved. A possible modulation of the QBO by solar wind dynamic pressure may provide a partial explanation for their observation. Our results on solar wind dynamic pressure–related perturbation of the QBO add another element to this general picture of the QBO variability mechanisms. As tropical upwelling is generally lower during the austral winter and spring, a seasonal temperature perturbation linked to high P_{sw} may create an environment that favors significantly more easterly shear zone to descend from higher altitudes into the lower stratosphere and/or a slower descent of the QBO overall.

[35] It is worth noting that the seasonal temperature change that we find to be related to solar wind dynamic pressure in the stratosphere is ~1 K in the lower stratosphere and ~2 K in the upper stratosphere. These values are comparable to the interannual variability of tropical temperature associated with ENSO, volcanic eruptions, and the QBO [*Randel et al.*, 2004; *Fueglistaler et al.*, 2009].

[36] Our analysis regarding inclusion or exclusion of the data affected by the major volcanic eruptions or the major El Niño events showed no obvious bias on the statistically significant modulation of P_{sw} on the QBO. Nevertheless, more detailed study is needed to separate the effects of volcanoes, the ENSO, solar UV, and solar wind. One obvious method is multivariate regression. However, the main drawback of such analysis is that these mechanisms may not be mutually exclusive, and they could act together in a nonlinear fashion to either enhance or cancel the overall responses. In addition, multivariate regression effectively reduces the degree of the freedom in the data and results in statistically less significant results. Thus, it remains technically challenging to separate the different origin or cause on the basis of limited observational data, especially when the interaction is nonlinear between the regressive variables.

[37] *Lu et al.* [2008a] showed that the winter Northern Annular Mode is positively correlated with P_{sw} when the 11 year SC is at its maximum phase, while negative correlation between P_{sw} and the stratospheric NAM exists in spring at solar minimum. Those extratropical signals suggested a weakening of the BD circulation during winter months with reduced upwelling in the low-latitude lower stratosphere when both solar activity and solar wind dynamic pressure are high. The reversed relationship in spring implies enhanced BD circulation and anomalous stronger upwelling at low latitudes. Our findings here (see Figure 9) further confirm a seasonal modulation of the BD circulation, reflected by a significant increase of tropical lower stratospheric temperature, which corresponds to an anomalously weaker upwelling near the tropics. It is also consistent with earlier findings of *Lu et al.* [2007] based on the geomagnetic *Ap* index.

[38] It should be noted that distinct differences exist in the extratropical and tropical responses. First, the tropical response reported here does not change sign between winter and spring as opposed to the P_{sw} projection onto the NAM, which tends to switch from positive to negative from NH winter to spring [*Lu et al.*, 2008a]. Second, the tropical P_{sw} signals are independent of the 11 year SC, while conversely, the extratopical signals are only statistically significant once the data are separated into high and low solar activity. Finally, a significant response of the QBO to the P_{sw} variation is detected during SH late winter and spring rather than during NH winter, when the most robust P_{sw} versus NAM relationship holds. Hence, there is no clear evidence to suggest that the tropical response is a direct result of the NH extratropical response reported previously by *Lu et al.* [2008a]. This implies that a change in extratropical wave forcing alone is insufficient to explain the signals observed in both the NH polar region during the boreal winter and spring and the equatorial region during the austral winter and spring. There must be either extra or different mechanisms at play to cause the incompatible signals in the extratropical NH winter.

5. Summary

[39] By using daily and monthly data extending from 1963 to 2009, we have revealed a significant solar wind dynamics pressure (P_{sw}) relationship to the annual cycle of temperature in the tropical tropopause region, which in turn affects the QBO during austral late winter and spring. The main characteristics of this relationship can be summarized as follows.

[40] 1. Significant changes in both the strength and phase occurrence frequency of the QBO in the lower stratosphere

	Westerly->Easterly												Easterly->Westerly											
10 hPa										06														
										04														
										90							03							
										80							98						00	
									85	75							94						81	
						02			69	61							84	07			05		76	
			88		95	99		08	66	59							68	96	70	79	91	72	62	
	64		83	93	78	73		97	57	55					87		54	89	56	74	60	65	58	
	J	F	M	A	M	J	J	A	S	O	N	D	J	F	M	A	M	J	J	A	S	O	N	D
20 hPa																							07	
											06												05	
					83						04	97									03		89	
			09		67						02	85								96	98		79	
			86	91	64	93				95	99	61	86	75			92	87	01	68	84	96	72	
	58	60	72	88	53	76			78	73	69	55	66	59			63	82	77	54	60	70	56	
	J	F	M	A	M	J	J	A	S	O	N	D	J	F	M	A	M	J	J	A	S	O	N	D
30 hPa													04											
													97											
			07		00								80	08					92					
		74	05		91	09						88	73	90					87					
	03	65	79	62	81	86	83					69	71	69		75			82				01	98
	98	56	60	58	72	67	76			93	95	53	57	61	06	66	59		63		77	94	54	84
	J	F	M	A	M	J	J	A	S	O	N	D	J	F	M	A	M	J	J	A	S	O	N	D
50 hPa						07																		
						03																		
				98	05	91											08	90						
				96	70	89											06	80						
				94	65	79									02	04	97	69						
			01	84	60	74	62								99	85	71	66	75			87		
	77	87	68	54	56	72	58					81	95	78	55	73	61	57	59		63	82		92
	J	F	M	A	M	J	J	A	S	O	N	D	J	F	M	A	M	J	J	A	S	O	N	D
70 hPa																								
						98			05															
						96			79															
						84	01		74										07			75		
			92			78	73		71	07	08						99	06	85			66		
			82	87		70	72		65	89	03			93			95	97	78	90		59	87	91
	63		77	68	94	60	56		54	58	91		09	55	53		73	69	71	80	61	57	63	82
	J	F	M	A	M	J	J	A	S	O	N	D	J	F	M	A	M	J	J	A	S	O	N	D

Figure 11. Histograms of the QBO phase transitions (zero crossings based on daily data) at (top to bottom) 10, 20, 30, 50, and 70 hPa grouped by month. Westerly to easterly transitions are displayed on the left, while easterly to westerly transitions are shown on the right. The climatology and the variability faster than 6 months were both removed before the transitions were determined. The years of the transitions are denoted by two digital numbers.

are found in the lower stratosphere in relation to the solar wind dynamic pressure. The signature is manifested by stronger and more frequent occurrence of easterly anomalies associated with high P_{sw} during July–October in the lower stratosphere. The effect is related to much higher frequency variations of P_{sw} than the 11 year solar cycle, and the most significant response is found at 30–50 hPa.

[41] 2. The annual averaged vertical profile of the tropical zonal wind anomalies caused by P_{sw} perturbation is characterized by a vertical three-cell anomaly pattern with westerly anomalies in the troposphere, easterly anomalies at 20–70 hPa, and westerly anomalies at 3–10 hPa. Despite its smaller amplitude in comparison to those obtained from austral winter and spring months, this well-structured wind anomaly pattern is statistically significant all year around. There is additional seasonal variation superimposed on this annual average with noticeable upward movement during the boreal winter and spring and downward movement during the austral late winter and spring, consistent with the known annual cycle effect with stronger upwelling during NH winter and weaker upwelling during SH winter.

[42] 3. The tropical temperature response to P_{sw} is characterized by anomalous warming of 2 K at 30–50 hPa and up to 1 K near the tropopause during the boreal winter and spring accompanied by up to 0.5 K cooling in the troposphere during the austral winter and spring. There is an anomalous downward propagation of positive temperature anomalies from the upper stratosphere to the lower stratosphere over a period of a year starting from December. The combined effect may cause a systematic modulation of the annual cycle in the tropical stratospheric temperature and wind.

[43] It remains to be understood what mechanism can cause the observed seasonal modulation of the annual cycle by solar wind dynamic pressure leading to statistically significant changes of the QBO. As the annual cycle in tropical temperature and zonal wind is strongly coupled to the annual cycle in stratospheric water vapor, ozone, and wave activity from the troposphere [*Mote et al.*, 1995; *Fueglistaler et al.*, 2009], a next step would be to look for changes in convective water vapor, ozone, and/or upward propagating equatorial waves. Through such related research, the pathways by which the effect of solar wind variability, which is observable mainly in the lower thermosphere and above, may propagate to the lower atmosphere can be better understood.

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