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Decadal-scale changes in the effect of the QBO on the northern stratospheric polar vortex

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[1] This study documents decadal-scale changes in the Holton and Tan (HT) relationship, i.e., the influence of the lower stratospheric equatorial quasi-biennial oscillation (QBO) on the northern hemisphere (NH) extratropical circulation. Using a combination of ECMWF ERA-40 Reanalysis and Operational data from 1958–2006, we find that the Arctic stratosphere is indeed warmer under easterly QBO and colder under westerly QBO. During November to January, composite easterly minus westerly QBO signals in zonal wind extend from the lower stratosphere to the upper stratosphere and are centered at $\sim$5 hPa, 55–65°N with a magnitude of $\sim$10 m s^{-1}. In temperature, the maximum signal is near $\sim$20–30 hPa at the pole with a magnitude of $\sim$4 K. During winter, the dominant feature is a poleward and downward transfer of wind and temperature anomalies from the midlatitude upper stratosphere to the high latitude lower stratosphere. For the first time, a statistically significant decadal scale change of the HT relationship during 1977–1997 is diagnosed. The main feature of the change is that the extratropical QBO signals reverse sign in late winter, resulting in fewer and delayed major stratospheric sudden warmings (SSWs), which occurred more often under westerly QBO. Consistent with earlier studies, it is found that the HT relationship is significantly stronger under solar minima overall, but the solar cycle does not appear to be the primary cause for the detected decadal-scale change. Possible mechanisms related to changes in planetary wave forcing are discussed.

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1. Introduction

[2] The quasi-biennial oscillation (QBO) in the equatorial stratosphere consists of descending alternating westerly and easterly winds with an average cycle length of $\sim$28 months, varying from 22 months to 34 months [*Reed*, 1965; *Naujokat*, 1986]. Wind direction reversals start near 3 hPa ($\sim$40 km), and from this level the QBO phase fronts descend to a dissipation level near 18 km altitude ($\sim$90 hPa) with a descent rate of $\sim$1–2 km/month [*Baldwin and Dunkerton*, 1998; *Giorgetta et al.*, 2002]. The oscillation peaks at the equator with a latitudinal half-width of $\sim$12° [*Reed*, 1965]. At individual stations, the QBO has a peak-to-peak amplitude of $\sim$55 m s^{-1} at about 15 hPa, while in the lower stratosphere at 40–50 hPa, the maximum amplitude is about 30 m s^{-1} [*Baldwin and Gray*, 2005].

[3] Studies have been undertaken to understand the QBO influences on the extra-tropical circulation in the stratosphere [*Angell and Korshover*, 1970, 1975; *Dunkerton and Baldwin*, 1991; *Holton and Austin*, 1991; *Baldwin and Dunkerton*, 1998; *Baldwin et al.*, 2001; *Gray et al.*, 2001a; *Naito et al.*, 2003]. Observational studies tend to suggest a colder and stronger polar vortex during westerly QBO, warmer and a more disturbed polar vortex during easterly QBO. *Holton and Tan* [1980, 1982] proposed a mechanism to explain the phenomenon in terms of planetary wave propagation that is guided by the zero-wind line in the winter subtropics [*Matsuno*, 1970, 1971; *Tung and Lindzen*, 1979]. The rationale is that the vertically propagating stationary planetary waves from the extratropical troposphere can only propagate through a waveguide of westerly wind. During the easterly QBO phase, the zero-wind line moves to the subtropics of the winter hemisphere. It narrows the width of the planetary waveguide in the extratropical lower stratosphere. The narrowed waveguide leads to refraction of the planetary waves away from the subtropical region and consequent redirection poleward. When such wave events with large amplitudes break or dissipate, the resulting additional wave drag slows the polar vortex and warms the polar stratosphere. In extreme cases, they cause major stratospheric sudden warmings (SSWs) [*McIntyre*, 1982]. Conversely, if the QBO is westerly, the waves are less restricted latitudinally, resulting in an anomalously colder and less disturbed polar vortex. Thus fewer major SSWs occur during the westerly phase of the QBO. Such QBO influence on the polar vortex through planetary wave propagation is commonly referred to as the Holton-Tan (HT) mechanism. The corresponding statistical

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correlations between the stratospheric equatorial wind and the extratropical temperature or wind are referred to as the *HT relationship* hereafter.

[4] It has been shown that the HT relationship with "warm disturbed easterly phase" and "cold undisturbed westerly phase" does not always hold up [*Hamilton*, 1998]. *Labitzke and van Loon* [1988] have suggested that the periods when the HT relationship holds up coincide well with when the 11-year solar cycle is at its minimum but the relationship reverses during solar maximum, although *Gray et al.* [2001b] found that the relationship substantially weakened during solar maximum but that there was no evidence for a reversal. By using National Meteorological Center (NMC) data from 1962–1994, *Naito and Hirota* [1997] found that the HT relationship is statistically significant only for the period of 1962–1977; for the entire period of 1962–1994, the HT relationship was statistically significant only in early winter (i.e., November and December). *Naito and Hirota* [1997] and *Gray et al.* [2001b] attributed the stronger HT relationship in pre-1977 period to relatively weak solar modulation during the time as it covers one solar maximum and two solar minima, while 1978–1994 period has two solar maxima and one solar minimum. They concluded that the HT relationship holds in early winter and the solar cycle modifies its influence in late winter. By using NCEP/NCAR reanalysis data from 1952 to 2001 at 50 hPa, *Hu and Tung* [2002] repeated the original analysis of *Holton and Tan* [1980]. Similar to *Naito and Hirota* [1997], they found that the HT relationship remains valid in early winter but fails to hold in late winter. However, they concluded that the solar cycle has little effect on wave amplitudes in the lower stratosphere, implying that the solar cycle may not be the responsible factor for the substantially weakened HT relationship in late winter. By using ERA-40 zonal wind from 1979–2001, *Gray et al.* [2004] found that, although easterly anomalies in the extratropical stratosphere are more likely to be associated with the easterly QBO, the only month in which the HT relationship remains significant at the 95% confidence level is November.

[5] The discrepancies in the HT relationship were previously attributed to the combination of large interannual variability in the polar region, volcanic eruptions, an overlapping signal with ENSO, the brevity of data and/or limited height range of data sets [*Baldwin et al.*, 2001]. Nevertheless, the evidence in the literature strongly suggests that early and late winter QBO signals in the extratropical temperature are different from each other [*Holton and Tan*, 1980; *Naito and Hirota*, 1997; *Hu and Tung*, 2002]. In addition, there may be structural changes in the HT relationship during last 50 years and the 11-year solar cycle might not be the only modulating factor. However, investigation into other possible modification of the HT relationship is yet to be undertaken.

[6] Given that the largest extratropical QBO signals are found in the NH winter [*Baldwin and Dunkerton*, 1998], this study makes a statistical assessment of the decadal-scale variation of the HT relationship in the NH winter. We aim to address three main objectives: (1) to determine if the HT relationship holds up for the period of 1958–2006; (2) to examine whether the HT relationship has weakened or reversed during that period, and (3) to study whether and how the 11-year solar cycle may relate to the decadal-scale changes of the HT relationship.

2. Data and Methods

[7] Our analysis uses monthly mean wind and temperatures from ECMWF (European Centre for Medium Range Weather Forecasting) ERA-40 Reanalysis (September 1957 to August 2002) and ECMWF Operational analyses (September 2002 to December 2006).

[8] The ERA-40 Reanalysis used ECMWF's 3D variational data assimilation system with a spectral resolution of T159, corresponding to a 1.125° horizontal resolution in latitude and longitude. The model had 60 levels in the vertical between the surface and 0.1 hPa (~65 km). Only the 23 standard pressure surfaces from 1000 hPa to 1 hPa are used here. The data in this height range were assimilated using direct radiosonde and satellite measurements (when available), while the data beyond 1 hPa represent primarily the modeled results [*Uppala et al.*, 2005]. *Baldwin and Gray* [2005] showed that the QBO extracted from ERA-40 is consistent with rocketsonde winds (that were not assimilated) measured at Ascension and Kwajalein up to 2–3 hPa, even for those years before satellite era.

[9] The ECMWF Operational data used ECMWF's 4D variational data assimilation scheme and were output from the ongoing analyses produced by the ECMWF Integrated Forecasting System (IFS) model. The most recent IFS uses T799 horizontal resolution and 91 vertical levels to 0.01 hPa. The data from September 2002 to the present-day are available on the same 1.125° grid and on 21 pressure levels, identical to the ERA-40 data except without the 600 and 775 hPa levels. Rather than using 21 levels for the entire analysis period, we chose to linearly interpolate the Operational data to these missing levels after the data were zonally averaged.

[10] Our correlation and compositing analyses are all carried out using deseasonalized zonal-mean zonal wind and temperature. The deseasonalization is carried out by taking monthly mean values based on the records for the entire period of 1958–2006. Changes in the availability of satellite data and the operational assimilation system, beginning in 2002, mean that the data record, especially near 1 hPa, may have varied in quality.

[11] The 50 hPa zonal wind is chosen to define the phase of the equatorial QBO, in order to be consistent with *Holton and Tan* [1980]. In this study, the equatorial zonal winds are extracted at 0.56 °N, 50 hPa from the combined ERA-40 and Operational records. In fact, 50 hPa winds are the closest to the optimum level suggested by *Baldwin and Dunkerton* [1998], who found that the largest QBO signals were obtained with a QBO defined at 40–50 hPa for the NH, and 20–30 hPa in the SH. The westerly and easterly phases were defined as the deseasonalised monthly zonal-mean zonal wind ≥ 2 m s^{-1} and ≤ -2 m s^{-1}, and are hereafter referred to as w-QBO and e-QBO, respectively. However, we found that other choices of threshold values in the range of ± 0–5 m s^{-1} produce similar outcomes, though smaller thresholds result in slightly smaller correlation coefficients while larger thresholds result in higher correlation but with limited numbers of sample points.

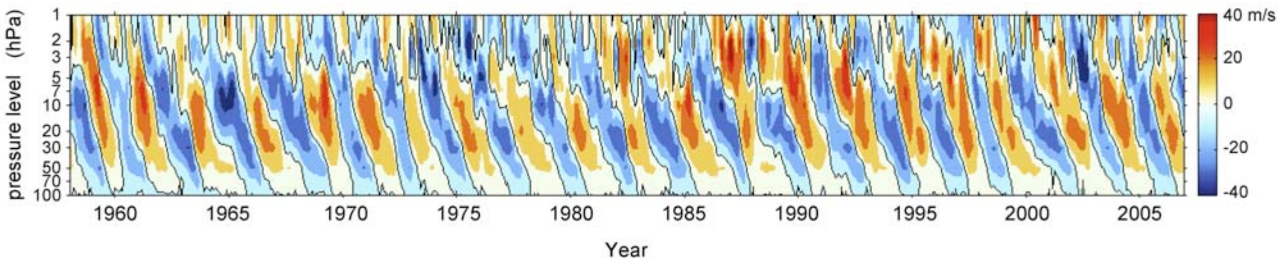

Figure 1. The QBO time series (i.e., de-seasonalized monthly mean zonal mean zonal wind at the equator) in m s^{-1} extracted from ERA-40 and ECWMF operational on 11 pressure levels from 1958 to 2006. Red and blue colors represent westerly and easterly winds. The black solid lines are zero wind lines.

[12] A list of the major SSW events occurring during 1958–2001 was compiled recently by *Charlton and Polvani* [2007]. In addition, three major SSW events that occurred after 2001 and before 2007 are identified using the same criteria, i.e., January, 2003, January, 2004, and January, 2006 [Charlton, personal communication, 2007]. All these events are used here to examine whether the e-QBO indeed favors the occurrence of major warmings, as proposed by *McIntyre* [1982], or whether such a QBO-phase versus SSW relationship has changed when examined over a longer duration data period.

[13] During 1958–2006, major perturbations to the stratospheric circulation resulted from three major volcanic eruptions; Agung in March, 1963, El Chichón in March, 1982 and Pinatubo in June, 1991. To avoid contamination by the anomalous warming caused by volcanic aerosols in the stratosphere, the 24 months following these major eruptions are excluded from the spatial analysis. However, in order to maintain the temporal continuity in the analyses of the individual time series, these volcanically contaminated data are kept in the analyses presented in section 3.2. Nevertheless, we found that the overall results are not sensitive to keeping or removing the 24 months data after volcanic eruptions.

[14] The main diagnostic tools employed in this study are correlation studies and composite analysis. Details around statistical significant tests used are given in Appendix A.

3. Results

3.1. QBO Composites During Early and Late Winter

[15] Figure 1 shows the QBO time series at 10 pressure levels from 100 to 1 hPa and from 1958–2006. The descending alternating easterly and westerly winds are the dominant feature of the QBO, particularly between 5–70 hPa, with maximum amplitude at $\sim$40 m s^{-1}. The difference in descent rates of the easterly and westerly shear zones can be clearly seen. A comparison with the corresponding time series of single station radiosonde data [*Naujokat*, 1986] and rocketsonde data [*Gray et al.*, 2001b] suggests that the QBO is captured exceedingly well up to 2–3 hPa, even in the pre-satellite era in the ERA-40 data [*Baldwin and Gray*, 2005]. Figure 1 shows that enhanced westerly anomalies appear above 10 hPa during the second half of the 1980s and much of the 1990s, suggesting a possible structural change of the QBO in the upper stratosphere. This agrees with what has been reported by *Punge and Giorgetta* [2007]. A closer examination of the change suggests that the QBO switched to easterly soon after September 1978. It then switched back to westerly around June, 1981; a discernible enhancement of westerly anomalies has occurred since then. Thus it is not clear if those enhanced westerly anomalies seen in the upper stratosphere were a result of data assimilation pre/post-satellite era.

[16] Figure 2 shows the composites of zonal-mean zonal wind (lined contours) and temperature (color shaded) for e-QBO (left-hand panels) and w-QBO (right-hand panels), for November to January averaged anomalies (top panels), and February to March averaged anomalies (bottom panels) for the entire period of 1958–2006. Note that using averaged anomalies for any consecutive two months from November to January may produce similar spatial patterns to those shown in the top panels of Figure 2. It shows, in early to middle winter (i.e., November to January) and under e-QBO, the largest easterly anomalies found at $\sim$60°N, 5 hPa in the extratropical stratosphere are associated with a generally warmer polar vortex extending from 3 hPa to 300 hPa with the largest temperature anomaly located at $\sim$90°N, 30 hPa. The opposite holds for the w-QBO composites, in which a stronger, colder stratospheric vortex exists. On average, the extratropical stratospheric winds are up to $\sim$10 m s^{-1} stronger under w-QBO than under e-QBO, while the Arctic stratospheric temperature is up to 3–4 K warmer under e-QBO than under w-QBO.

[17] During late winter (i.e., February and March) and under e-QBO, the easterly anomaly has moved slightly poleward and downward and the largest anomaly is located near 65°N, 20 hPa. Warming remains below 30 hPa with its center located at $\sim$90°N, 150 hPa. From 60°N poleward, while the lower stratosphere is warmer, the Arctic upper stratosphere is noticeably cooler. The opposite spatial pattern holds under w-QBO. Thus the dominant feature of the late winter HT relationship is a downward movement of westerly or easterly winds from the midlatitude upper stratosphere, and cooling or warming (dependent upon QBO phase) in the Arctic upper stratosphere with an opposite signed cell directly below. Consequently, the HT relationship evolves from early winter to late winter. At a single fixed level, the HT relationship appears to change through the winter, as the circulation and temperature anomalies descend through that level. In general, we found that the HT relationship appears to be strongest during

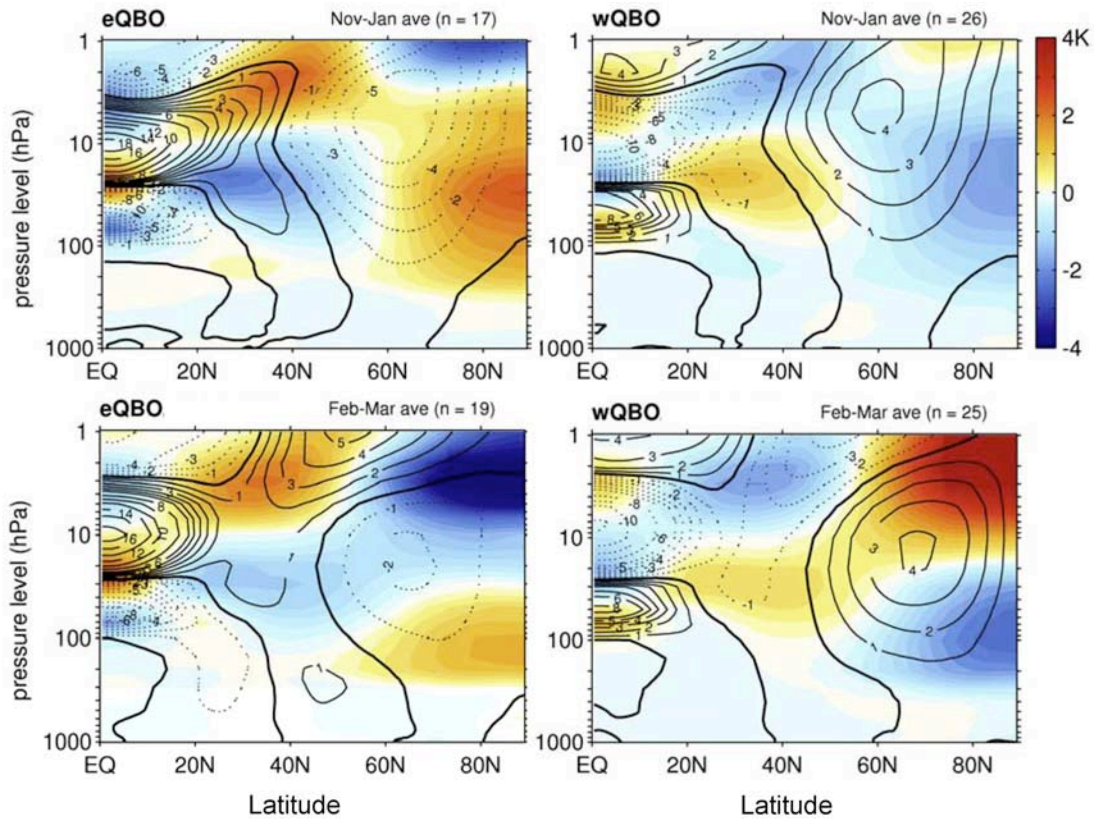

Figure 2. e-QBO (left-hand panels) and w-QBO (right-hand panels) composites of the zonal mean zonal wind (lined contours) and temperature anomalies (color shaded contours) for November to January mean (top panels), and for February and March mean (bottom panels). Thick solid lines represent zero wind. The QBO phases, the calendar months, with the total number of data samples indicated on the top of each panel.

November to December and descent during February to March.

3.2. Temporal Variations of the QBO and Its Extratropical QBO Signals

[18] In this section, we report a decadal-scale change in the correlation between the equatorial QBO zonal wind and both the extratropical wind and temperature records. Time series were extracted from three representative locations in the Arctic stratosphere. Note that similar statistical results can be obtained if the time series are extracted from nearby locations within $\sim 10^\circ - 15^\circ$ in latitude and $\sim 5 - 10$ km in altitude.

[19] Figure 3 shows the November to March averaged time series of (a) the zonal wind anomaly at 54.4°N, 10 hPa, (b) the temperature anomaly at 65.5°N, 50 hPa and (c) the temperature anomaly at 65.5°N, 200 hPa. On each of the plots the November-March averaged equatorial QBO has been superimposed (gray lines). For the whole period 1958–2006, the correlation coefficient (r) between the equatorial QBO and the extratropical wind is 0.64, and the correlation coefficients between the equatorial QBO and the two polar temperature time series are -0.50 and -0.57, respectively. These results show that in the NH the vortex is generally stronger and colder under w-QBO and weaker and warmer under e-QBO. The correlations are all significant at a confidence level above 99%, implying that the HT relationship holds for the averaged condition for the extended winter period. They suggest that about 30–40% of the extratropical variations in zonal wind and temperature for the extended winter may be explained by the HT relationship.

[20] There is a noticeable structural change in the QBO averaged over the extended winter period. From the mid-1970s to the mid- or late 1990s, the QBO preferentially appears in its westerly phase more than in its easterly phase. During 1975–1995, for instance, only 1979, 1984, and 1989 are clearly e-QBO. Such a structural change can also be demonstrated through its correlations with the extratropical time series. More specifically, we found substantially weakened correlations during 1977–1997 with $|r| < 0.3$ and confidence levels below 80%. In both wind and temperature, the high correlations evident in the first and last subperiods are not present during 1977–1997. These correlation changes appear to have affected a large vertical range.

[21] Figure 4 shows the same as Figure 3 but here the averages are taken for February and March only. Though the structural change in the equatorial QBO itself becomes less clear, the differences between the correlation coefficients and confidence levels for the first and last subperiods and for the midperiod of 1977–1997 become even larger. For instance, in both 1958–1976 and 1998–2006 periods, the correlation coefficients are positive for wind and negative for temperature while the reverse is true for 1977–1997. The correlations are statistically significant at confi-

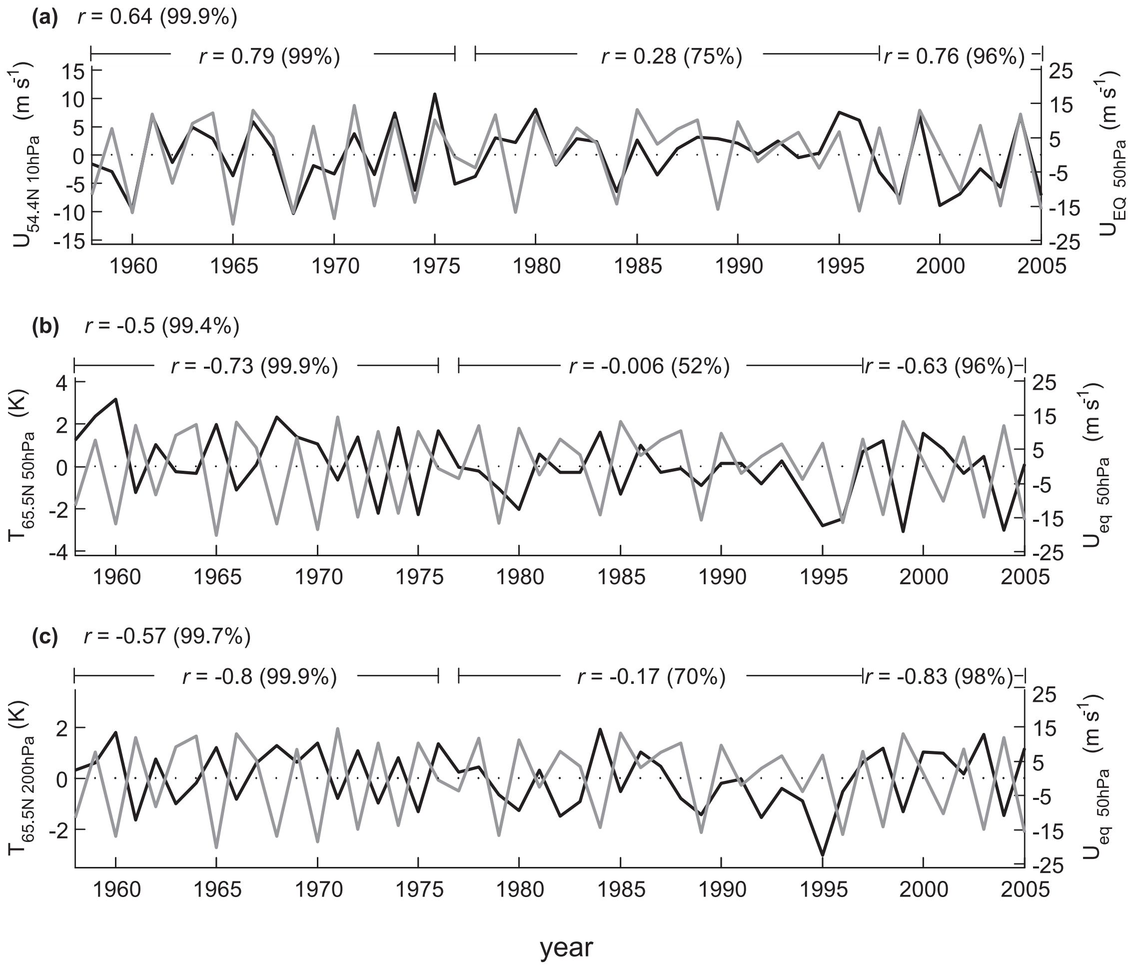

Figure 3. Time series (black lines) of the extratropical zonal-mean (a) zonal wind at 54.4°N, 10 hPa and (b) and (c) temperature anomalies at 65.5°N, 50 hPa and at 65.5°N, 200 hPa, respectively, plotted against zonal-mean zonal wind at the equator at 50 hPa averaged over the extended winter period (November–March) (gray lines). The correlation coefficients and confidence levels (in brackets) for the entire period of 1958–2006 are given on the top of each sub-plot together with those for the sub-periods of 1958–1976, 1977–1997, and 1998–2006.

dence levels above 99% only in 1958–1976. We found that the subperiod differences in correlations associated with November to January months are considerably smaller (not shown). Thus late winter changes are primarily responsible for the substantially weakened HT relationship during 1977–1997. An additional new result is that, over the whole period of 1958–2006, the QBO signature in the extratropical temperature is actually stronger in the lowermost stratosphere (i.e., 200 hPa) than at its conventional height of 50 hPa.

[22] To demonstrate the magnitude of the change more systematically, running correlation coefficients for the all cases shown in Figures 2 and 3 over blocks of data spanning 21 years are computed and are shown in Figure 5. The 21-year length of the running window is chosen to ensure a degree of statistical stability in the estimated correlations, while providing some localization in time. We verified that qualitatively similar results are obtainable by using a running correlation window from 17 to 25 years.

[23] The left-hand panels of Figure 5 show the running correlation coefficients for the three cases where the data are averaged over November to March. In each panel, the range (maximum–minimum) of correlation coefficients is 0.56, 0.75, and 0.64, respectively. The right-hand panels of Figure 5 show that a similar change in the HT relationship is maintained for February-March averages. The ranges are 0.94, 0.81, and 0.81, respectively, which are larger than those associated with November-March averages. Around 1985, the correlation changes from positive to negative for the winds and from negative to positive for the temperatures. All six running correlation curves peak or trough around 1987 and share an essentially similar shape.

[24] The hypothesis of a systematic temporal change in the HT relationship needs to be tested objectively. The null hypothesis we assume here is that there is no temporal structure change in the HT relationship and the observed variations in the running correlation shown in Figure 5 are merely due to statistical fluctuations. First of all, using the method proposed by *Neumaier and Schneider* [2001], our fitting results suggest that good models for both the equatorial QBO and the NH extratropical time series are stochastic AR1 processes. It allows us to perform the

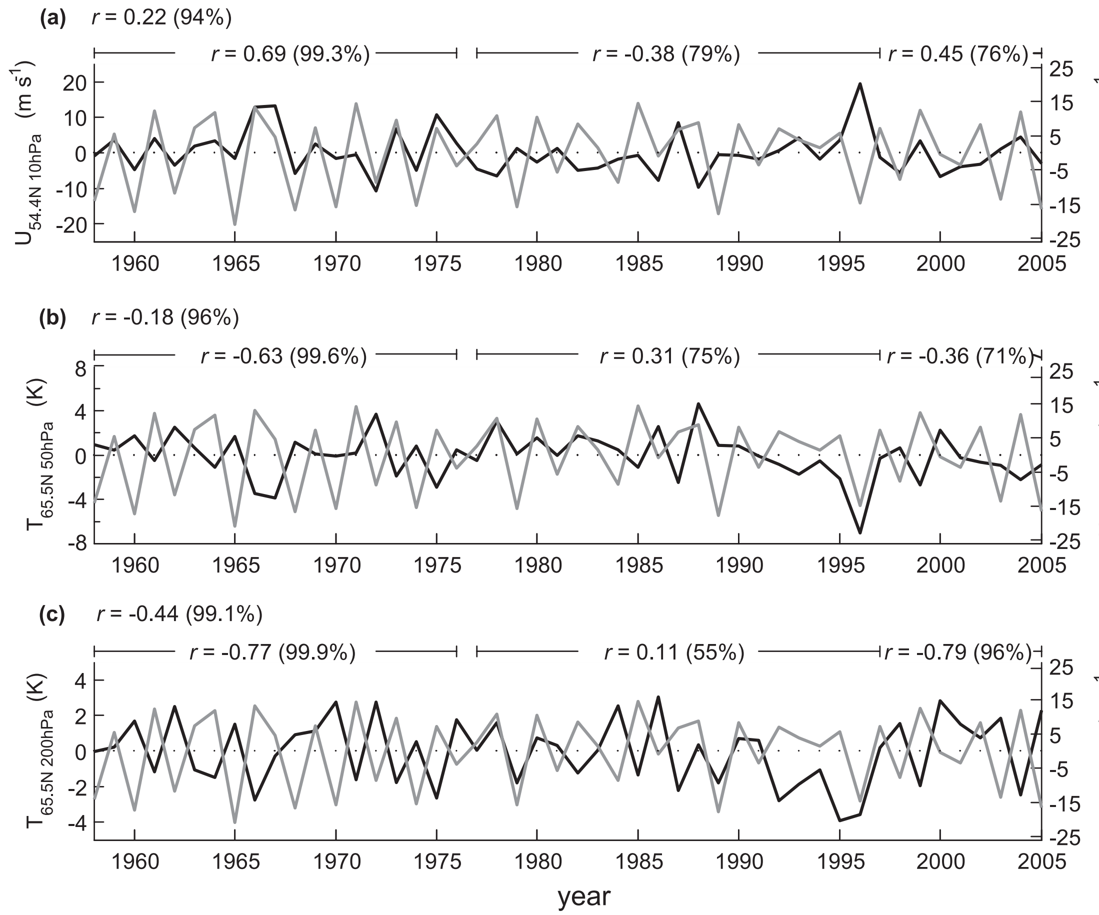

Figure 4. Same as Figure 3 but for late winter (February–March) averages.

significance test proposed by *Wang et al.* [2006] and a *p*-value is generated for each correlation case. In relation to the null hypotheses under consideration here, the *p*-value is large if the observed range of running correlation could reasonably occur by chance, and small if the observed range is unlikely to occur by chance. A *p* value of 0.05 corresponds to the 95% confidence level.

[25] The *p*-values for the differences in running correlations between the QBO and extratropical wind and temperature anomalies are 0.201, 0.065, and 0.106 for the cases of November-March mean. None of these results is statistically significance at the 95% level. The February-March mean results are 0.014, 0.046, and 0.045, respectively. The evidence for the late winter mean is stronger across all three time series as the *p*-values are all below 0.05, allowing us to reject the null hypothesis with greater than 95% confidence.

[26] Though we find here that the decadal-scale changes in the HT relationship during late winter are statistically significant, it still leaves room for other factors, such as the effect of internal atmospheric variability or the 11-yr solar cycle, to cause temporal structured fluctuations in the HT relationship. Ultimately, longer data series will provide a more concrete conclusion.

3.3. Seasonal Progressions of Composite Differences

[27] In this section, the spatial and seasonal variations of the HT relationship are studied to determine whether or not the variation in HT relationship shown in section 3.2 is a property that applies only to certain preselected locations. Latitude/height composite analyses are performed for the entire period of 1958–2006 first. The composite analysis for the first and last subperiods (i.e., 1958–1976 and 1998–2006) is then performed as a single data group, and the results are compared with those for the middle period (1977–1997). We chose to use January of 1977 and January of 1998 to separate the subperiods as we found that the greatest overall reduction in both correlation and statistical significance occurs when the data are separated in such a way. The results are shown in Figures 6 and 7 below.

[28] Figure 6 shows the seasonal progression of the QBO composite differences (i.e., e-QBO minus w-QBO) of the zonal-mean zonal wind anomalies in vertical-meridional cross section from October to April for the NH. The 1st, 2nd, and 3rd columns are for the whole period (1958–2006), the combined two end subperiods (1958–1976 & 1998–2006) and the middle period (1977–1997), respectively. The areas within the gray contours represent statistical significances at confidence levels equal or greater than 95%.

[29] For the period of 1958–2006 (1st column), a three-cell vertical structure of the QBO, similar to that shown by *Pascoe et al.* [2005], is evident in the tropics. From October to March, negative wind differences can be observed in the Arctic stratosphere, implying a more disturbed polar vortex with persisting easterly equatorial anomalies near 50 hPa. The data suggest that the easterly anomalies first originate in

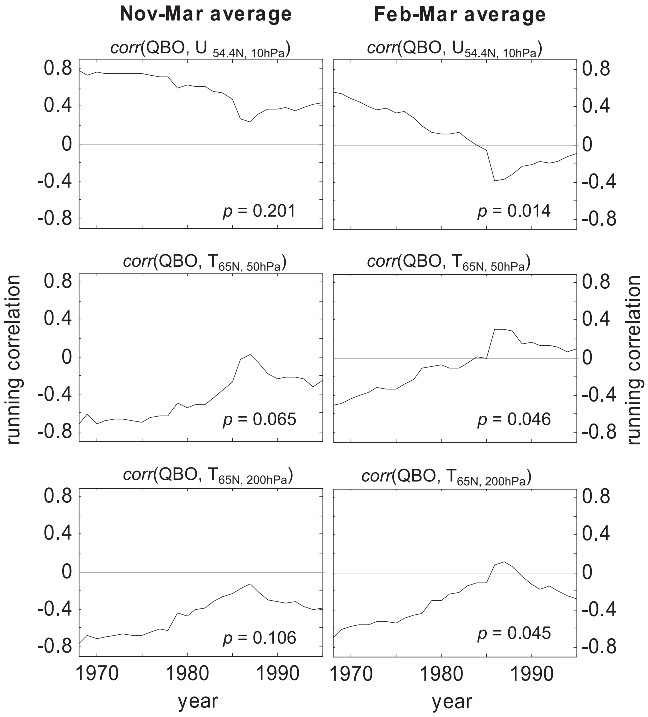

Figure 5. Running correlation coefficients between the equatorial QBO and the NH polar wind anomaly (first row) & temperature anomalies (second and third rows) with a 21-year window, using November to March mean (left-hand panels) and February to March mean (right-hand panels). The time series of the wind and temperature anomalies are the same as those shown in Figures 3 and 4. The *p* values represent the levels of statistical significance in terms of the difference between the maximum and minimum values of the running correlation. See text for details.

the upper stratospheric subtropics at 22°N, 3–5 hPa during September (not shown). From November to January, these easterly anomalies are enhanced and move poleward. Their downward movement starts in late winter from February to March. During those movements, the center of the easterly anomalies shifts slightly from 50–55°N, 1–5 hPa to 55–65°N, 5–10 hPa. By April, the easterly anomalies appear to have descended into the troposphere. These QBO composite differences primarily feature the characteristics of typical e-QBO composites, with a smaller contribution from w-QBO composites (not shown). Overall, the extratropical QBO signals are stronger in early winter (i.e., November–January) and weaker in mid to late winter (i.e., February–March), measured by total area in the extratropics enclosed by the gray lines, i.e., the 95% confidence levels of the signals.

[30] For the combined 1958–1976 & 1998–2006 periods (2nd column of Figure 6), the overall QBO signal patterns are quite similar to those during 1958–2006, but with much stronger easterly anomalies in the Arctic stratosphere. In comparison to those for 1958–2006, notably larger magnitudes of the easterly anomalies are observable in December and March. Statistically significant (at 95% confidence levels or above) composite differences can be found from November through March in the Arctic stratosphere, except for in January. Slight poleward and downward movement of the easterly anomalies can be observed as winter progresses.

[31] For the 1977–1997 period (3rd column of Figure 6), the extratropical QBO composite differences are distinctly different from those shown in the other two columns. Easterly anomalies appear in the extratropical stratosphere in November to January and reverse into westerly anomalies in February through April. This suggests a strengthened polar vortex under e-QBO which moves poleward and downward. During this period, the typical extratropical HT QBO signature can only be found in November.

[32] Figure 7 shows the corresponding QBO composite differences in temperature. During 1958–2006 (1st column of Figure 7), the tropical and midlatitude QBO signals show a well-defined pattern of the QBO-induced secondary circulation, which is connected by meridional and vertical positive/negative temperature anomaly cells. At midlatitudes, a three-cell structure in height is present with warming in the upper stratosphere (~4 K), cooling in the lower to mid stratosphere (~−3 K) and weaker warming in the lowermost stratosphere (~0.3 K). Those maxima are centered at heights of about 3 hPa (40 km), 30 hPa (25 km) and 200 hPa (11 km). The Arctic lower stratosphere is anomalously warmer by up to 4 K throughout the winter. During February, however, the mid to upper stratosphere (1–10 hPa, 65–90°N) is colder (by up to 13 K). Overall, positive QBO signals in temperature are found in the upper midlatitude stratosphere as well as the Artic lower stratosphere, while negative QBO signals are found at midlatitudes of the middle stratosphere.

[33] For the combined 1958–1976 & 1998–2006 periods (2nd column of Figure 7), the tropical to midlatitude QBO signals remain very much the same as those for 1958–2006. In the Arctic stratosphere, up to 10 K temperature differences can be observed during December. The positive QBO signals in the Arctic lower stratosphere are noticeably stronger during late winter.

[34] For the 1977–1997 period (3rd column of Figure 7), the tropical QBO signals are similar in sign to those for 1958–2006, as would be expected because of the way in which the e-QBO and w-QBO years are defined. However, the magnitudes of cooling in the tropical lower and upper stratospheres are ~1–2 K smaller and they are barely significant at a 90% confidence level. In early winter, the three-cell structure in the subtropics to midlatitudes remains, with warming in the upper stratosphere (15–40°N, 1–10 hPa), cooling in the lower to mid stratosphere (15–

Figure 6. Composite differences between e-QBO and w-QBO for zonal-mean zonal wind anomalies (m s^{-1}), for the period of 1958–2006, 1958–1976 & 1998–2006, 1977–1997 respectively for each of the months from October to April. The data have been first de-seasonalized using monthly mean data from 1958–2006, grouped into e-QBO, and w-QBO. The differences of the calculated mean values for each group are described in the text. The areas enclosed within the gray lines indicate that the differences are statistically significance from zero with a confidence level of 95% or above, calculated using a Monte Carlo trial based non-parametric test.

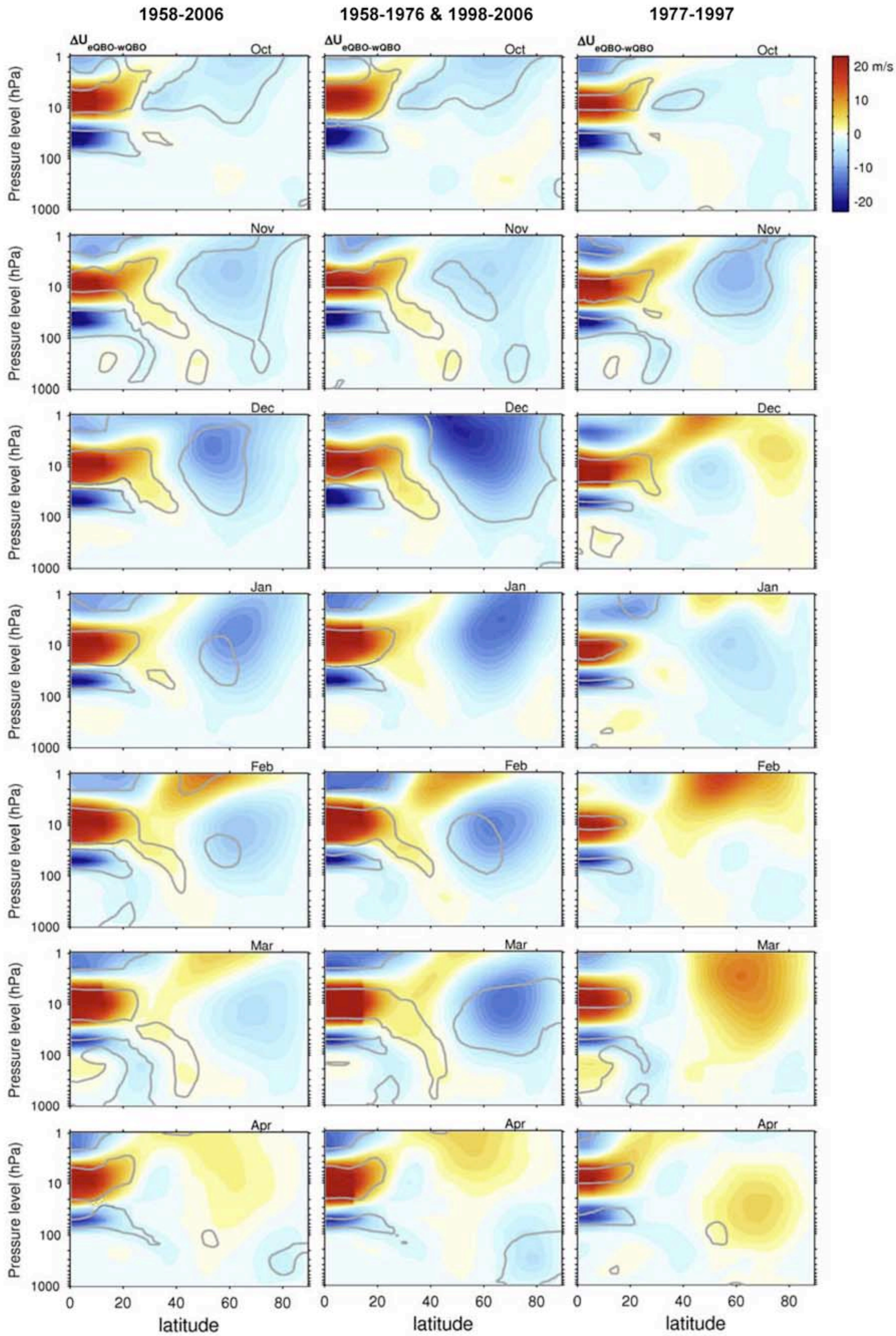
1958-2006
1958-1976 & 1998-2006
1977-1997
ΔU eQBO-wQBO
Oct
Nov
Dec
Jan
Feb
Mar
Apr
Pressure level (hPa)
1
10
100
1000
0
20
40
60
80
latitude
20 m/s
10
0
-10
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Figure 6

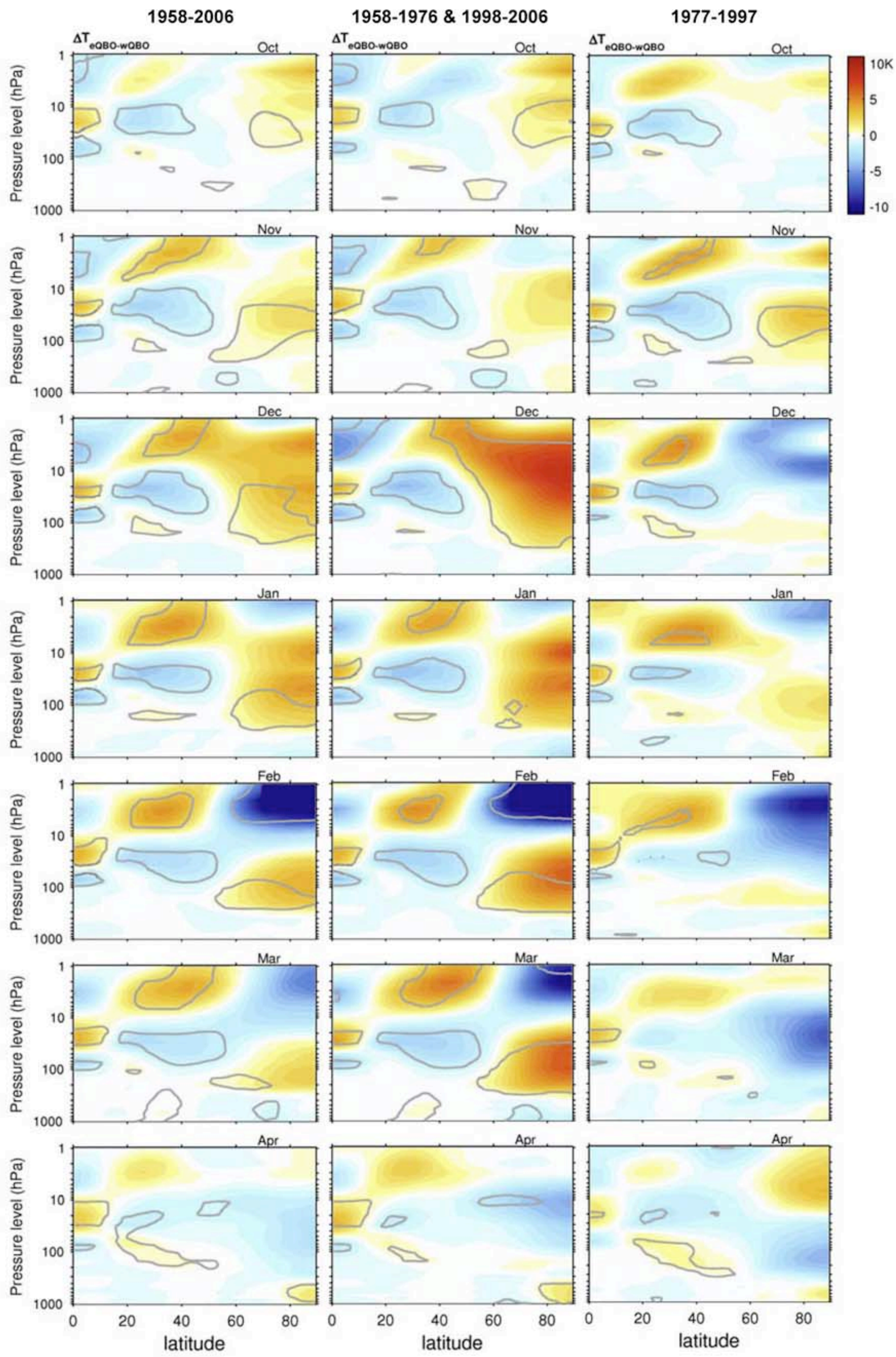

Figure 7. Same as Figure 6 but for the temperature anomalies.

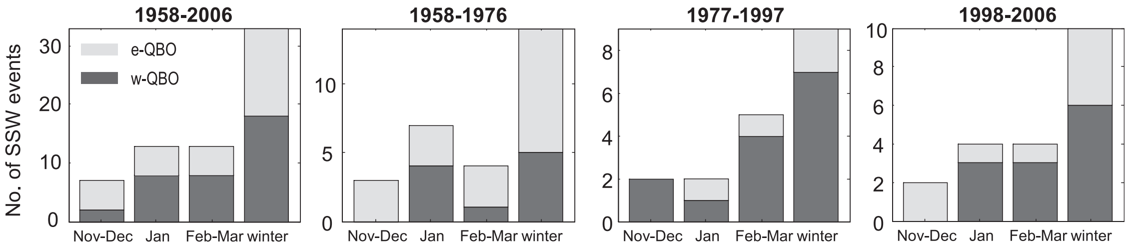

Figure 8. Major SSW occurrence grouped by early, middle, late and extended winter and the phases of the QBO, for the periods of 1958–2006, 1958–1976, 1977–1997, and 1998–2006 (from left to right). Westerly and easterly QBO are defined as >/< 0, respectively.

50°N, 10–100 hPa) and lesser warming in the lower most stratosphere (20–30°N, 100–200 hPa). Although the structure remains during late winter, it becomes substantially weakened and are not significant even at the 90% confidence level. In the extratropics, there is a suggestion of weak warming in February, which is only observable in the lowermost stratosphere of the Arctic. Cooling, on the other hand, is visible in the lower stratosphere during March.

[35] Theoretical studies have suggested that the QBO influence on temperature is a result of downward and upward motion associated with the QBO-induced meridional circulation [*Plumb and Bell*, 1982]. This is consistent with the warming and cooling structure resulting from the wave-induced transfer of easterly/westerly momentum to the mean flow in the high-latitude stratosphere [*Garcia*, 1987; *Gray and Pyle*, 1989]. The QBO-induced meridional circulation starts at the equator and is balanced by upward and downward motion in the subtropics and midlatitudes. Such a general pattern of dynamic influences of the QBO is shown clearly in most of the panels of Figures 6 and 7, except for those for 1977–1997, particularly during late winter. This suggests possible regime shifts around 1977 and 1997.

3.4. Decadal-Scale Change of the Major SSWs

[36] *McIntyre* [1982] suggested that the HT relationship may affect the likelihood of the SSWs. That is, the easterly winds present in the subtropics during e-QBO years may act as a wave barrier and hence may favor the occurrence of a major SSW. In this section, we examine the occurrence of SSWs to investigate whether a coherent temporal change in the occurrence frequency of major SSWs is evident.

[37] Figure 8 shows the distributions of major SSW events grouped for the periods of 1958–2006, 1958–1976, 1977–1997, and 1998–2006. In total, 33 SSW events occurred in 1958–2006, i.e., 33 events in 49 years, averaging 0.67 SSWs per year. In the whole period 1958–2006, there is no clear preference for either phase of the QBO; the occurrence ratio is 15:16 between e-QBO and w-QBO. For the three consecutive subperiods, the average number of major SSWs per year is 0.74 per year (14/19), 0.43 (9/21) and 1.11 (10/9), respectively. Thus the major SSWs have occurred less frequently during 1977–1997.

[38] Differences in occurrence patterns can also be observed when the major SSWs are grouped according to the phases of the QBO. For the entire period of 1958–2006, there are more major SSWs occurring under e-QBO in early winter but the opposite holds in mid to late winter. In 1958–1976, 9 out of 14 events occurred under e-QBO, reinforcing the concept of a strong influence of the HT relationship. However, during 1977–1997, only 2 out of 9 events occurred under e-QBO, suggesting either a weakened or even a reversed HT relationship. During this period, noticeably more events occurred in late winter rather than in midwinter, implying a significant delay in the timing of major SSWs. In 1998–2006, an approximately equal number of major SSWs occurred under the two QBO phases (i.e., 4:6), though more events occurred under e-QBO in early winter and under w-QBO in mid to late winter.

[39] It is not clear from this data set why the occurrence of SSWs appears to be influenced by the phase of the QBO in some periods but not in others. One possibility, suggested by *Dunkerton et al.* [1988] is that the existence of a deep layer of equatorial easterly anomalies or westerly anomalies is more important than easterly or westerly winds at any single level. More events (or modeling studies) are needed in order to investigate this.

3.5. 11-yr Solar Cycle Modulation

[40] Another possible influence on the frequency of SSWs is the 11-yr solar cycle (SC), which could influence SSWs directly by the presence of an 11-yr SC wind anomaly in the subtropical upper stratosphere [*Kodera and Kuroda*, 2002; *Gray et al.*, 2004] or indirectly by modifying the equatorial QBO [*Salby and Callaghan*, 2000; *McCormack*, 2003; *Pascoe et al.*, 2005]. Previous studies have shown that the strongest apparent SC modulation in the NH occurs in January and February [*Labitzke and van Loon*, 1988; *Labitzke et al.*, 2006]. In this section, using the extended data from 1958 to 2006, we examine if the 11-yr solar cycle indeed modulates the QBO influences in the extratropical stratosphere and if it is the responsible factor for the decadal-scale change of the HT relationship.

[41] Figure 9 shows the correlations between January-February mean equatorial QBO at 50 hPa and January-February zonal-mean zonal wind anomalies (1st row), and January-February temperature anomalies (2nd row). The 1st column shows data for all years, the 2nd and 3rd columns show solar maximum (HS) and solar minimum (LS) years respectively. For the latter columns, the years were grouped when December-Februay mean F10.7-cm solar radio flux (F_s) is high (HS) and low (LS) respectively. HS (LS) years are defined as when the standardized F_s is > (<) than 0.2 (−0.2). The three conditions under consideration are referred to as all-data, HS and LS hereafter.

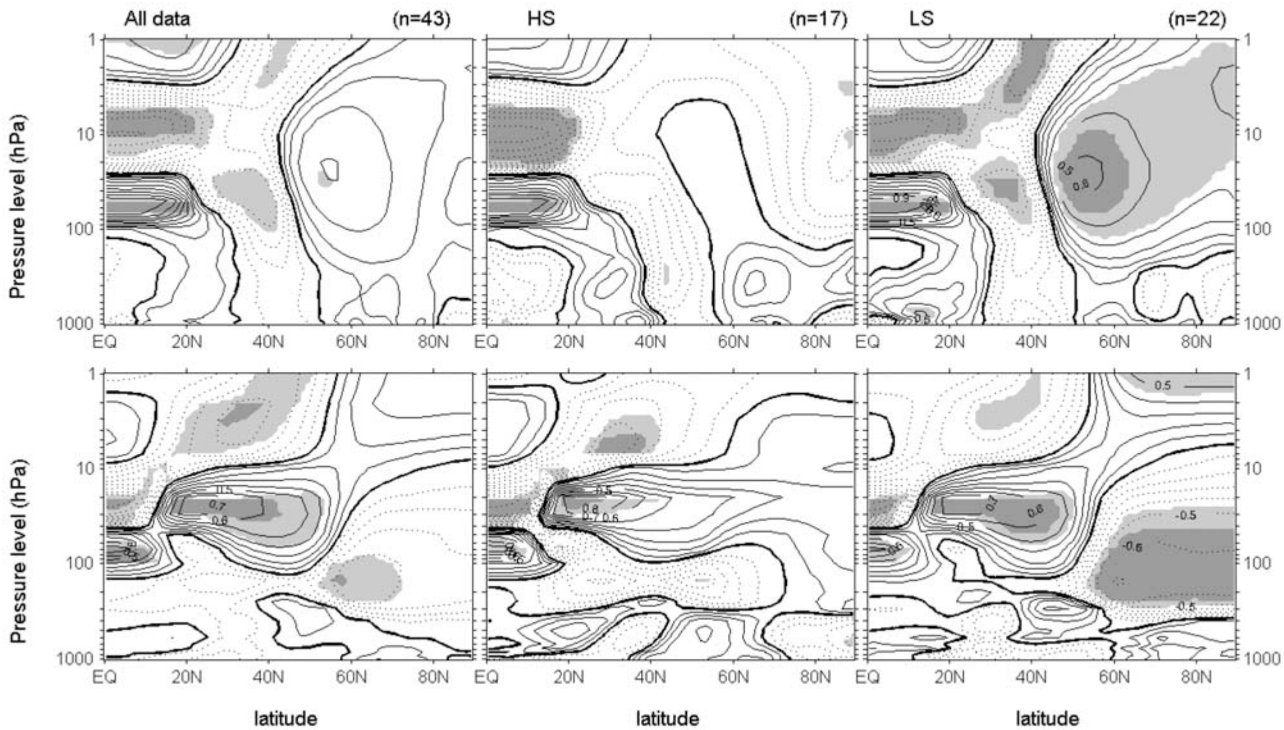

Figure 9. Linear correlations between January-February mean QBO and January-February mean-zonal mean zonal wind anomalies (1st row), and temperature anomalies (2nd row), under all solar conditions (1st column), when December-February mean F10.7 solar radio flux F_s is high (2nd column), and when F_s is low (3rd column). High/low F10.7 is defined as the standardized F_s >/< than +/−0.2. The contour interval is ±0.1 and the thick black contour is zero correlation. The light and dark shaded areas indicate that the differences are statistically significance from zero with a confidence level of 90% and 95% or above, respectively.

[42] As shown in previous studies [*Labitzke and van Loon*, 1988; *Naito and Hirota*, 1997; *Gray et al.*, 2001b, 2004], Figure 9 suggests that the HT relationship is barely significant when all data are used and fails to hold under HS, but becomes significant at a confidence level of 95% under LS. Under LS, significant positive correlations are present in the high latitude wind fields while negative QBO signals occur at midlatitudes. Up to 40% of the variations in the extratropical mid to lower stratosphere can be accounted for by the HT relationship during LS years. However, during HS years this value falls to only 16% or less, showing that the HT relationship is substantially disturbed during solar maximum.

[43] If solar modulation were the primary cause for the QBO signals pre and post 1977, then, for a given solar phase, correlation patterns should remain approximately the same across different periods. However, we found that, under LS, the extratropical QBO signals are substantially enhanced and up to 70–80% of the variations in the stratospheric polar winds and temperature can be accounted for by the QBO alone if only the data from 1958–1976 are used (not shown). In 1977–1997, however, the HT relationship is weaker and not significant under all data, HS and LS. For a given solar phase, nearly the same numbers of data samples exist in 1958–1976 and in 1977–1997, implying that there is no clear bias toward HS or LS for these two subperiods under investigation. Thus the 11-yr solar cycle does not appear to be the responsible factor for the considerably weakened HT relationship during 1977–1997.

4. Discussions

[44] The results presented here provide additional observational analyses to previous findings from observational [*Baldwin and Dunkerton*, 1991, 1998; *Baldwin et al.*, 2001], theoretical [*Plumb and Bell*, 1982; *Haynes*, 1998] and modeling [*Dunkerton*, 1985; *Gray and Pyle*, 1989] studies. Those early studies suggested that the vertical structure of the QBO circulation is a two-cell structure symmetrically straddling the equator. More recently, *Gray et al.* [2004] and *Pascoe et al.* [2005] demonstrated clearly that there is an additional cell in the upper stratosphere. Here, we found that this oscillating pattern of the QBO extends vertically into the upper stratosphere as a three-cell structure both near the equator and in the subtropics. These results are consistent with recent analysis from satellite measurements [*Huang et al.*, 2006].

[45] The ultimate origin for the abrupt changes around 1977 and 1997 in the HT relationship must originate primarily from changes in planetary wave activity or changes in the stratospheric waveguide conditions, or perhaps likely a combination of both. In the past ∼25 years, the chemical composition of the stratosphere has changed substantially due to anthropogenic greenhouse emissions and ozone-depletion. Simulations from coupled chemistry-

climate models suggest that the ozone recovery rates may contribute to the increase in the strength of the Brewer-Dobson (BD) circulation, which may, in turn, be influenced by increases in greenhouse gas concentrations [*Austin and Wilson*, 2006]. GCM simulations also suggest that enhanced BD-circulation leads to weaker westerly winds and higher than average temperatures in the extratropics during late winter and spring [*Butchart et al.*, 2006]. A weakened westerly waveguide may reduce the amount or/and the amplitude of the upward propagation planetary waves. Consequently, it may reduce the differences in waveguide condition between the two QBO phases, thus reducing the effect of the HT relationship.

[46] Vertically propagating planetary waves from the troposphere control the intensity of the equator-to-pole transport of stratospheric ozone by the BD-circulation and thereby modulate the total ozone content at mid and high-latitudes. *Hu and Tung* [2003] suggested that changes in stratospheric ozone may lead to an anomalously induced temperature gradient between mid and high latitudes. This would modify the refraction of planetary waves and thereby either suppress or enhance the propagation of planetary waves into the stratospheric polar region, which then could lead to anomalous cooling or warming in the polar region. *Nathan and Cordero* [2007] found that wave-induced ozone heating can increase wave drag by more than a factor of two in the photochemically controlled upper stratosphere and decrease it by 25% in the dynamically controlled lower stratosphere, suggesting a nonlinear coupling between planetary waves and ozone. Nevertheless, even though changes in chemical composition of the stratosphere are likely to have some effects or perturbations on the waveguide, and therefore the HT relationship, those changes have been too gradual to provide a primary explanation for the abrupt changes near 1977 and 1997.

[47] Near the surface, climate regime shifts occurring in the mid-1970s and the mid-1990s in the North Pacific region have been reported [*Nitta and Yamada*, 1989; *Deser and Blackmon*, 1995; *Yasuda and Hanawa*, 1997]. Since the early 1990s, empirical evidence suggested that a major climate regime shift took place in the mid-1970s, with widespread consequences for the biota of the North Pacific Ocean and Bering Sea [*Hare and Mantua*, 2000]. Recent literature has also suggested another possible regime shift in 1997/1998 winter in the same region [*Minobe*, 2002; *Rodionov and Overland*, 2005]. A predominant feature of those regime shifts is that the sea surface temperature (SST) difference before and after the shift has two action centers dominated by the Arctic and tropical air mass [*Nakamura et al.*, 1997]. For instance, changes on the Arctic influences in the region are exhibited by a substantially warmer SST with weakened surface westerly anomalies before the mid-1970s and colder with enhanced westerly winds afterward. The enhanced surface westerly anomalies reinforce the underlying SST anomalies, resulting in an increase in the intensity and regularity of warm phases of ENSO. It is possible that those near surface changes may alter the relative location and/or intensity of the waves propagating into the stratosphere. Recent studies [*Manzini et al.*, 2006; *Taguchi and Hartmann*, 2006] have shown a link between ENSO and SSWs, which may result in a disruption of the HT relationship. Though it is beyond the scope of this paper, investigation into possible decadal changes of planetary wave activity could provide at least part of the reason for the observed regime shifts.

5. Conclusions

[48] We have re-examined the Holton and Tan (HT) relationship using 49-year ERA-40 and ECWMF Operational combined data. Our findings are the following.

[49] (1) The HT relationship holds in general for the period of 1958–2006, confirming that the QBO-planetary wave mechanism plays an important role in the Arctic dynamics. During November to January, consistent positive or negative QBO signals extend throughout the stratosphere. They are centered at ~5 hPa, 55–65°N with a magnitude of ~10 m s^{-1} in zonal-mean zonal wind, and at ~20–30 hPa at the pole with a magnitude of ~4 K in temperature. In late winter, the dominant feature is a downward movement of westerly or easterly wind anomalies from the midlatitude upper stratosphere, and cooling or warming in the Arctic upper stratosphere with an opposite-signed cell directly below.

[50] (2) The downward movement of the extratropical QBO signals between early and late winter explains why different HT relationships were found and reported for a fixed pressure level [*Holton and Tan*, 1980; *Naito and Hirota*, 1997; *Hu and Tung*, 2002]. At 30–50 hPa, the HT relationship is stronger during early winter and weaker during late winter. In the lowermost stratosphere, the HT relationship holds generally true for the entire winter. In the upper stratosphere, such as 2–3 hPa, the HT relationship is strong but reversed in late winter but weak in early winter.

[51] (3) The HT relationship has not been stationary over time. It was substantially weakened in 1977–1997. The timing of the weakening seems to be synchronous with the climate regime shifts over the North Pacific, with possible additional influences from the changes associated with stratospheric ozone depletion.

[52] (4) During these 49 years (i.e., 1958–2006), the HT relationship in the NH midwinter fails to hold during the maximum phase of the 11-yr solar cycle. This is consistent with the previous findings [*Labitzke and van Loon*, 1988; *Gray et al.*, 2001b; *Labitzke et al.*, 2006]. However, the 11-yr solar cycle does not appear to be the primary responsible factor for the decadal-scale change during 1977–1997.

Appendix A: Significant Tests

[53] Statistical significances are tested using Monte Carlo trial-based nonparametric methods. The Monte Carlo trial-based test proposed by *Wang et al.* [2006] is used to determine the significance of the maximum differences of the running correlation coefficients shown in Figure 5.

[54] First, we ensured that the two time series under consideration can be represented satisfactorily by a first-order auto-regressive (AR1) process using the method of *Neumaier and Schneider* [2001]. That is, whether the two time series under consideration can be represented by two AR1 processes $\{X_t\}$ and $\{Y_t\}$ with:

$$\begin{aligned} X_t - \mu_X &= \alpha_X(X_{t-1} - \mu_X) + \varepsilon_t \\ Y_t - \mu_Y &= \alpha_Y(Y_{t-1} - \mu_Y) + \eta_t \end{aligned} \tag{A1}$$

where (ε_t, η_t) obeys a bivariate normal distribution such that the correlation between ε_t, and η_t is r for any given time t; this implies that, when $\alpha_X = \alpha_Y = \alpha$, the cross correlation between $\{X_t\}$ and $\{Y_t\}$ is $r\alpha^\tau$ at lag τ. Secondly, a large number of (ε_t, η_t) time series can be generated by Gibbs sampling as correlated pairs of time series, in which the correlation coefficient r is determined by performing a linear correlation between the two time series. The simple model of (A1) is then used to generate a large number (e.g., 100,000) pairs of synthetic time series that all have the same length as the original two time series. This allows a large number of synthetic differences between the maximum and minimum values of running correlation between those pairs of synthetic time series to be calculated. Finally, ranking the difference between the maximum and minimum values of running correlation estimated from the original two time series by using the difference distribution generated by the Monte Carlo trials determines its significance level. For simplicity, we call the resulting significance level as p-value, following conventional statistics terminology.

[55] Similar Monte Carlo significance tests are also used to test the statistical significance of the correlation or composite difference between two subsamples. In both cases, two subsamples from the original time series with the lengths equal to the two original subsamples are generated randomly and then the correlation or the difference between their mean values is computed. At each grid point, this procedure is repeated 10,000 times and a distribution of the correlation or the difference is constructed. The correlation or the composite difference between the original subsamples is then compared to this Monte Carlo simulation generated distribution. The rank of the actual correlation or difference among these randomized trials determines its significance level. We say that a signal exists if the correlation or difference within a region is statistically significant at a confidence level at 95% or above.

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