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Review Article
High-latitude influence of the quasi-biennial oscillation

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The interannual variability of the stratospheric winter polar vortex is correlated with the phase of the quasi-biennial oscillation (QBO) of tropical stratospheric winds. This dynamical coupling between high and low latitudes, often referred to as the Holton–Tan effect, has been the subject of numerous observational and modelling studies, yet important questions regarding its mechanism remain unanswered. In particular it remains unclear which vertical levels of the QBO exert the strongest influence on the winter polar vortex, and how QBO–vortex coupling interacts with the effects of other sources of atmospheric interannual variability such as the 11-year solar cycle or the El Niño Southern Oscillation. As stratosphere-resolving general circulation models begin to resolve the QBO and represent its teleconnections with other parts of the climate system, it seems timely to summarize what is currently known about the QBO's high-latitude influence. In this review article, we offer a synthesis of the modelling and observational analyses of QBO–vortex coupling that have appeared in the literature, and update the observational record.

Key Words: Holton–Tan effect; polar vortex; interannual variability; decadal variability; stratosphere–troposphere coupling; seasonal predictability; middle atmosphere; annular modes

Received 17 September 2012; Revised 1 February 2013; Accepted 5 February 2013; Published online in Wiley Online Library 28 March 2013

Citation: Anstey JA, Shepherd TG. 2014. High-latitude influence of the quasi-biennial oscillation. *Q. J. R. Meteorol. Soc.* **140**: 1–21. DOI:10.1002/qj.2132

1. Introduction

The quasi-biennial oscillation (QBO) consists of alternating westerly and easterly zonal wind regimes that descend from the tropical upper stratosphere to the tropical tropopause at a rate of ≈1 km per month, with ≈28 months being required for a complete cycle (Figure 1(a)). The QBO is well known to influence the atmospheric circulation at altitudes and latitudes outside the tropical stratosphere; an extensive review of the QBO, including its manifestations outside the Tropics, has been given by Baldwin *et al.* (2001). In this article we provide an expanded and updated review of one particular area of interest related to the QBO: its impact on the atmospheric circulation at high latitudes during winter, often referred to as the Holton–Tan effect. Dynamical coupling between the QBO and the winter stratospheric polar vortices of both hemispheres has been extensively studied using both observations and models (albeit with the Northern Hemisphere – NH – having received more attention than the Southern – SH), yet the mechanism of coupling between low and high latitudes is still not well understood. Here we attempt a summary of the relevant literature on this subject by collecting in one place a useful list of references, an update of some key diagnostics, a synthesis of current understanding, and some discussion of open questions.

The wintertime variability of the stratospheric polar vortex, particularly the Stratospheric Sudden Warmings (SSWs) that occur roughly every two years in the NH (Charlton and Polvani, 2007), has been a topic of interest for more than fifty years. An understanding of polar vortex variability on intraseasonal and interannual

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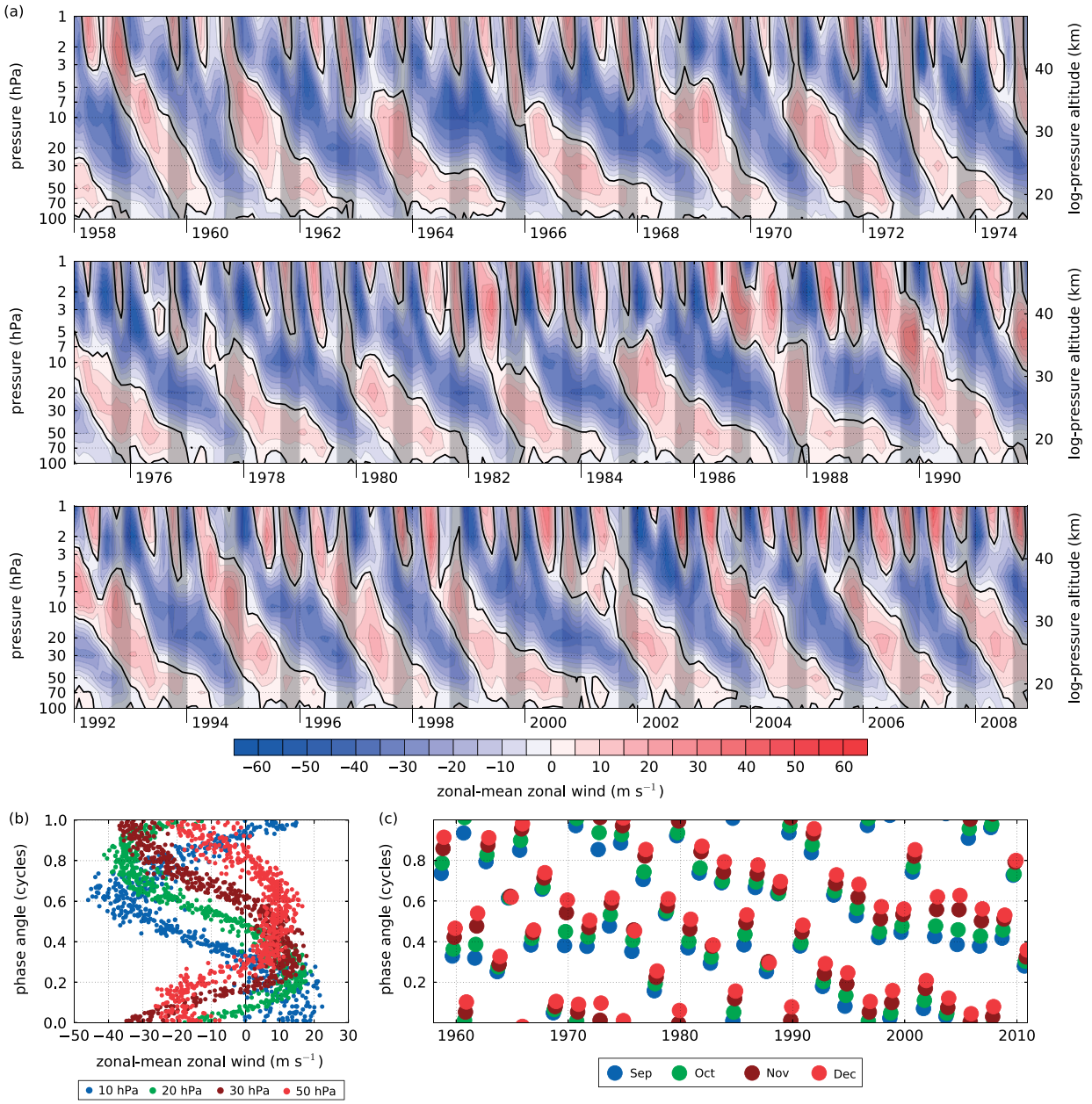

Figure 1. (a) Time series of 1958–2008 equatorial monthly-mean zonal-mean zonal wind, $\overline{u}$, averaged over 2°S–2°N, from ECMWF reanalyses (ERA-40 from January 1958 to August 2002 and ERA-Interim from September 2002 to December 2008). The contour interval is 5 m s^{-1} and shading indicates the September–December period of each year, i.e. SH spring and NH autumn/early winter. (b) Scatterplot of QBO phase angle versus QBO winds at altitudes of 10, 20, 30 and 50 hPa (≈32, 27, 25 and 21 km altitude, respectively, assuming a pressure scale height of 7 km). The phase angle is defined following Wallace *et al.* (1993) using 1958–2010 ECMWF reanalysis deseasonalized monthly-mean $\overline{u}$ in the 70–10 hPa layer (five vertical levels), and is normalized so that a progression through the range 0 to 1 represents one complete QBO cycle (hence phase angles of 1 and 0 are equivalent). QBO winds are equatorial $\overline{u}$ defined as in (a), i.e. they are not deseasonalized. Since the QBO phase angle increases with time almost monotonically, the vertical direction in (b) is the direction of increasing time. Simultaneous values of the QBO winds at different vertical levels are found by taking horizontal slices through the plot. (c) Time series of the QBO phase angle, showing only September–December values. Note that time increases both from left to right and also, due to the phase angle increasing with time, from bottom to top. (The plot could be visualized as the projection of a cylinder, with the direction of increasing time wrapping diagonally about the cylinder.)

time-scales is relevant to issues such as ozone depletion and recovery (e.g. Austin *et al.*, 2003; Shepherd, 2008), stratosphere–troposphere coupling including the possible downward influence of the stratosphere on the troposphere (e.g. Baldwin and Dunkerton, 2001; Shepherd, 2002), and the seasonal predictability of NH winter weather regimes (e.g. Thompson *et al.*, 2002). Stratospheric polar vortex variability may be influenced by numerous external factors including the QBO, the 11-year solar cycle (SC), the El Niño Southern Oscillation (ENSO), and other sources of tropospheric variability. Evidence from mechanistic models of varying complexity also suggests that strong vortex variability may occur under conditions of unvarying external forcing (e.g. Holton and Mass, 1976; Scott and Haynes, 1998; Scott and Polvani, 2004). In the NH winter, large intrinsic dynamical variability has made it difficult to use the relatively short observational record (≈50 years) to unambiguously attribute vortex behaviour to external influences.

The basic explanation for the large interannual variability of the NH stratospheric winter vortex was given by Matsuno (1971), who described the mechanism underlying SSWs. Planetary-scale Rossby waves are generated in

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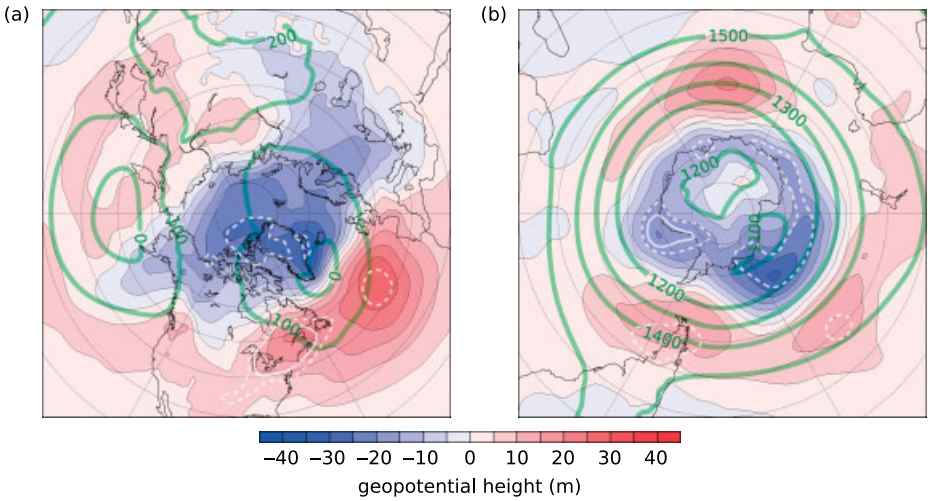

Figure 2. (a) Composite westerly-minus-easterly (W–E) difference in January mean 1000 hPa geopotential height ($Z1000$) in the NH, using 1957–2011 ECMWF reanalysis data (combined ERA-40 and ERA-Interim as described in Figure 1). QBO phase is defined by the sign of January $\overline{u}_{\mathrm{EQ}}$ at 50 hPa, yielding 32 and 22 winters in the W and E groups, respectively. This figure updates Figure 3 of Holton and Tan (1980), which used 1962–1977 NMC data (section 1 gives details). The climatological January $Z1000$ field is shown in green contours. Contour intervals are 5 m and 100 m for the W–E difference and the climatology, respectively. (b) is as (a), but for November mean 850 hPa geopotential height ($Z850$) in the SH, with QBO phase defined by the sign of July $\overline{u}_{\mathrm{EQ}}$ at 20 hPa (yielding 21 and 32 years in the W and E groups, respectively). In both panels, the statistical significance of the W–E differences is indicated by white contours, with dashed (solid) contours denoting the 90% (95%) significance level. Significance is determined by a two-sided Student's t-test applied to the monthly means, assuming each year to be independent.

the troposphere and may propagate upward into the stratosphere during winter when the zonal-mean zonal wind, $\overline{u}$, is westerly. When these waves dissipate in the stratosphere through wave–mean flow interaction processes (Andrews and McIntyre, 1976), they force a poleward meridional mass circulation and decelerate $\overline{u}$ away from the radiative equilibrium state of a strong and cold circumpolar vortex (e.g. Shepherd, 2000). Given sufficiently strong wave forcing, as well as poleward focusing of wave propagation, the polar cap temperature may warm dramatically, resulting in a SSW (Dunkerton *et al.*, 1981). Even if a full-fledged SSW does not develop, wave-driven perturbations in the strength of the polar vortex occur throughout the winter season in both the NH and SH (e.g. Kodera and Kuroda, 2002).

A mechanism for QBO–vortex coupling involving modulation of the position of the subtropical zero-wind line $\overline{u}(\phi, z) = 0$, where ϕ is latitude and z is altitude, was proposed by Holton and Tan (1980, hereinafter HT80). Since the zero-wind line is a critical line for stationary planetary waves (i.e. waves with zonal phase speed of zero relative to the ground), its location is expected to affect the propagation and dissipation of these waves, and thus affect the wave forcing of the polar vortex. The zero-wind line separates the stratospheric westerly winds ($\overline{u} > 0$) of the winter hemisphere, in which planetary waves may propagate, from the easterly winds ($\overline{u} < 0$) of the summer hemisphere, in which these waves are evanescent (Charney and Drazin, 1961). HT80 suggested that by modulating the latitudinal extent of the winter westerlies, the QBO may act either to focus planetary wave activity toward high latitudes or to allow its passage into the Tropics. Increased extratropical confinement of wave activity, corresponding to a poleward shift of the zero-wind line associated with the QBO easterly (QBO-E) phase, is expected to weaken the vortex while the opposite situation during the QBO westerly (QBO-W) phase is expected to strengthen it.

Choosing the QBO wind at 50 hPa ($z \approx 21$ km, in the lower stratosphere) to define the phase of the QBO, and using 16 years of National Meteorological Center (NMC) data covering the period 1962–1977, HT80 showed that the composite westerly-minus-easterly (W–E) difference in extratropical 50 hPa geopotential height ($Z50$) from November to March was consistent with this hypothesis – i.e. QBO-W was on average associated with lower polar $Z50$, and hence a stronger and colder polar vortex, than QBO-E. Additionally, the W–E difference in January extratropical 1000 hPa geopotential height ($Z1000$) displayed a similar pattern (which had also been found in an earlier study by Ebdon, 1975), suggesting a link between QBO phase and extratropical surface climate. Although most studies of the Holton–Tan effect have focused on stratospheric dynamics, the possibility of tropospheric impacts is important because of the high predictability of the QBO. Figure 2(a) uses 1957–2011 reanalysis data from the European Centre for Medium-range Weather Forecasts (ECMWF) to update HT80's January $Z1000$ W–E difference, and similarly to their result it shows negative $Z1000$ anomalies at polar latitudes with a surrounding ring of positive anomalies. The W–E difference strongly resembles a positive Northern Annular Mode (NAM) pattern, with the Atlantic component resembling the positive phase of the North Atlantic Oscillation (NAO). The Southern Annular Mode (SAM) exhibits a similar response to the QBO, albeit with different seasonality, as shown in Figure 2(b) by the November 850 hPa geopotential height ($Z850$) W–E difference. The amplitude of the response in both hemispheres corresponds roughly to a difference of about one standard deviation in the annular mode index (cf. Thompson and Wallace, 2000, Figure 1).*

Despite the apparent observational support for their hypothesis, HT80 stressed that it was not known how to predict from first principles the response of planetary waves

*Note that the equivalent SH plot for $Z1000$ (not shown) yields essentially the same pattern and magnitude as the $Z850$ response shown in Figure 2(b). Also note that Figure 2(a) may be compared with Figure 31 of Baldwin *et al.* (2001), which shows the DJF $Z1000$ W–E difference for the 1964–1996 period.

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to changes in the background mean state, and that their proposed mechanism should not be regarded as definitive. It is clear that the propagation and dissipation of the waves is affected by the background zonal-mean zonal wind, which in turn is altered by the waves (e.g. as happens during a SSW). But the lack of a suitably general theory to predict the wave response to a given background flow that includes a tropical stratospheric zonal wind anomaly, $\Delta\overline{u}_{\mathrm{EQ}}$, makes it difficult to predict the response of the vortex to the QBO. While some progress has been made using small-amplitude approximations that assume a scale separation between the waves and the mean flow (Karoly and Hoskins, 1982; Hauck and Wirth, 2001; Garfinkel *et al.*, 2012), planetary waves in the NH are generally of large amplitude and have spatial scales comparable to those of the mean flow. It is not known how the evolution of the highly nonlinear wave–mean flow interactions that characterize the stratospheric 'surf zone' of breaking planetary waves (McIntyre and Palmer, 1983; Shepherd, 2000) should respond to $\Delta\overline{u}_{\mathrm{EQ}}$, although the broad latitudinal extent of the surf zone lends plausibility to the notion that a response of some kind should occur. The wave response to subtropical zero-wind line position might be the key issue, but the nature of this response remains unclear (Holton and Tan, 1982; Tung and Lindzen, 1979). Hence it is not known from first principles what effect a given $\Delta\overline{u}_{\mathrm{EQ}}$ will have on the evolution of the polar vortex. Nevertheless, the basic assumption that causality of the QBO–vortex relation does arise from the Tropics – rather than in the opposite direction, from Extratropics to Tropics – is fairly certain. The explanation of the QBO as being driven by vertically propagating tropical waves (Lindzen and Holton, 1968; Holton and Lindzen, 1972) superseded previous attempts to explain it as being driven by some kind of quasi-biennial forcing originating in midlatitudes. Hence it is, as HT80 put it, 'reasonable to pose the problem in reverse and attempt to determine whether there are extratropical QBOs driven by the equatorial QBO'.

Studies to date have given some support to the Holton–Tan hypothesis: in numerous observational and modelling analyses, the seasonal development and interannual variability of the extratropical winter stratosphere have shown a sensitivity to the tropical wind state. Yet since HT80 it has become clear that the NH QBO–vortex correlation obtained for the 1962–1977 period is stronger than that obtained in the subsequent observational record, with the early-winter response proving more robust than the late-winter signal. Other sources of interannual variability such as ENSO and the 11-year solar cycle coexist with the QBO, and the observational record remains too short for statistical separation of these competing influences. Studies with mechanistic models of varying complexity have allowed for detailed study of the extratropical dynamical response to imposed $\Delta\overline{u}_{\mathrm{EQ}}$ perturbations, but comparison of their results with observations can be difficult due to the idealizations employed. In particular, imposed forms of $\Delta\overline{u}_{\mathrm{EQ}}$ often have not been realistic facsimiles of observed QBO phases. Atmospheric general circulation models (GCMs) may offer realistic representation of tropospheric variability, which ultimately forces stratospheric variability, as well as allow for feedback of the stratospheric state on the tropospheric flow. Yet these models often possess unrealistic $\Delta\overline{u}_{\mathrm{EQ}}$ due to the difficulties such models often have in resolving the QBO (although the practice of imposing the QBO artificially by relaxation methods has become a useful expedient for studying the Holton–Tan effect with a GCM).

It is now becoming more common for GCMs to resolve the QBO (e.g. Takahashi, 1996; Scaife *et al.*, 2000; Giorgetta *et al.*, 2006; Anstey *et al.*, 2010), potentially allowing realistic simulation of interactions between polar vortices (in either hemisphere), the QBO, and other sources of interannual variability. Hence it seems useful to assess what is currently known about the wintertime high-latitude influence of the QBO, or Holton–Tan effect. We first address observational studies (section 2), emphasizing where detection of a QBO response has been complicated by competing influences such as ENSO or the 11-year solar cycle. Modelling studies (section 3) are categorized by the complexity of the atmospheric model, treating mechanistic models (section 3.1) and GCMs (section 3.2) separately. Emphasis is placed on the nature of the tropical wind perturbations employed in the models and the seasonal variation of their vortex responses, so as to help facilitate meaningful comparison between models and observations. Conclusions and a brief discussion of outstanding issues are given in section 4.

2. Observational studies

Apparent high-latitude signatures of the QBO have been observed since the 1960s, following the discovery of the tropical oscillation (Ebdon, 1960; Reed *et al.*, 1961; Veryard and Ebdon, 1961), but these early efforts were hampered by the limited duration of the observational records available at that time. We consider here only studies subsequent to HT80, which was the first to consider a record of reasonable length (16 years)[†]. Defining a vortex response to QBO phase requires objective definitions of both QBO phase and vortex state. QBO phase is often defined by categorizing winters into QBO-W or QBO-E groups based on the sign of $\overline{u}_{\mathrm{EQ}}$ at a particular vertical level. HT80 used the 50 hPa level, which allowed 15 out of their 16 winters to be placed into the QBO-W or QBO-E groups due to 50 hPa QBO transitions occurring mostly during NH summer in the 1962–1977 period (HT80, p 2201). Vortex state has been defined using various metrics including geopotential (Φ), $\overline{u}$ and $\overline{T}$, polar cap temperature, polar temperature (T_{NP}, for the North Pole), SSW occurrence, and annular mode index. The response is often defined as the difference of the composite-mean vortex states of the two QBO groups (e.g. W–E as in HT80 Figures 2 and 3, and as shown here in Figures 2–4). Intraseasonal variation in the extratropical response has been found in both hemispheres: the NH response appears to be more robust earlier in the winter, while distinct early and late-winter responses are found in the SH. Many studies have considered the QBO response in conjunction with responses to the 11-year solar cycle or ENSO.

An illustration of the seasonally varying extratropical QBO response in both hemispheres is given in Figure 3. The response is shown as the W–E composite-mean difference of smoothed daily $\overline{u}$, where the QBO-W and QBO-E groups are defined by the sign of single-level QBO wind as was done in Figure 2 (i.e. using January 50 hPa $\overline{u}_{\mathrm{EQ}}$ for the NH and

[†]Holton and Tan (1980, 1982) give references and some discussion of earlier works that used shorter data records.

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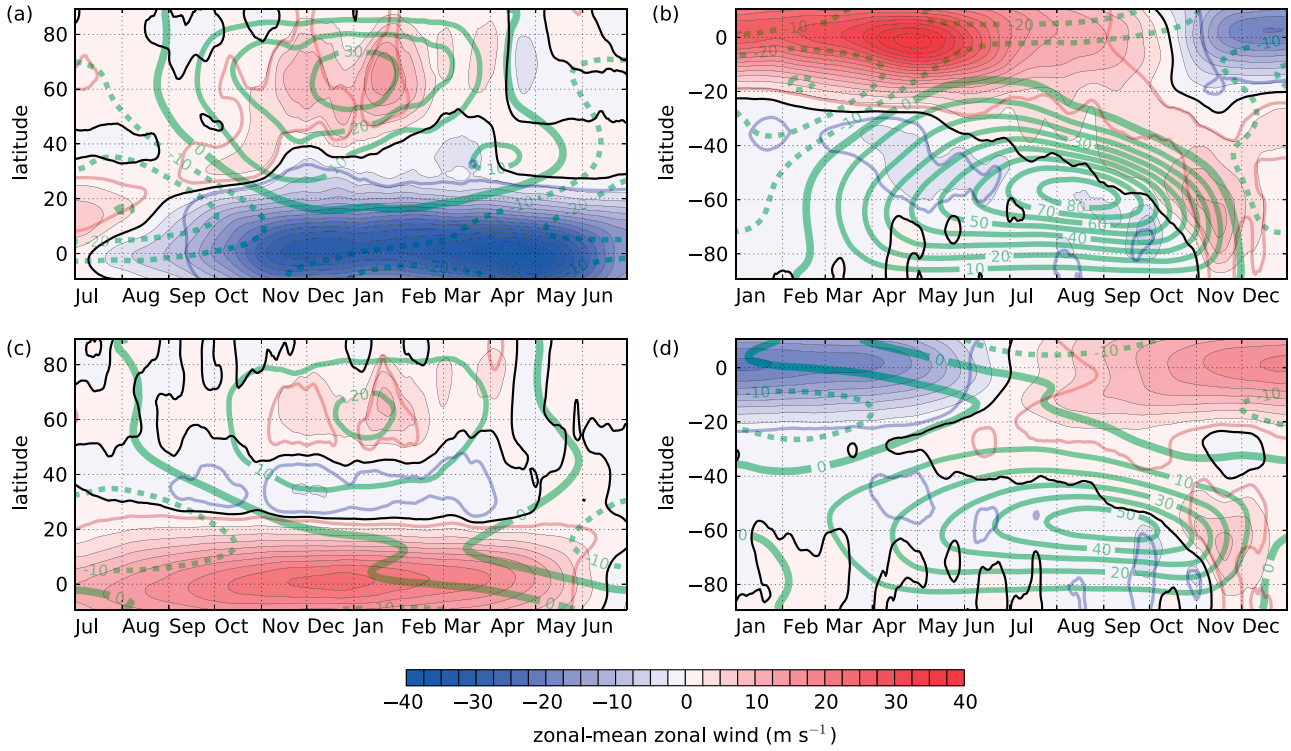

Figure 3. Seasonal development of the high-latitude zonal-mean zonal wind QBO response in (a, c) the NH and (b,d) the SH, at altitudes (a, b) 10 hPa and (c, d) 50 hPa. Daily zonal-mean zonal wind from 1957–2011 ECMWF reanalysis data (as described in Figure 1) is smoothed with a 15-day running mean and composited for westerly (W) and easterly (E) QBO phases, defining QBO phase by the sign of $\overline{u}_{\mathrm{EQ}}$ at 50 hPa in January for the NH composites and 20 hPa in July for the SH composites. The westerly-minus-easterly (W–E) composite difference is shown in filled contours with 2.5 m s^{-1} spacing and the climatological winds are shown in green contours with 10 m s^{-1} spacing, with positive (negative) values solid (dashed) and the zero-wind-line bold solid. The time axis is displaced by six months between hemispheres so that the plots are centred on midwinter for each hemisphere. Statistical significance at the 95% level for positive (negative) W–E differences is indicated by red (blue) contours. The significance is determined by applying a two-sided Student's *t*-test to the 15-day means centred at each calender day, and assuming each year to be independent.

July 20 hPa $\overline{u}_{\mathrm{EQ}}$ for the SH). As indicated above, there is no unambiguously optimal choice of either vortex metric or QBO phase with which to define the QBO response. Hence we have chosen here to present results based on a simple QBO phase definition (sign of $\overline{u}_{\mathrm{EQ}}$) and a typical choice of extratropical field ($\overline{u}$). Classifying winters by only the QBO phase gives no explicit consideration to competing influences such as ENSO and the solar cycle, and hence their effects may be aliased in the responses shown in Figure 3. While it would be hoped that such effects would average to zero given a sufficiently long time series (provided that these competing factors have no long-term correlation with the QBO itself), it is doubtful that the observational record is long enough for this assumption to be valid. Bearing these caveats in mind, several features in Figure 3 are notable. The QBO response appears as a modulation of the seasonal cycle (shown in green contours) with different timing in each hemisphere: it is strongest from early winter to midwinter in the NH and, at high latitudes, in late winter in the SH. As such, the NH midwinter response has often been interpreted as a modulation of SSW activity and the SH response as a modulation of the final warming (i.e. the transition to summer stratospheric easterlies). The vertically deep nature of the extratropical response (comparing the 10 hPa and 50 hPa panels of Figure 3) is consistent with the QBO modulating the vertically coherent leading modes of extratropical variability, i.e. the NAM and SAM (Baldwin and Dunkerton, 1999; Thompson and Wallace, 2000). Differences in the tropical winds themselves ($\Delta\overline{u}_{\mathrm{EQ}}$) exhibit a much stronger vertical variation, illustrating the fact that composite differences such as these do not provide an unambiguous test of the Holton–Tan hypothesis. The vertical structure of both $\Delta\overline{u}_{\mathrm{EQ}}$ and the extratropical response is seen in greater detail in Figure 4, which shows the zonal cross-section (i.e. vertical and meridional structure) of the W–E differences in January (for the NH) and November (for the SH, although note that some indication of the NH November response is also present in Figure 4(b)).[‡]

[‡]An estimate of the statistical significance of the W–E differences, using Student's *t*-test with the assumption that each year is independent (i.e. represents one degree of freedom), is indicated in Figures 2, 3, and 4 (see figure captions for further details). This is a commonly used method, but it is not definitive; various significance tests have been used in the literature, and some studies have given detailed consideration to statistical methods (e.g. Ruzmaikin *et al.*, 2005; Camp and Tung, 2007b, among others). We consider a detailed description of statistical methods to be outside the scope of this review, although we report the significance levels that were obtained in some of the studies. Results of statistical tests are naturally contigent upon assumptions about the underlying distribution of the data, including the length of the available record (i.e. whether the available data adequately sample the phase space of the underlying dynamics). A key issue in diagnosing the QBO response is whether QBO influence has been adequately separated from competing influences. More generally, it is unclear how long a record is required to separate the signal of a vortex response to any particular forcing – such as the QBO, ENSO, or solar cycle – from the noise of the internal variability of the extratropical atmosphere (e.g. Yoden *et al.*, 2002; Fomichev *et al.*, 2007).

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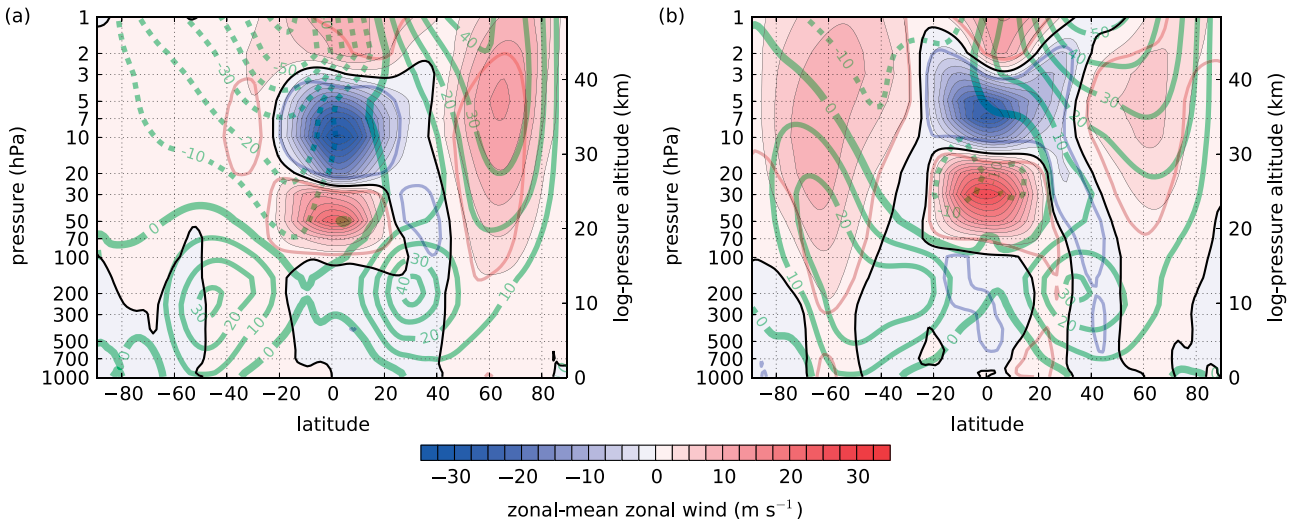

Figure 4. Vertical and meridional structure of the zonal-mean zonal wind QBO response in (a) January, defining QBO phase by the sign of 50 hPa January $\overline{u}_{\mathrm{EQ}}$ (as was done for the NH in Figures 2 and 3), and (b) November, defining QBO phase by the sign of 20 hPa July $\overline{u}_{\mathrm{EQ}}$ (as was done for the SH in Figures 2 and 3). Data are 1957–2011 ECMWF reanalysis as in previous figures. As in Figure 3, the W–E composite difference is shown in filled contours with 2.5 m s^{-1} spacing and the climatological winds are shown as green contours with 10 m s^{-1} spacing with positive (negative) values solid (dashed) and the zero-wind-line bold solid. Statistical significance at the 95% level for positive (negative) W–E differences is indicated by red (blue) contours, with the significance determined by a two-sided Student's *t*-test applied to the monthly means, assuming each year to be independent.

2.1. *Northern Hemisphere*

The possible links between QBO phase and SSW occurrence have been explored extensively. Labitzke (1982) examined a 26-year record of 30 hPa T_{NP} and found midwinter SSWs to be less common during QBO-W, although periods of solar maxima (i.e. maxima of the 11-year sunspot cycle) appeared to be an exception to this rule. McIntyre (1982) hypothesized that a deep layer of equatorial easterlies was likely to enhance the probability of SSW occurrence, noting that the modelling study of Dunkerton *et al.* (1981) had found that a poleward movement of the $\overline{u}(\phi, z) = 0$ line appeared to accompany SSW development. McIntyre (1982) stressed the importance of a 'preconditioning' of the stratospheric circulation by wave forcing, wherein the vortex becomes more likely to undergo a SSW if the accumulated effects of wave forcing over the course of the winter have caused it to contract and/or become displaced from the pole; he suggested that the QBO phase might influence the evolution of the vortex towards a preconditioned state due to its effect on the accumulated wave forcing. However, Wallace and Chang (1982) and Holton and Tan (1982) noted that SSWs are likely to be affected by other sources of interannual variability such as ENSO, and that the available observational record did not (at that time) provide a convincing separation of possible ENSO and QBO effects. Wallace and Chang (1982) also emphasized that QBO influence need not be manifest only through SSWs – which do not exclusively encapsulate the variability of the vortex – noting that HT80 had attempted some separation of SSW and non-SSW effects by considering early (November–December) and late (January–March) winter separately. (Major SSWs tend to occur only from December onward.) Similarly, Labitzke (1982) considered November separately from other NH winter months.

Following the suggestion of McIntyre (1982), an explicit association between deep QBO phases and the occurrence of SSWs was investigated by Dunkerton *et al.* (1988), who found that major SSWs tended not to occur with the QBO in a deep westerly phase. They emphasized that defining QBO phase by a deep layer of equatorial winds was not equivalent to the 50 hPa definition used by HT80, since the downward descent of QBO phases often stalls near the 30 hPa level during NH winter; such stalling can be seen in Figure 1(a) as giving a boot-like shape to some of the QBO-W phases (e.g. 1964/65, 1988/89). Figure 1(b) also shows the likelihood of deep QBO phases by showing the relationship between QBO winds at different vertical levels (e.g. during QBO-W at 50 hPa, the 20 hPa QBO is no more likely to be W than it is to be E, because the two levels are roughly 90° out of phase). Labitzke (1987) and Labitzke and van Loon (1988) cast further doubt on the robustness of the QBO–SSW association by showing an apparent interference between the 11-year solar cycle and the Holton–Tan effect: defining QBO phase by 50 hPa $\overline{u}_{\mathrm{EQ}}$, they found that during QBO-W winters a strong positive correlation occurred between mean January–February 30 hPa T_{NP} and solar variability (as defined either by sunspot number or by 10.7 cm radio flux), while a weak negative correlation was found during QBO-E.[§] The correlations in the two studies were, respectively, 0.78, 0.76 (−0.31, −0.45) for the QBO-W (QBO-E) groups, and the QBO-W group correlations were argued to have high statistical significance.

An update of the relationship between the QBO and SC is given in Figure 5(a), which shows the joint distribution of the daily vortex state in the QBO–SC phase space (using $\overline{u}$ at 10 hPa, 60°N to characterize the January–February vortex state instead of 30 hPa T_{NP}). Dividing this phase space into four quadrants and taking the mean of the vortex state over each quadrant yields the picture shown in Figure 5(c), which is consistent with Labitzke and van Loon (1988): during QBO-W phases, high solar activity (SC-max) appears to weaken the vortex relative to low solar activity (SC-min), while during QBO-E the opposite relation is obtained (i.e. SC-max strengthens the vortex

[§]Labitzke (1987) actually used November–February mean 30 hPa T_{NP}, but Labitzke and van Loon (1988) found that January and February gave the main contributions to the solar cycle signal, and thus removed early winter (November and December) from their analysis.

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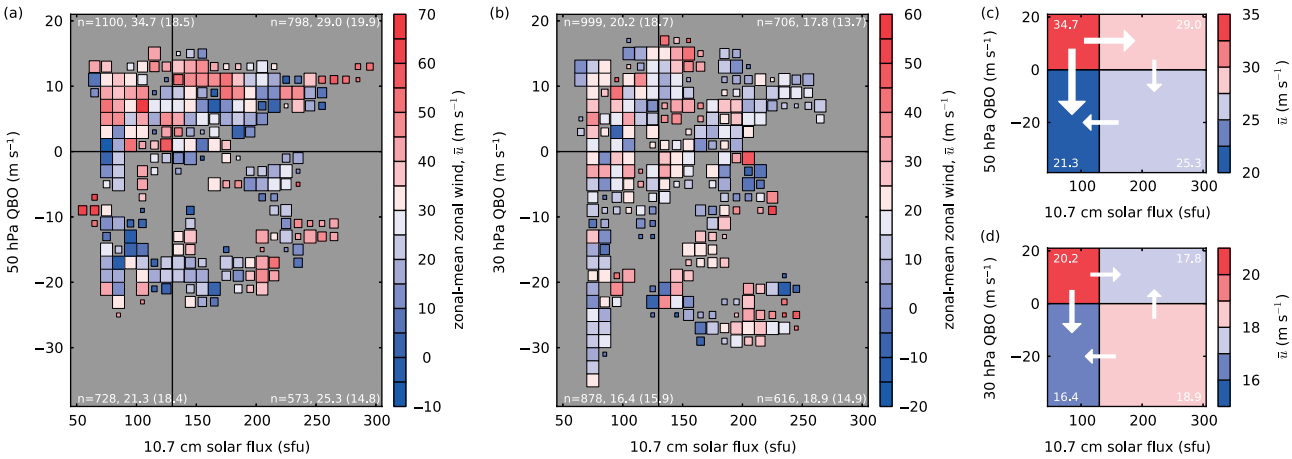

Figure 5. (a) Joint distribution of vortex strength during January and February with respect to QBO phase (defined by daily 50 hPa $\overline{u}_{\mathrm{EQ}}$) and the 11-year solar cycle phase (defined by 10.7 cm flux). The vortex is defined by daily $\overline{u}$ at 10 hPa, 60°N using 1957–2011 reanalysis data (as described in Figure 1), which contains 54 complete NH winters. A 31-day running mean is used to smooth the daily mean 10.7 cm flux values. The distribution is estimated as a joint histogram, with QBO and solar bins defined by increments of 2 m s^{-1} and 10 sfu, respectively (1 sfu = 10^{-22} W m^{-2} Hz^{-1}). The colour in each box corresponds to the mean value of vortex $\overline{u}$ for all days within that box, with a colour scale interval of 5 m s^{-1} . Grey indicates no data. An indication of the number of days in each box, n, is given by the box size: if $n < 10$, then the box area is proportional to n. Solid black lines separating each plot into four quadrants designate the four groups based on combinations of two-state QBO (W or E) and solar cycle (min or max) phases. The value of 130 sfu chosen to separate SC-min and SC-max is based on Naito and Hirota (1997), Figure 1(b). White text indicates for each group the total number of days, group-mean value of vortex $\overline{u}$, and (in brackets) standard deviation of vortex $\overline{u}$ with respect to the group mean. Note that (a) may be compared to Figure 8 of Naito and Hirota (1997), which similarly shows the distribution of the vortex state in QBO–SC phase space (for the 1953–1995 period) but characterizes the vortex state by the occurrence of major SSWs. (b) is as (a) but for daily vortex $\overline{u}$ during February and March, with QBO phase defined by daily 30 hPa $\overline{u}_{\mathrm{EQ}}$. (c, d) are as (a, b), respectively, but with data binned only into the four QBO–SC groups. White text indicates the mean vortex $\overline{u}$ for each group, and note that all panels (a–d) have different colour scales. Arrows across the quadrant boundaries in (c,d) point in the direction in QBO–SC phase space in which the vortex is more disturbed (i.e. weaker and warmer, on average). Arrow sizes are scaled by the significance level of the differences between group means (W-min minus E-min, etc.), where significance is estimated by a two-sided Student's t-test applied to the daily data. The number of days required for an independent sample is estimated using the e-folding times for the 10 hPa NAM autocorrelation shown in Figure 1(a) of Baldwin *et al.* (2003). The largest arrow, for the W–E difference under SC-min during Jan–Feb, corresponds to the 99% significance level under this test; the next highest level is for min–max under QBO-W during Jan–Feb (88%). For Feb–Mar, W–E under SC-min has the highest significance (81%). The precise values of these significances are sensitive to the assumed e-folding time, but their relative values are not; hence the arrow sizes in (c, d) provide a schematic representation of the relative strength of these signals against the large underlying vortex variability (at least as quantified by this statistical test, for this vortex metric). Similar results are obtained if the test is applied to bimonthly means with each year assumed independent.

relative to SC-min). This behaviour may be regarded as a QBO modulation of SC influence, whereby the QBO state acts to reverse the sign of the SC–vortex relationship, or equivalently as a SC modulation of QBO influence, whereby the SC state acts to change the strength of the QBO–vortex relationship. (Note from Figure 5(c) that the W–E difference retains the same sign under both SC-min and SC-max, but is substantially weaker under SC-max). Baldwin and Dunkerton (1989) performed further statistical tests of this relation and noted that stratifying the data by QBO phase would tend to produce high correlations between any number of low-frequency harmonics (used as proxies for the solar cycle) and T_{NP}, provided that the level defining QBO phase was determined *a posteriori* by choosing the level that optimized the correlation (i.e. that made it strongly positive or negative). While their statistical test nevertheless obtained a significance level of ≈ 95% for the QBO-W group correlation, it highlighted the strong sensitivity of such tests to their underlying assumptions. Labitzke and van Loon (1988) and Baldwin and Dunkerton (1989) also noted that the spatial pattern (in latitude–longitude) of the extratropical response to the solar cycle bore a strong similarity to the QBO or ENSO response. Baldwin and Dunkerton (1989) showed high correlations between 50 hPa T_{NP} and 50 hPa T at other latitudes and longitudes in the NH, implying that mean January–February stratospheric T possesses few degrees of freedom; thus any time series strongly correlated with polar-cap T would tend to be strongly correlated with T over the whole Extratropics, consistent with the QBO, ENSO and SC responses all showing similar spatial patterns.

The analysis of HT80 for the NH was updated by Dunkerton and Baldwin (1991). They showed that while QBO influence appeared to weaken in their longer record of NMC analyses (1964–1988), its statistical significance remained high even when considering the possibility of the vortex exhibiting a lag-1 year anticorrelation unrelated to the QBO. Defining QBO phase according to the 40 hPa equatorial wind was found to give the strongest correlations with the Extratropics. The 1979–1988 data, which incorporated satellite measurements, allowed the Holton–Tan effect to be diagnosed at altitudes up to 1 hPa. Using the Eliassen–Palm flux (hereinafter 'EP flux') as the measure of planetary wave activity (Andrews *et al.*, 1987), Dunkerton and Baldwin (1991) showed that differences in the spatial distribution of extratropical stratospheric EP flux convergence between the QBO-W and QBO-E categories were consistent with planetary wave dissipation acting to force both a mean meridional circulation and the observed $\overline{u}$ difference. Although this consistency lent support to the notion that planetary wave dissipation causes QBO–vortex coupling, it was noted that waves and mean flow are closely coupled irrespective of QBO state; this was demonstrated from high correlations between EP flux convergence and $\overline{u}$ on daily time-scales, and the prevalence of strong planetary wave activity during both QBO phases was shown. A further update of HT80 was provided by Naito and Hirota (1997) using Freie Universität Berlin (FUB) data for 1957–1994, who found that the early winter (November–December)

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QBO signal appeared to be unrelated to solar cycle variations while the late winter (January–February) QBO signal obeyed the modulation described by Labitzke and van Loon (1988). Naito and Hirota (1997) proposed that the reason why HT80 found a strong QBO–vortex coupling was that their analysis period (1962–1977) contained two solar minima and one solar maximum. They suggested that the polar vortex is mainly influenced by the QBO in early winter and by the solar cycle in late winter.

A physical mechanism for a solar cycle influence on polar variability was suggested by Kodera (1991, 1995). Analysis of 12 years of NH extratropical NMC data suggested that the four combinations of QBO and SC phase – W/max (QBO-W and SC-max), W/min, E/max and E/min – induced extratropical responses with distinct spatial patterns in early winter yet similar patterns in late winter, with only the polarity of the late-winter response depending on the forcing (i.e. on the combined QBO–SC state). Based on the common pattern shown by the late-winter responses, it was argued that the early-winter influence of these forcings created a late-winter polar signal by means of downward and poleward propagation of the $\overline{u}$ anomaly due to wave–mean flow interaction, characterizing this as a natural mode of extratropical stratospheric variability, analogous to behaviour seen in idealized stratospheric models (Holton and Mass, 1976), which could be excited by forcings such as the QBO or the SC. A similar view was advocated by Baldwin and Dunkerton (1991), who diagnosed QBO–extratropical correlations using 13 years of NMC data extending up to 1 hPa. They showed high correlations during NH winter between $\overline{u}$ in the layer above 10 hPa and the 40 hPa QBO zonal wind, extending from NH polar latitudes to 40°S, with the NH signal poleward of 40°N being strongest from early to midwinter (see their Figure 2). They also stressed the vertical coherence of the NH signal, which extended over the whole depth of the extratropical stratosphere; Figure 4(a) illustrates this spatial structure using the updated record. Baldwin and Dunkerton (1991) noted that the strong signal at high altitudes may or may not indicate a causal role for the flow at those altitudes (since correlation does not imply causality), and suggested that the Holton–Tan effect should be regarded as an important mode of stratospheric interannual variability that, while triggered by the QBO, may also be triggered by other mechanisms.

Pursuing a similar line of argument, Dunkerton and Baldwin (1992) performed an Empirical Orthogonal Function (EOF) analysis of NH $\overline{u}$ and zonal-mean temperature $\overline{T}$ and showed that the leading mode of variability coincided well with the spatial structure of correlations between the QBO and the Extratropics during December, January and February, but less well with the solar cycle correlations in the same months. They noted that most of the interannual variance of mean December–February (DJF) $\overline{T}$ at high latitudes could be represented by two signals: the QBO, and a term representing the modulation of QBO influence by 'quasi-decadal variability' (QDV). The QDV-modulation terms also yielded a spatial correlation pattern, in February, similar to that of the leading EOF of $\overline{T}$. This led them to argue that polar variability was best thought of as an internal mode triggered by the QBO but modulated on a decadal timescale by QDV – with the QDV possibly, but not necessarily, being due to solar cycle influence. They suggested that determining the true role of the solar cycle would require a longer record. Kodera (1993) further clarified the behaviour of the QDV by calculating the running correlation between T_{NP} and 45 hPa QBO wind and argued that decadal variability in the strength of QBO–vortex coupling was not a spurious effect, as had previously been argued by Salby and Shea (1991). The decadally varying modulation of the QBO signal was diagnosed by demodulation of the polar temperature (Figure 3(c) of Kodera, 1993) without any reference to the solar cycle. Thus while the existence of a QDV modulation of the QBO signal was succinctly demonstrated, its explicit connection to the solar cycle remained unclear.

2.2. *Southern Hemisphere*

Although most studies have focused on the NH, a high-latitude QBO signal is also seen in the SH. The stratospheric polar vortex in SH winter is stronger, colder and longer-lived than its NH counterpart, and only one SSW has occurred in the observed record (during September 2002). HT80 used data from SH stations centred along 145°E longitude to show that during September–November (SH spring) the extratropical 50 hPa QBO response defined by 50 hPa QBO phase showed stronger westerlies during QBO-W, although with weak statistical significance. Also examining SH spring, Garcia and Solomon (1987) suggested that an apparent biennial modulation of October minimum polar ozone values might be due to QBO influence, and Newman and Randel (1988) noted this variability to be coherent with variations in planetary wave forcing. Exploring this behaviour in more detail, Lait *et al.* (1989) argued that QBO influence on polar downwelling could significantly modulate Antarctic ozone depletion due to the polar temperature being close to the threshold for chlorine activation. Spectral peaks suggestive of QBO influence seen in ozone in both hemispheres (Tung and Yang, 1994) were suggested by Baldwin and Tung (1994) to be caused by QBO-induced anomalies associated with similar spectral peaks seen in extratropical zonal angular momentum and EP flux divergence (using 1978–1993 data for the SH and 1964–1993 data for the NH).[¶]

A comparison of QBO–vortex coupling in the two hemispheres was made by Baldwin and Dunkerton (1998), using National Centers for Environmental Prediction (NCEP) analyses spanning 1978–1996 on pressure levels from 1000 hPa to 1 hPa. They represented the QBO state in terms of a 'phase angle' defined from the two leading EOFs of the tropical radiosonde vertical profiles of monthly-mean zonal wind (Wallace *et al.*, 1993); Figure 1(b) shows the relationship between this phase angle and equatorial winds at selected altitudes. Analogously to using winds at a single altitude to define QBO phase, Baldwin and Dunkerton (1998) searched for the values of the phase angle that yielded the strongest extratropical response in each hemisphere. The optimal NH signal was found to be roughly equivalent to defining QBO phase by the 40 hPa equatorial wind (consistent with Dunkerton and Baldwin,

[¶]It should be noted that many of the studies examining QBO–vortex coupling in the SH have focused on ozone, rather than on dynamical quantities such as temperature and zonal wind (as are often used for the NH). Some additional references are given in the works that we cite here. The global distribution of the QBO in ozone has been extensively studied; Baldwin *et al.* (2001) give further details.

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1991), while roughly the 25 hPa equatorial wind was found to optimize the SH signal. The timing and magnitude of the QBO signal during SH winter was found to differ markedly from that of NH winter: QBO-W (at 25 hPa) coincided with anomalous westerlies only in midlatitudes during May–October, followed by a strengthened polar vortex – i.e. a delay in the final warming – in November. In contrast, the NH QBO signal reaches polar latitudes early in winter, and no robust modulation of the final warming has been observed. These features of the QBO response, shown in Figures 2 and 3 of Baldwin and Dunkerton (1998), are apparent in Figure 3 (i.e. in the updated record). Naito (2002) also considered the QBO signal in the SH, finding similar results to Baldwin and Dunkerton (1998). Analyzing the influence of the QBO on EP fluxes, Naito (2002) argued that, in the SH Extratropics, QBO modulation of upward wave fluxes resided primarily in zonal wavenumbers $k = 1$–3 while QBO modulation of equatorward wave fluxes was dominated by $k = 4$–6, suggesting distinct QBO influences on planetary- and synoptic-scale waves. Interhemispheric and seasonal variations in QBO–vortex coupling were further characterized by Hitchman and Huesmann (2009), who suggested that autumn and spring responses to middle-stratospheric QBO winds were similar in the two hemispheres, but that a midwinter response to lower-stratospheric QBO winds existed only in the NH. Hitchman and Huesmann (2009) also examined QBO modulation of stratospheric Rossby wave breaking using diagnostics of the frequency and strength of potential vorticity meridional gradient reversals.

2.3. QBO phase

While correlations between the Extratropics and the QBO may be optimized by choosing a single vertical level of equatorial $\overline{u}$ to define the QBO phase – e.g. 40 hPa for the NH and 25 hPa for the SH in Baldwin and Dunkerton (1998) – the reason why these particular levels give the strongest signal is unknown. Using 1964–1990 rocketsonde data extending to 58 km, Gray *et al.* (2001b) demonstrated high correlations between January–February T_{NP} at 30 hPa and equatorial winds near the stratopause during the preceding August–September. (In particular, the anomalous shear $\Delta\partial\overline{u}_{\mathrm{EQ}}/\partial z$ at $z = 50$ km in September gave a correlation of -0.68 and described 46% of the variance of January–February T_{NP}, higher than the 16% of variance described by December $\Delta\overline{u}_{\mathrm{EQ}}$ at $z = 24$ km.) The focus on equatorial winds at lower altitudes by earlier authors may have been motivated by the availability of a continuous and lengthy record of tropical radiosonde data; radiosondes can only reach a maximum altitude of roughly 30 km (10 hPa) before their balloons burst. Using ECMWF 40-year reanalysis (ERA-40) data, a QBO signal has been shown to extend across the whole depth of the tropical stratosphere with a three-cell vertical structure, although its magnitude in the 40–50 km layer is diminished in comparison with the 20–40 km layer: W–E composite differences at the Equator yield a peak-to-peak QBO strength of 30 m s^{-1} at 23 km, 25 m s^{-1} at 35 km, and 10 m s^{-1} at 48 km (Pascoe *et al.*, 2005). These magnitudes are consistent with the vertical structure of the W–E composite differences shown in Figure 4. The relatively weak QBO amplitude at altitudes above 3 hPa ($z \approx 40$ km) is unlikely to strongly modulate the position of the $\overline{u}(\phi, z) = 0$ line, since wind variability in this region is dominated by the semiannual oscillation (as shown in Figure 1(a)). However, it is by no means certain that $\overline{u}(\phi, z) = 0$ modulation is the dominant cause of QBO–vortex coupling. Kodera (1991) and Ruzmaikin *et al.* (2005) argued that the QBO-induced mean-meridional circulation at low latitudes (Plumb and Bell, 1982), which exhibits a seasonal asymmetry about the Equator and is stronger in the winter hemisphere (Jones *et al.*, 1998; Kinnersley, 1999), may alter low- and mid-latitude $\overline{u}$ so as to affect planetary wave propagation and thereby generate the extratropical response.

The choice of QBO phase affects the detection of QDV in the high-latitude QBO signal. Dunkerton *et al.* (1988) noted that the 11-year solar cycle appeared to modulate QBO–vortex coupling when 50 hPa $\overline{u}_{\mathrm{EQ}}$, but not 30 hPa $\overline{u}_{\mathrm{EQ}}$, was used to define QBO phase, and Camp and Tung (2007b) showed that this remains true in a longer record (1954–2001, using January–February T_{NP} to define the vortex; their Figure 3b). If a deep layer of westerly or easterly equatorial winds does in fact induce the strongest vortex response, then it is unclear why 50 hPa is a better choice than 30 hPa; Dunkerton *et al.* (1988) suggested that the 50 hPa QBO index should be relied upon only when its use can be justified *a priori*. Unfortunately it is not clear on physical grounds what choice of QBO phase is most appropriate (as was discussed in section 1). Although modelling studies have yielded some insight into this question (as will be discussed in section 3), it nevertheless remains an open question. However, observations of the QBO winds show that the vertical structure of anomalous $\overline{u}_{\mathrm{EQ}}$ can vary widely from one winter to the next, as seen in Figure 1(a) by comparing, for example, the shaded September–December regions. The QBO's period of ≈ 28 months (Baldwin *et al.*, 2001) is responsible for much of this variation, since it leads to 'beating' between the QBO and the annual cycle. An exact 28-month period would imply that 7 years is required for any given QBO phase alignment with respect to the annual cycle (hereafter simply 'QBO phase alignment') to repeat. In reality the QBO period varies over time, with QBO phase transitions exhibiting a weak tendency to synchronize with the seasonal cycle (Dunkerton, 1990). Hence the beating is irregular, and the QBO phase alignment exhibits an irregular QDV, which includes some tendency toward episodes of synchronization with the seasonal cycle (Read and Castrejón-Pita, 2012). These features can be discerned in the time series of QBO phase angle shown in Figure 1(c). Regular beating corresponds to times when the phase angle drifts at approximately a fixed rate with respect to the annual cycle (e.g. 1975–1990) while seasonal synchronization corresponds to times of little or no drift (e.g. 1995–2010), which give the diagram a banded appearance. The QBO appears to vary between these two behaviours, with Figure 1(c) suggesting a decadal time-scale for switching between them, although a longer observational record would help to clarify whether this interpretation is robust.

Slow modulation of QBO phase alignment, or beating, was shown by Gray and Dunkerton (1990) to cause slow modulation (i.e. QDV) of the subtropical ozone QBO signature in their two-dimensional photochemical model, similar to that observed in ozone. Since QBO phase alignment will affect the vertical and latitudinal structure of anomalous $\overline{u}_{\mathrm{EQ}}$ that prevails during any given winter, it might also affect dynamical QBO–vortex coupling (i.e. the QBO signal at higher latitudes). Anstey and Shepherd

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(2008) used ERA-40 data (1958–2002) to argue that the weak seasonal synchronization of QBO phase transitions (Dunkerton, 1990) causes a QDV in QBO phase alignment that leads to QDV in the NH late-winter vortex response (defined by February–March 50 hPa 60°N $\overline{u}$), and suggested this could explain the weakened QBO–vortex coupling found during the 1977–1997 period by Lu *et al.* (2008). Using NCEP (1948–2008), ERA-40 and FUB (1957–1997) data, Christiansen (2010) showed that bimodality in the probability distribution of the extended winter mean (October–March) NH vortex, as defined by the leading EOF of 20 hPa or 30 hPa geopotential height poleward of 20°N, coincided well with bimodality in the distribution of 50 hPa extended winter mean QBO winds over the whole record; a weakening of this QBO bimodality in the 1979–1999 period due to the weak seasonal synchronization of QBO phase transitions was argued to be the likely cause of a weakened QBO–vortex coupling during that time.

2.4. Solar cycle

The hypothesis that QDV of QBO–vortex coupling may be caused by the 11-year SC has largely survived the test of time provided by the longer observational record available since Labitzke and van Loon (1988); this has been shown in a number of studies, and was also illustrated in Figure 5. Using FUB data for the 1955–1999 period, which spans four solar cycles, Gray *et al.* (2001b) examined QBO modulation of the SC signal. Defining QBO groups based on 50 hPa $\overline{u}_{\mathrm{EQ}}$, correlations of 0.68 (−0.14) between mean January–February 30 hPa T_{NP} and 10.7 cm radio flux were found in the QBO-W (QBO-E) groups. Labitzke (2005) used NCEP/NCAR reanalysis data for 1958–2003 (1968–2003) for the NH (SH) to compare the vortex response to the QBO and solar cycle in both hemispheres. Defining QBO phase by 45 hPa $\overline{u}_{\mathrm{EQ}}$, correlations between February 30 hPa T_{NP} and the 10.7 cm radio flux were found to be 0.65 (0.02) in the QBO-W (QBO-E) groups, echoing Labitzke's (1987) result of a highly significant correlation for the QBO-W group and a statistically insignificant correlation for the QBO-E group. The spatial pattern of the NH February 30 hPa Φ solar cycle response (i.e. SC-max minus SC-min) resembled the NAM, while that of the SH September 30 hPa Φ response resembled the SAM; in both hemispheres, the response showed a positive (negative) annular mode pattern for the QBO-W (QBO-E) group. Labitzke *et al.* (2006) extended these results for the NH using a 65-year record dating back to 1942 (achieved using historical reconstructions of QBO phase and vortex strength for the 1942–1947 period, and NCEP/NCAR reanalyses for 1948 onwards), covering seven (six) solar minima (maxima). The correlations between February 30 hPa Φ at the North Pole and the 10.7 cm radio flux were 0.69 (−0.30) in the QBO-W (QBO-E) groups (again using 45 hPa QBO phase to define the groups). While the correlation between T_{NP} and the solar cycle in the QBO-W group has remained strong in the longest available observational records, the correlation in the QBO-E group remains weak.

It remains unclear why a strong high-latitude response to the 11-year solar cycle becomes apparent only when the data are stratified by QBO phase. Kuroda and Kodera (2002) used 1979–1999 NMC data up to 1 hPa to characterize the solar cycle response without reference to the QBO. They argued that in both hemispheres the effect of the solar cycle was to modulate the seasonal development of the vortex, with SC-max favouring an earlier strengthening of the subtropical stratopause jet. They noted the potential of the QBO to affect this process, but suggested that their 20-year record was too short to reliably diagnose a combined QBO-SC signal. Using the longer 1954–2004 record of NCEP/NCAR reanalysis, which extends only up to 10 hPa, Camp and Tung (2007b) used a linear discriminant analysis (Wilks, 2006) of February–March 10–50 hPa layer-mean $\overline{T}$ at latitudes poleward of 30°N to analyse the spatial signature of combined QBO and solar cycle influence, the pattern of which was found to resemble that of SSW events (and hence also the NAM). Defining QBO phase by 30 hPa $\overline{u}_{\mathrm{EQ}}$, their analysis showed the combined QBO-W plus SC-min group, corresponding to a colder and less disturbed vortex, to be statistically well separated from the other three groups, all of which corresponded to the NH vortex being warmer and more disturbed on average. Hence they suggested that either QBO-E or SC-max may disturb the vortex by triggering a SSW event, but that once disturbed the vortex becomes insensitive to further perturbations; thus the influence of QBO-E and SC-max is not linear (i.e. not simply additive), which may explain the weak correlation between late-winter T_{NP} and the solar cycle that occurs during QBO-E winters (Labitzke, 1987; Gray *et al.*, 2001b; Labitzke, 2005). In this framework, as noted by Camp and Tung (2007b), stratification of the data by QBO phase was not required to reveal the solar cycle signal apart from the necessity of using the QBO-W plus SC-min group to define the reference state (of a quiescent vortex).

Figure 5 attempts to summarize the combined QBO-SC influence on the vortex as seen in the updated record, albeit using a different vortex metric ($\overline{u}$ at 10 hPa, 60°N) than those employed by Labitzke and van Loon (1988) or Camp and Tung (2007b). The QBO phase is defined for the January–February period by 50 hPa $\overline{u}_{\mathrm{EQ}}$ and for the February–March period by 30 hPa $\overline{u}_{\mathrm{EQ}}$, following the levels used by Labitzke and van Loon (1988) and Camp and Tung (2007b) respectively. Comparison of the mean states of the four QBO–SC groups (Figure 5(c, d)) shows that the following features occur during both Jan–Feb and Feb–Mar periods: W-min emerges as the least disturbed vortex state, the W–E difference is larger during SC-min than SC-max, and the sign of the SC difference changes with the phase of the QBO (i.e. under QBO-W, SC-max yields a more disturbed vortex than SC-min, and vice versa under QBO-E). These features are indicated schematically by arrows in Figure 5(c, d), which point in the direction of increasing vortex disturbance: all arrows point away from W-min, and upper and lower arrows point in opposite directions. During Feb–Mar the contrast amongst the more disturbed vortex states (i.e. all states except for W-min) appears less dramatic than that seen during Jan–Feb, suggesting consistency with Camp and Tung (2007b). Under SC-max, the sign of the W–E difference changes between the Jan–Feb and Feb–Mar periods, but the magnitude of the W–E difference under SC-max is weak during both periods and hence may simply reflect noise. An indication of the sampling variability is given in Figure 5(a, b): binning the daily data into smaller groups, rather than the coarse-grain view implied by the four-state QBO–SC phase space, illustrates the strong underlying variability of the vortex. It also indicates the potential sensitivity of the Figure 5(c, d) results to changes in the thresholds that separate the four QBO–SC groups.

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The lack of data in some interior regions of Figure 5(a, b), indicated by grey, should also be noted; these regions of the QBO–SC phase space have not been sampled by the observational record.

2.5. *El Niño Southern Oscillation*

The polar vortex appears to also be influenced by ENSO, complicating detection of the QBO response (including QDV). For the 1963–1977 period, similar to the 1962–1977 period considered by HT80, van Loon *et al.* (1982) found warm ENSO (WENSO) to favour a warm and weak winter (DJF) stratospheric vortex while cold ENSO (CENSO) favoured the opposite state; this result was based on eleven extreme ENSO events in this 15-year period (compared with 15 out of 16 of HT80's years classified as QBO-W or QBO-E), and it was noted that five out of seven of HT80's QBO-E winters coincided with WENSO events. Using 1957–1978 data, Wallace and Chang (1982) considered the potential influence of both ENSO and QBO on an equal footing and found an ENSO signal consistent with van Loon *et al.* (1982) and a QBO signal consistent with HT80. However as already noted in section 2.1 they found that WENSO and QBO-E tended to coincide, as did CENSO and QBO-W, preventing separation of the two signals.[‖] Comparing ENSO and QBO signals for the DJF period in 1957–1985 data, van Loon and Labitzke (1987) showed that while the ENSO signal remained consistent with previous results, separation from the QBO remained an issue; they did, however, show a QBO signal consistent with HT80 under neutral ENSO conditions (i.e. filtering out strong WENSO and CENSO events), and also emphasized the zonal asymmetry of the ENSO response as being distinct from the QBO response, which appeared more zonally symmetric. Updating these findings for the 1957–1991 period, Hamilton (1993) confirmed the zonally asymmetric ENSO response, but found no statistically significant $\overline{\Phi}$ ENSO response poleward of 20°N, in contrast to a significant zonal-mean QBO response (defined by 50 hPa QBO phase). Hamilton (1993) noted that if one assumed the QBO response to be physically real (motivated by the QBO yielding stronger correlations than ENSO), then filtering out the QBO response reduced the significance of the ENSO signal, suggesting that 34 years of data were insufficient to separate the two effects. Baldwin and O'Sullivan (1995), using 1964–1993 data, also found the DJF NH vortex QBO response to be stronger than the ENSO response. They supposed that ENSO's extratropical influence would manifest itself through three major patterns of NH tropospheric variability – the Pacific North America (PNA), Western Pacific Oscillation (WPO) and Tropical/Northern Hemisphere (TNH) modes – known to correlate with ENSO activity. Compared to the QBO response (defined by 40 hPa QBO phase), the signature of these tropospheric modes was found to be weak, mainly confined to the zonal wave components and decaying with altitude, whereas the QBO signal was strongest in the zonal mean and increased with altitude. Overall, these studies indicated a weak ENSO influence on the NH winter stratosphere during DJF.

More recent studies have detected more robust ENSO responses. For the 1959–2005 period, Camp and Tung (2007a) performed linear discriminant analysis of 10–50 hPa layer-mean $\overline{T}$ at latitudes poleward of 30°N and found a strong ENSO signal in the NH extratropical stratosphere during DJF (rather than February–March, for which Camp and Tung (2007b) used the same analysis method to separate QBO and SC effects). They found the latitudinal structure of ENSO-induced vortex anomalies to be distinct from that of the QBO-SC response, insofar as it bore less resemblance to a characteristic SSW pattern; to explain the discrepancy, they speculated that ENSO might act to induce warmings at higher altitudes than those induced by QBO-SC. Garfinkel and Hartmann (2007) used composite analysis to examine ENSO and QBO effects in 1957–2007 data. Defining the vortex by area-weighted $\overline{T}$ poleward of 70°N at 10 hPa, they showed the November–February ENSO signal for all years – i.e. without separating by QBO phase – to be highly significant. The QBO signal for October-January under neutral ENSO conditions also showed high significance, but obtaining an ENSO signal under neutral QBO conditions proved difficult because of the bimodality of the 50 hPa QBO index used (although a WENSO signal under neutral QBO was found to be significant). Further attempting to separate the two influences, a comparison of the four possible combinations of phases was performed. A significantly warmer vortex resulted from ENSO under QBO-W conditions (i.e. WENSO minus CENSO, limited to years with QBO-W) and from QBO under CENSO (i.e. QBO-E minus QBO-W, limited to years with CENSO); the similarity of this finding to that of Camp and Tung (2007b), wherein a vortex-disturbing influence (e.g. QBO-E, WENSO, or SC-max) only becomes apparent when compared against a suitably non-perturbed state (e.g. under QBO-W, CENSO and SC-min), was noted. Similarly, Wei *et al.* (2007) showed for the 1957–2002 period that the high-latitude QBO response (with QBO phase defined by 50 hPa DJF $\overline{u}_{\mathrm{EQ}}$) was highly significant during CENSO events but nonexistent during WENSO events, suggesting the Holton–Tan relation to be disrupted by WENSO-induced upward fluxes of tropospheric wave activity. Further exploring the possible tropospheric avenues for ENSO influence, Garfinkel and Hartmann (2008) – updating and expanding earlier work by Baldwin and O'Sullivan (1995) – found a highly significant stratospheric response to the PNA mode, and argued this to be the predominant mechanism by which ENSO influenced the vortex. Interestingly, it was also noted that WENSO exhibited different extratropical tropospheric teleconnections under QBO-W (mainly PNA) than under QBO-E (mainly WPO and TNH) in the observational record.

2.6. *Summary of observational studies*

The observed NH and SH polar vortex responses to the QBO are characterized by vertically deep hemispheric-scale fluctuations of the stratospheric wintertime circulation (Figures 3, 4). This suggests that QBO influence modulates the natural (or 'internal') large-scale variability of the extratropical winter stratosphere (Baldwin and Dunkerton, 1991; Dunkerton and Baldwin, 1992; Kodera, 1993), which has been shown to covary with annular-mode tropospheric

[‖]There is no known mechanism by which the phase of ENSO and the QBO would be linearly related. Modulation of the wave forcing of the QBO by ENSO-driven fluctuations in equatorial wave activity may affect the rate of QBO phase progression (Geller *et al.*, 1997; Taguchi, 2010), but this would not create a linear correlation between ENSO and QBO phases.

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variability (Baldwin and Dunkerton, 1999, 2001; Thompson and Wallace, 2000). This vertical coupling presumably underlies the surface signature of the QBO (Figure 2), which has been found in spectral analysis of long time series of sea-level pressure (100 years) and surface air temperature (150 years) by Coughlin and Tung (2001)**, and through which a connection between the QBO and NH winter weather regimes comparable in strength to the influence of ENSO has been demonstrated (Thompson *et al.*, 2002). The extratropical QBO signal is often optimized by choosing a specific vertical level to define QBO phase (e.g. 50 hPa QBO winds for the NH), which implies that the magnitude and timing of apparent QBO influence is sensitive to the chosen definition of QBO phase. The periodicity of the QBO in time and altitude (Figure 1(a, b)) implies that this choice is not unique, but unfortunately it is not known *a priori* which QBO altitudes most strongly influence the extratropical response. The strength of the correlation between 50 hPa equatorial winds and NH stratospheric polar vortex strength originally found by HT80 in their 16-year dataset has to a degree been retained in the longer observational records (mainly provided by global reanalysis datasets) that are now available. Based on the longer record, it has been argued that the strength of QBO-vortex coupling fluctuates on decadal timescales (Kodera, 1993; Lu *et al.*, 2008). Such QDV, if it is real rather than simply an artefact of sampling or of the definition of QBO phase, may be due to the 11-year solar cycle (Labitzke and van Loon, 1988; Labitzke, 2005; Labitzke *et al.*, 2006), ENSO (Camp and Tung, 2007a; Garfinkel and Hartmann, 2007, 2008), or internal stratospheric variability (Anstey and Shepherd, 2008; Christiansen, 2010), or some other as-yet undemonstrated mechanism. A persistent difficulty faced by observational studies of stratospheric vortex variability remains the statistical separation of these competing influences.

3. Modelling studies

Faced with the complications of establishing cause-and-effect relations in an open system such as the atmosphere, models provide a means to test dynamical hypotheses by isolating causal factors. It will now be apparent, from the studies examined in the preceding section, that detection of a high-latitude QBO response is complicated by the lack of a theory that would predict such a response from first principles as well as by the existence of competing influences such as ENSO and the 11-year solar cycle. Modelling studies have attempted to address both of these issues, and accordingly we partition our review of modelling studies into two general categories: mechanistic studies, which have tended to examine the first issue (basic response to a tropical wind perturbation $\Delta\overline{u}_{\mathrm{EQ}}$; section 3.1), and GCMs, which have usually examined the second (QBO influence combined with competing influences; section 3.2). 'Mechanistic' here denotes a simplified model intended to capture essential mechanisms governing large-scale dynamics. A mechanistic model may, for example, integrate the primitive equations on a sphere (as a GCM would) but employ simpler representations of physical processes, such as Newtonian cooling toward a prescribed radiative equilibrium profile (instead of the more complex spectral parametrization of radiative processes that would be used in a GCM) and an imposed wave-like Φ perturbation at a lower boundary taken to be the tropopause (in lieu of explicit representation of tropospheric dynamics as would occur in a GCM). In contrast, GCMs arguably offer – at the expense of higher computational cost – a more realistic representation of processes such as the tropospheric forcing of upward propagating waves that perturb the stratosphere, of other known sources of interannual variability besides the QBO (solar cycle, ENSO, etc.), and of the possible influence of the stratosphere on the troposphere (including manifestations of the QBO in surface climate). This extra complexity can make the results of GCM experiments difficult to interpret; because of their relative simplicity, mechanistic models are naturally suited to the testing of specific dynamical hypotheses.

3.1. *Mechanistic models*

The hypothesis that SSWs are affected by the QBO phase was first tested in a mechanistic model by Bridger (1984), who used a global three-dimensional (3D) primitive equations model of the middle atmosphere with a lower boundary at 100 hPa. Upward-propagating planetary waves were forced by imposing, in the Extratropics, a fixed zonally asymmetric perturbation at the lower boundary of geopotential height amplitude $\Phi_0 = 350$ m and zonal wavenumbers $k = 1$ or 2. The model was run under conditions representative of NH winter, and a $\Delta\overline{u}_{\mathrm{EQ}}$ perturbation was imposed as either a deep westerly or easterly layer; the depth of the layer was ≈ 40 km, and it occupied the altitude range $z \approx 10$–50 km. No evidence for a connection between QBO phase and SSW occurrence was found. The problem was revisited by Holton and Austin (1991), using a similar model with a fixed $k = 1$ perturbation at the model's 100 hPa lower boundary, run under January–March (JFM) radiative conditions, with imposed westerly and easterly $\Delta\overline{u}_{\mathrm{EQ}}$ in the lower stratosphere (≈ 15 km deep, located at $z \approx 20$–35 km). They found that for strong wave forcing ($\Phi_0 \geq 400$ m) there was no apparent relation between SSW occurrence and the sign of $\Delta\overline{u}_{\mathrm{EQ}}$, similar to Bridger's (1984) results. Upon introducing variations in the forcing amplitude Φ_0 – which could be thought of as representing variations in tropospheric wave forcing – SSWs were found to be sensitive to $\Delta\overline{u}_{\mathrm{EQ}}$ for an intermediate range of Φ_0. For small wave forcing amplitudes ($\Phi_0 \leq 200$ m) no SSWs occurred at all, irrespective of $\Delta\overline{u}_{\mathrm{EQ}}$. Hence their results suggested that while very quiescent or very disturbed NH winters are insensitive to the tropical wind state, there is an intermediate range of forcing amplitudes for which a sensitivity to QBO phase does exist. They found that $\Delta\overline{u}_{\mathrm{EQ}} < 0$ tended to hasten the onset of SSWs, apparently in agreement with the sign of the observed Holton–Tan effect.

Subsequent and similar mechanistic modelling studies have tended to support Holton and Austin's (1991) results. O'Sullivan and Young (1992) found that positioning $\Delta\overline{u}_{\mathrm{EQ}}$ in the lower stratosphere ($z \approx 20$–35 km) yielded a stronger extratropical response than when $\Delta\overline{u}_{\mathrm{EQ}}$ was shifted upwards by 9 km; they also found a strengthened response for

**There are no reliable measurements of tropical stratospheric winds extending as far back as 100 or 150 years; Coughlin and Tung (2001) identified a spectral peak at the QBO frequency in sea-level pressure and surface air temperature data projected onto the annular mode, and attributed its existence to QBO influence. A reconstruction of the QBO back to 1900 has been created by Brönnimann *et al.* (2007), based mainly on the QBO signature in surface pressure due to its modulation of the solar semi-diurnal tide.

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increased latitudinal width of $\Delta\overline{u}_{\mathrm{EQ}}$ (an increase of $\approx 4^\circ$ in the half-width led to roughly a doubling of the extratropical response) as well as increased vertical depth of $\Delta\overline{u}_{\mathrm{EQ}}$. They emphasized the importance of the development of a nonlinear Rossby wave critical layer – which is often referred to as the 'stratospheric surf zone' (as described in section 1) – in communicating the influence of $\Delta\overline{u}_{\mathrm{EQ}}$ to high latitudes, showing that the case of small Φ_0 led to absorption of waves in a thin low-latitude critical layer and consequently no propagation of the signal to high latitudes. O'Sullivan and Dunkerton (1994) imposed a $\Delta\overline{u}_{\mathrm{EQ}}$ similar to that of O'Sullivan and Young (1992), and were the first to carefully consider the role of the seasonal cycle of radiative forcing. They noted that by focusing attention on NH late winter (January–February), Holton and Austin (1991) had avoided the period during which the observed QBO signal is strongest, and that O'Sullivan and Young (1992) used unrealistically weak radiative forcing. O'Sullivan and Dunkerton (1994) found, for perpetual solstice integrations, that Φ_0 behaved as a bifurcation parameter and that the value of Φ_0 separating quiescent from disturbed winters was smaller when $\Delta\overline{u}_{\mathrm{EQ}} < 0$ than when $\Delta\overline{u}_{\mathrm{EQ}} > 0$ (i.e. stronger wave forcing was required to produce a disturbed winter for their 'QBO-W' case). When a seasonal cycle was introduced, the location (in terms of Φ_0) of the bifurcation point became somewhat smeared out. They also showed that including parametrized gravity wave drag in their model tended to reduce the strength of the vortex response to realistic levels: without parameterized drag, the model response reflected a contrast between a disturbed polar vortex and a vortex in radiative equilibrium, yielding an extratropical $\overline{u}$ difference far in excess of the observed signal. A failing of their model, they noted, was that it produced a vortex response only in December–February, whereas the observed response also occurs earlier in winter – a deficiency shared by previous studies (Holton and Austin, 1991; O'Sullivan and Young, 1992).

Two-dimensional (2D) single-layer models – intended to represent a deep, vertically averaged stratospheric layer – have also been used to examine QBO–vortex coupling. Such models have been shown to give useful insight into the behaviour of nonlinear Rossby wave critical layers (Juckes and McIntyre, 1987), or Rossby wave breaking (RWB), which may communicate the influence of $\Delta\overline{u}_{\mathrm{EQ}}$ to high latitudes. O'Sullivan and Salby (1990) used a single-layer model with radiative damping under perpetual January conditions to investigate the effect of $\Delta\overline{u}_{\mathrm{EQ}}$ on the behaviour of the critical layer. They noted that the large latitudinal scale of such a critical layer, spanning tens of degrees of latitude under realistic NH winter forcing conditions, provides a plausible means of coupling between low and high latitudes. But the vortex response depends also on the radiative damping, which opposes the tendency of RWB to displace and distort the polar vortex by relaxing potential vorticity gradients towards a zonally symmetric radiative equilibrium state (Butchart and Remsberg, 1986). In O'Sullivan and Salby's (1990) model, the ability of RWB to overcome the radiative drive and break up the vortex was more pronounced (for a given Φ_0) when the $\overline{u}(\phi) = 0$ line occurred further poleward in the NH.

Neglecting radiative effects, the response of high latitudes to $\Delta\overline{u}_{\mathrm{EQ}}$ was studied in adiabatic one-layer models by Chen (1996) and O'Sullivan (1997). Chen (1996) found that the strength of the subtropical $\overline{u}$ minimum in the winter hemisphere could affect the location and intensity of the stratospheric surf zone and was capable of shielding high latitudes from the influence of $\Delta\overline{u}_{\mathrm{EQ}}$. O'Sullivan (1997) focused on the effect of RWB on the QBO jet itself (i.e. the tropical stratospheric winds), finding that RWB tended to occur on the flanks of the QBO rather than at the Equator, but also illustrated the vortex erosion caused by RWB. Whether RWB weakens or strengthens the vortex depends on the presence of radiative damping. Polvani *et al.* (1995) contrasted the nature of vortex erosion and surf zone structure under both adiabatic and radiatively damped conditions. While not considering QBO–vortex coupling specifically, they showed that when their shallow-water model was initialized with a $\overline{u}(\phi)$ profile typical of NH winter mid-stratospheric conditions, with the $\overline{u}(\phi) = 0$ line at $\phi \approx 20^\circ$N, the combined effects of RWB and radiative restoration acted to strengthen the vortex and maintain the jet peak at its original latitude of $\phi \approx 55^\circ$N (Polvani *et al.*, 1995, Figure 9). This behaviour contrasted strongly with their adiabatic case, for which 'the vortex has no means of withstanding the erosion and simply retreats to higher latitudes'. (Similar behaviour was found by O'Sullivan, 1997). Thus stronger RWB led to a stronger vortex, provided that radiative relaxation was present.

As well as varied radiative forcing conditions, mechanistic studies have considered various idealized forms for $\Delta\overline{u}_{\mathrm{EQ}}$. The 3D-modelling studies cited above (Bridger, 1984; Holton and Austin, 1991; O'Sullivan and Young, 1992; O'Sullivan and Dunkerton, 1994) all imposed $\Delta\overline{u}_{\mathrm{EQ}}$ as a deep, single layer of either westerly or easterly winds. The real QBO typically exhibits oppositely-signed $\overline{u}$ at lower and mid-stratospheric levels with strong vertical shears separating the two regions, as shown in Figure 1 for $\overline{u}_{\mathrm{EQ}}$ and in Figure 4 for the vertical structure of W–E $\Delta\overline{u}_{\mathrm{EQ}}$. (Holton and Austin, 1991, did prescribe upper and lower $\Delta\overline{u}_{\mathrm{EQ}}$ of opposite sign, but their upper cell was weak and did not induce a shift of the $\overline{u}(\phi, z) = 0$ line between hemispheres; hence their $\Delta\overline{u}_{\mathrm{EQ}}$ resembled a single-layer perturbation.) While a deep layer is a useful idealization for studying the first-order effect of $\Delta\overline{u}_{\mathrm{EQ}}$ on the Extratropics, its use makes the connection between these studies and the observed QBO signal ambiguous. Such studies also neglect the downward propagation of QBO phase (Figure 1(a)). In a later study, Gray *et al.* (2001a) relaxed a 3D mechanistic model towards observed tropical radiosonde winds (over a 33-year period, 1964–1997) and tropical rocketsonde winds (over a 23-year period, 1964–1987). Radiosonde data do not extend higher than ≈ 32 km (10 hPa), while rocketsondes reach as high as ≈ 58 km (0.3 hPa); hence Gray *et al.*'s (2001a) study contrasted the QBO influence for cases of shallow ($z \approx 16$–32 km) and deep ($z \approx 16$–58 km) $\Delta\overline{u}_{\mathrm{EQ}}$ – and, additionally, incorporated realistic QBO phase propagation. They found the vortex response to be stronger for the case of deep $\Delta\overline{u}_{\mathrm{EQ}}$, suggesting that the lowermost equatorial stratosphere may not be the most important region from which low-latitude influence extends to higher latitudes. Gray *et al.* (2001a) ruled out the possibility that the response (in their model) could be caused *only* by $\Delta\overline{u}_{\mathrm{EQ}}$ at higher altitudes, finding that no vortex response occurred when equatorial winds were relaxed to observed winds in the 40–58 km and 30–58 km ranges (consistent with O'Sullivan and Young's, 1992, results). These results suggested the importance of $\Delta\overline{u}_{\mathrm{EQ}}$ over the whole depth of the stratosphere.

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Further exploring the effect of a very deep $\Delta\overline{u}_{\mathrm{EQ}}$, Gray *et al.* (2003) imposed a single-layer $\Delta\overline{u}_{\mathrm{EQ}}$ over the range $z \approx 16-65$ km, using the same 3D model as Gray *et al.* (2001a). Similar to the results of Holton and Austin (1991), they found that polar variability was significantly influenced by $\Delta\overline{u}_{\mathrm{EQ}}$ for an intermediate range of Φ_0 values (with $\Delta\overline{u}_{\mathrm{EQ}} < 0$ leading to more disturbed winters). They emphasized that extratropical planetary waves have deep vertical wavelengths, roughly the depth of the stratosphere, and hence should be sensitive to stratospheric $\overline{u}$, including $\Delta\overline{u}_{\mathrm{EQ}}$, over the whole stratosphere. In another study using the same model, Gray (2003) contrasted the effects of $\Delta\overline{u}_{\mathrm{EQ}}$ at $z \approx 20-40$ km and $z \approx 40-60$ km, finding the midwinter vortex to be most sensitive to the upper ($z \approx 40-60$ km) $\Delta\overline{u}_{\mathrm{EQ}}$, while the lower $\Delta\overline{u}_{\mathrm{EQ}}$ mainly affected the early winter evolution. Gray *et al.* (2004) considered yet another form of $\Delta\overline{u}_{\mathrm{EQ}}$, imposing easterly anomalies of ≈ 10 km depth at a series of altitudes that were moved upwards in 5 km increments: 20–30, 25–35, 30–40, 35–45, and 40–50 km. (The anomaly was strongest in a 10 km layer; tapering of the anomaly above and below this layer led to the actual layer of imposed easterly winds occupying ≈ 14 km depth.) The most pronounced effect occurred in the 30–40 km experiment, with the poleward-shifted $\overline{u}(\phi, z) = 0$ line inducing the vortex to cool and strengthen initially, followed by a SSW. All 20 ensemble members in the 30–40 km experiment followed this very regular behaviour, while the vortex variability was somewhat more random in the other cases (Gray *et al.*, 2004, Figure 4); hence the major effect of $\Delta\overline{u}_{\mathrm{EQ}}$ in this case was to regularize the vortex variability (i.e. SSWs still occurred, but their timing was much less random than in the other $\Delta\overline{u}_{\mathrm{EQ}}$ cases).

One issue that complicates the study of QBO–vortex coupling using observations, as well as GCMs, is the fact that stable statistics of polar vortex interannual variability may require very long records (Yoden *et al.*, 2002). Here the lower computational cost of mechanistic models gives them an advantage over GCMs. Naito *et al.* (2003) used a 3D primitive-equations model, including a simplified troposphere, with $k = 1$ surface topography and performed 12 000-day integrations under nine different idealized $\Delta\overline{u}_{\mathrm{EQ}}$ cases. The vertical structure of $\Delta\overline{u}_{\mathrm{EQ}}$ was single-celled and strongest in the $z \approx 20-30$ km layer. A weaker vortex occurred for $\Delta\overline{u}_{\mathrm{EQ}} < 0$, as well as for extremely strong ($> 40\ \mathrm{m\,s^{-1}}$) positive $\Delta\overline{u}_{\mathrm{EQ}}$, and index of refraction arguments were used to explain this result. Naito and Yoden (2006), also using very long integrations, examined the effect of $\Delta\overline{u}_{\mathrm{EQ}}$ on planetary wave structures associated with SSWs. Eight $\Delta\overline{u}_{\mathrm{EQ}}$ cases were considered, with the vertical structure of $\Delta\overline{u}_{\mathrm{EQ}}$ in the different cases being somewhat representative of the QBO at different stages of its downward propagation (although $\Delta\overline{u}_{\mathrm{EQ}}$ was strongest in the lower stratosphere and extended to an unrealistically low altitude, occupying the layer between 200 hPa and 5 hPa). Equatorward planetary wave propagation was enhanced for the $\Delta\overline{u}_{\mathrm{EQ}} > 0$ cases, resulting in weaker warmings occurring for the same amount of upward-propagating wave activity (i.e. wave forcing was less efficient in forcing SSWs for the $\Delta\overline{u}_{\mathrm{EQ}} > 0$ cases). Also noted was a statistically significant difference in upward EP flux in the troposphere at $\phi \approx 45°$N: both before and after SSW events, greater upward flux occurred for the $\Delta\overline{u}_{\mathrm{EQ}} > 0$ cases, although latitudinally averaged tropospheric flux (over 30°N–86°N) was less well correlated with polar temperatures in those cases. These results echo a point emphasized by Gray *et al.* (2003): due to the large vertical wavelength of planetary waves, it is probably wrong to imagine the wave source (in the troposphere) as being completely separate from the wave sink (the stratosphere). The planetary wave response to a given $\Delta\overline{u}_{\mathrm{EQ}}$ evidently can involve changes in wave structure (hence, in EP flux) over a deep vertical scale encompassing both the stratosphere and troposphere.

Mechanistic models have largely not been used to examine the combined influence of the QBO and solar cycle on the vortex, where a large sample size may be useful in separating the two influences; an exception is Ito *et al.* (2009). Using the same model as Naito and Yoden (2006), they obtained a large ensemble (39 600 days for each of 46 types of QBO–SC forcing conditions) of perpetual-winter integrations with which to examine the NH response (cf. Figure 5(a): the reanalysis period contains roughly 3200 Jan–Feb days and is most likely a non-ergodic sampling of the QBO–SC phase space). Radiative conditions were specified as Newtonian cooling toward a radiative equilibrium state $T_{\mathrm{rad}}(\phi, z)$, with T_{rad} representative of late winter SC-min conditions, and SC-max was specified as a stratopause-centred perturbation of this T_{rad}. Under QBO-W (E) conditions, SC-max was found to warm (cool) the vortex relative to SC-min, but the statistical significance of the SC signal was higher for the QBO-W group. Moreover, the vortex response to the SC was generally weak (≈ 2 K or smaller at 86°N) compared to the QBO response (≈ 5 K), albeit detectable due to the very large sample size considered. (It should be noted, however, that the chosen $\Delta\overline{u}_{\mathrm{EQ}}$ represented the most extreme W and E cases considered by Naito and Yoden, 2006.) Ito *et al.* (2009)'s results were thus consistent with the observational studies by Labitzke *et al.* (2006) and Camp and Tung (2007b) showing SC-max to warm the NH vortex during QBO-W but to weakly cool it, or have ambiguous effect, under QBO-E conditions. (Interestingly, Figure 3 of Ito *et al.* (2009) showed the sign of the max–min difference under QBO-E to depend on the meridional extent of the imposed SC forcing.)

Besides Gray *et al.* (2001a), the only other mechanistic modelling study to consider a downward-propagating QBO is that of Hampson and Haynes (2006), who specified $\Delta\overline{u}_{\mathrm{EQ}}$ in the $z \approx 15-35$ km layer with a two year period. They used a zonally truncated model (retaining only the zonal mean and the $k = 1$ harmonic) so as to allow a large number of integrations, and varied the phase alignment of their QBO with respect to the annual cycle in order to explore the influence of the vertical structure of $\Delta\overline{u}_{\mathrm{EQ}}$ on the extratropical response. QBO phase alignment was found to strongly affect the strength of the response; further manipulations to $\Delta\overline{u}_{\mathrm{EQ}}$ demonstrated early winter to be the most influential time for the QBO to affect the mid-to-late winter response. These results suggest the importance of the seasonal cycle in determining QBO influence; while the importance of early winter conditions for determining subsequent vortex evolution has been argued elsewhere (Scott and Haynes, 2002), many mechanistic QBO–vortex coupling studies have neglected the evolution of the seasonal cycle. Also notable is that Hampson and Haynes (2006) obtained a strong response with $\Delta\overline{u}_{\mathrm{EQ}}$ specified only in the lower stratosphere, in contrast to earlier studies by Gray and collaborators arguing the importance of $\Delta\overline{u}_{\mathrm{EQ}}$ spanning the depth of the stratosphere. But similarly to some earlier studies (e.g. O'Sullivan and Young, 1992; Gray *et al.*, 2001a),

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they found that upward movement of a $\Delta\overline{u}_{\mathrm{EQ}}$ of fixed vertical depth (of ≈ 10 km) weakened the extratropical response.

In summary, mechanistic modelling studies appear to demonstrate that a range of vortex responses to $\Delta\overline{u}_{\mathrm{EQ}}$ are possible. Precise comparisons of the different studies are difficult because of the varied nature of choices made for $\Delta\overline{u}_{\mathrm{EQ}}$, which sample a variety of possible vertical structures. However, all studies agree that a poleward-shifted $\overline{u}(\phi, z) = 0$ line may affect the polar vortex when the planetary wave forcing is of intermediate strength. Although the effect of poleward-shifted $\overline{u}(\phi, z) = 0$ is generally to weaken the vortex, the precise meaning of this statement depends on the averaging period that is used to define the vortex strength. Gray *et al.* (2004) showed that poleward-shifted $\overline{u}(\phi, z) = 0$ may initially strengthen the vortex (followed, with high predictability, by warmings), while Polvani *et al.* (1995) demonstrated that the combined effects of radiative damping and RWB may strengthen the vortex when the $\overline{u}(\phi, z) = 0$ line is located in the NH. Mechanistic studies have also shown that the strength and timing of the influence of $\Delta\overline{u}_{\mathrm{EQ}}$ on the vortex may be sensitive to the vertical structure of $\Delta\overline{u}_{\mathrm{EQ}}$ over the whole depth of the stratosphere, as well as the seasonal timing of $\Delta\overline{u}_{\mathrm{EQ}}$.

3.2. *General Circulation Models*

As noted in section 1, it is gradually becoming more common for GCMs to exhibit spontaneous QBOs, which is to say that a model resolves an oscillation in tropical stratospheric $\overline{u}$ that is driven by resolved and/or parametrized waves, and that the amplitude, period and descending shear zones of this oscillation resemble those of the observed QBO (Figure 1(a)). Spontaneous QBOs usually exhibit some unrealistic features, and since the composition of the wave spectrum forcing the QBO remains uncertain (Baldwin *et al.*, 2001; Alexander *et al.*, 2010), it is not yet well understood how to obtain a realistic QBO in a GCM for the right reasons. The results described in section 3.1 suggest that the extratropical response to the QBO is sensitive to the structure and timing of $\Delta\overline{u}_{\mathrm{EQ}}$ – i.e. the QBO amplitude, period, and vertical and latitudinal extent – and hence it is plausible that deficiencies in QBO representation may adversely affect a GCM's simulation of QBO–vortex coupling. Of course other aspects of a model, such as its extratropical climatology, are also likely to affect its high-latitude QBO response.

Alternatively, a QBO may be artificially induced in a GCM by imposing a zonal torque that relaxes the tropical $\overline{u}$ toward a specified wind profile. This allows high-latitude QBO influence to be studied using models that do not spontaneously resolve the QBO, and allows for either realistic or idealized $\Delta\overline{u}_{\mathrm{EQ}}$. Of course any feedback from the Extratropics on the Tropics is neglected in such experiments – e.g. the effect of changes in tropical upwelling forced by extratropical wave drag on the rate of change of QBO phase (Hampson and Haynes, 2004), or the possibility that an apparent solar cycle response in the tropical lower stratospheric temperature is caused by variations in the mean meridional circulation associated with modulations of SSW activity (Gray *et al.*, 2006). In this section we review GCM studies of QBO–vortex coupling that have utilized either spontaneous or imposed QBOs. Studies in which a spontaneous QBO was obtained in a GCM, but QBO influence at high latitudes was not examined, are not considered here.

Two early GCM studies are those of Hamilton (1998) and Niwano and Takahashi (1998). Hamilton (1998) used an imposed zonal torque to force a QBO with properties closely matching those of the observed QBO; a regular 2.25 year period was chosen. Niwano and Takahashi (1998) utilized a spontaneously generated, and hence somewhat less realistic, QBO: using a GCM with 0.5 km vertical resolution, they obtained a QBO with a 1.3 year period and a westerly bias that was particularly strong in the 50–80 hPa layer. Both studies found the polar vortex to be weaker during QBO-E, although not during November. (DJF and JFM responses were examined, respectively, by the two studies.) Hamilton (1998) obtained more robust statistics by employing a 48-year model run as opposed to Niwano and Takahashi's (1998) 15-year run, yet emphasized that a long record may be required for stable statistics of the high-latitude QBO signal to be obtained: the spatial structure of correlations between the QBO ($\Delta\overline{u}_{\mathrm{EQ}}$ at 40 hPa) and the Extratropics ($\overline{u}$ at all other gridpoints) differed between the first 23 years and the last 22 years of the 45-year analysis period, indicating significant sampling variability in a GCM run of roughly the same length as available reanalyses (e.g. ERA-40).

Other GCM studies that have considered QBO–vortex coupling as an internal mode of atmospheric variability (i.e. without considering external forcings such as ENSO or the SC) are Pascoe *et al.* (2006), Calvo *et al.* (2007) and Anstey *et al.* (2010). Such studies are useful because they allow for unambiguous detection of QBO influence, as well as its possible internally generated QDV. Like Hamilton (1998), Pascoe *et al.* (2006) used an imposed forcing to relax equatorial winds towards specified semiannual oscillation, shallow QBO (with $\Delta\overline{u}_{\mathrm{EQ}}$ occupying the range $z \approx 15$–35 km), and deep QBO (with $\Delta\overline{u}_{\mathrm{EQ}}$ occupying the range $z \approx 15$–65 km) states, performing a 28-year run for each case. The timing of midwinter SSWs was advanced in the deep QBO case – i.e. midwinter SSWs occurred earlier in the season for deep QBO than for shallow QBO. Early winter (November) warmings were equally affected by both shallow and deep QBOs, suggesting that the upper QBO altitudes ($z \approx 35$–65 km) were unimportant for the early winter response. Hence they emphasized that realistic upper stratospheric tropical wind variability may be required for GCMs to properly simulate NH midwinter vortex variability. Calvo *et al.* (2007) examined the Holton–Tan effect in a GCM with a spontaneous QBO (Giorgetta *et al.*, 2006). Using a 50-year model run containing 20 complete QBO cycles, the NH winter polar vortex was found to be stronger (weaker) when the QBO at 30 hPa was W (E). Similar to earlier GCM studies (Hamilton, 1998; Niwano and Takahashi, 1998), QBO influence occurred only from December onwards. Anstey *et al.* (2010) used a 150-year integration of a GCM with a spontaneous QBO and found different QBO phases, defined using a QBO phase angle constructed from the two leading EOFs of $\overline{u}_{\mathrm{EQ}}$ (Wallace *et al.*, 1993; Baldwin and Dunkerton, 1998), to optimize the early and late winter vortex responses in both the NH and SH. An early NH winter (November) response was detected, and it was noted that the model's realistic QDV of QBO phase transitions, due to interaction between the QBO and seasonal cycle (Dunkerton, 1990), could lead to QDV in the vortex response due only to internal atmospheric dynamics.

A number of GCM studies have focused on the combined effects of the QBO and 11-year solar cycle (SC), attempting to capture the behaviour illustrated in Figure 5. Kodera

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et al. (1991) performed ten 60-day runs under conditions of two different imposed QBO phases with somewhat realistic structure, and five different solar forcing conditions obtained by varying solar ultraviolet (UV) heating rates between 70% to 110% of their usual values (although such large variations in solar forcing are unlikely to be realistic). The combined QBO/SC vortex response was reminiscent of that observed by Labitzke and van Loon (1988). Palmer and Gray (2005) performed four 28-year runs of a GCM with imposed solar cycle forcing and a spontaneous QBO (Scaife *et al.*, 2000) and found similar results. However, like Kodera *et al.* (1991) their solar forcing was unrealistically large. Balachandran and Rind (1995) performed 3-year GCM runs with imposed QBO phases and using both strong (±25%) and weak (±5%) changes in UV heating rates to mimic SC-max,min forcings. For both weak and strong SC cases, SC-max acted to disrupt QBO–vortex coupling: polar $\overline{T}$ in the lower stratosphere ($z \approx 20$ km) was warmer under E/min than W/min (for SC-min = −5% or −25%) but showed little difference between E/max and W/max (for SC-max = +5% or +25%), consistent with the observational result of Labitzke (1987); cf. Figure 5(c). (Note that W/E here denotes $\overline{u}_{\mathrm{EQ}}$ in the $z \approx 20$–30 km layer.) Furthermore, the (arguably more realistic) ±5% case produced a combined QBO/SC modulation of lower stratospheric polar $\overline{T}$ essentially identical to that shown by Camp and Tung (2007b), with W/min giving the coolest polar $\overline{T}$ and the other three groups (E/min, W/max, E/max) giving roughly the same amount of warming relative to W/min; cf. Figure 5(d). (This can be seen from comparison of their Figures 5 and 9, which show the relevant E–W and max–min differences.) Extending the ±5% runs to ten years confirmed the robustness of this signal. Nevertheless, Balachandran and Rind (1995) noted that even their ±5% variation might be unrealistically large, and stressed the importance (and difficulty) of prescribing accurate solar variations when using a model to test the mechanisms underlying combined QBO/SC influence.

Aiming for realistic forcings in a GCM simulation of QBO/SC interactions, Matthes *et al.* (2004) imposed fixed QBO phases with realistic two-celled vertical structure (over the range $z \approx 20$–40 km) and solar cycle radiative forcing of realistic magnitude and spectral dependence, as well as prescribed solar-induced ozone changes.[††] Performing one 15-year run for each of the four QBO/SC cases, the results were similar to Balachandran and Rind (1995) in being consistent with Camp and Tung (2007b): T_{NP} at 10 hPa showed a distinct E–W difference in Jan–Feb under SC-min conditions (i.e. E/min warmer than W/min), but negligible E–W difference under SC-max (their Figure 10; E/W here denotes $\overline{u}_{\mathrm{EQ}}$ in the $z \approx 20$–30 km layer; cf. also Figure 5(c)). However, using six 20-year runs of a chemistry-climate model (CCM – i.e. a GCM with interactive chemistry), again with forcings designed to be realistic, Matthes *et al.* (2010) found that high-latitude lower stratospheric $\overline{T}$ during NH winter showed no response to solar forcing under either QBO-W or E conditions, although SH winter showed a statistically significant response (maximizing in Sept–Oct) that changed sign depending on the QBO phase. A different model was used here than in Matthes *et al.* (2004), and it was suggested that this model's unrealistic climatology and SSW frequency might explain the lack of a NH response.

It is becoming increasingly common that more comprehensive models – such as CCMs, which incorporate interactive chemistry as in Matthes *et al.* (2010) – are used to address various aspects of QBO high-latitude influence. Shindell *et al.* (1999) used a GCM with parametrized heterogeneous ozone chemistry to assess the effects of imposed steady QBO phases on the SH ozone hole. Lower $\overline{T}$ in the lower stratosphere during QBO-W caused ≈ 40% more ozone to be destroyed than during QBO-E, with the QBO signal not swamped by tropospheric variability due to the highly nonlinear (i.e. threshold T-dependence) nature of the chemistry involved in ozone depletion; the magnitude of QBO-induced variability compared well with the observed ozone variability. A strong QBO modulation of the response to volcanic eruptions was shown in case-studies of GCM integrations by Thomas *et al.* (2009), who additionally noted that QBO influence in NH winter sometimes peaked prior to the DJF period. Schmidt *et al.* (2010) addressed combined QBO-SC effects using a CCM with a spontaneous QBO and either perpetual SC-min or SC-max conditions, analysing 35 years for each case. During March in the NH, the SC-max case showed a stronger NH vortex under QBO-E than W (note that this behaviour is seen, weakly, in Figure 5(d)), but no significant QBO modulation of solar effects was found for earlier winter months. Similarly to Matthes *et al.* (2010), they suggested that their model's climatological state (in particular, the vortex exhibiting very strong early-winter variability) might underlie the discrepency with observations.

Another function of more comprehensive GCMs is to represent active coupling between the atmosphere and ocean. Most of the previously mentioned modelling studies have employed idealized conditions at their lower boundaries: use of prescribed climatological sea-surface temperatures (SSTs) is typical in atmosphere-only GCMs, while mechanistic models often use fixed Φ_0 as a proxy for tropospheric stationary waves. While such studies provide convincing evidence that QBO–vortex coupling is a real mode of atmospheric internal variability, it remains an important question to quantify its significance in relation to other sources of interannual variability in the climate system, such as ENSO. Calvo *et al.* (2009) used fixed-SST GCM simulations to compare the effect of warm ENSO (WENSO) conditions to that of climatological SSTs, and found the seasonal timing of WENSO-induced NH vortex warming to be delayed (advanced) when the 30 hPa QBO was W (E). Using both fixed-SST and coupled (i.e. with an interactive ocean) GCM experiments, Garfinkel and Hartmann (2010) found the extratropical tropospheric teleconnections of WENSO to depend upon QBO phase in a manner similar to (but more weakly than) that observed (Garfinkel and Hartmann, 2008), suggesting a possible tropospheric route for QBO influence on the winter stratospheric vortex.

Since most GCMs are designed to represent the troposphere well – indeed, most stratosphere-resolving GCMs were conceived as upward extensions of already-established tropospheric models – it is natural to use them to investigate QBO effects on surface climate, such as have been detected in observational studies (e.g. Coughlin and

[††]Specifically, Matthes *et al.* (2004) give the observed difference in total solar irradiance (TSI) between solar cycle maxima and minima as 0.1%, and the UV difference in the 200–300 nm range as 5%. Baldwin *et al.* (2001) quote the TSI difference as 0.1%, and the UV difference as less than 1% except at wavelengths less than 200 nm, where the difference rises to 8%. It remains unclear exactly what the correct variations in solar UV should be (Haigh *et al.*, 2010; Gray *et al.*, 2010).

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Tung, 2001; Thompson *et al.*, 2002). The extratropical surface signature of the QBO in winter (Figure 2) is generally assumed to be a manifestation of extratropical stratosphere–troposphere coupling that involves to some degree a causal role for the stratosphere (i.e. downward influence).[‡‡] Boer and Hamilton (2008) and Marshall and Scaife (2009) conducted seasonal forecasting experiments, using several different GCMs, demonstrating that knowledge of the QBO imparts additional skill to these models' predictions of NH winter climate. Boer and Hamilton (2008) quantified their results as a modest ($\approx 20\%$) amount of additional skill, beyond that imparted by ENSO, in the North Atlantic during DJF. These studies show that the extratropical tropospheric teleconnections of the QBO, manifested as a NAO-like pattern in Boer and Hamilton (2008) and a more zonally symmetric response in Marshall and Scaife (2009), yield some enhanced predictability in a region where improved seasonal forecasts are desirable.

To summarize the GCM studies, QBO–vortex coupling has been examined both on its own (as a mode of internal atmospheric variability) as well as in conjunction with the solar cycle and ENSO. Due to the inability of most GCMs to realistically represent the QBO from first principles, imposed forcing has been used to generate QBO-like $\overline{u}_{\mathrm{EQ}}$ states in many of these models, often with $\overline{u}_{\mathrm{EQ}}$ being steady in time although in some cases propagating QBOs have been used. (In comparison, recall the strong interannual variability of $\overline{u}_{\mathrm{EQ}}$ shown in Figure 1(a).) Most studies have reported QBO high-latitude influence in NH midwinter (DJF), but not in early winter (November) when a strong NH response is also observed (exceptions being Pascoe *et al.* (2006) and Anstey *et al.* (2010); Figure 3(a, c)). The relatively small number of studies that have considered the SH response (Shindell *et al.*, 1999; Anstey *et al.*, 2010; Matthes *et al.*, 2010) have found a QBO modulation of the final warming, consistent with the seasonal timing of the observed SH response (Figure 3(b, d)). Regarding the solar cycle, GCM results so far mostly support the observational inference that either QBO-E or SC-max can warm the NH vortex relative to W/min (Figure 5). Yet realistic simulation of such interactions remains in its nascent stages due to the difficulty of realistically representing any one of these physical processes (i.e. QBO, ENSO or SC) in a GCM, let alone their combined effects.

4. Discussion

Winter stratospheric polar vortex variability has been widely studied, as already noted in section 1, both for its own intrinsic interest and because of its implications for other aspects of climate, such as surface impacts (Figure 2). Both observations and models indicate that the QBO can modulate this variability in either hemisphere (as shown for reanalysis data in Figure 3). Polar vortices in the real atmosphere are influenced by many factors, of which the QBO is only one; hence these factors may be aliased in the apparent QBO response, or vice versa, as was described in section 2. It is therefore important to use models to establish the physical plausibility of such responses. Results from both mechanistic models and GCMs have tended to support the hypothesis of a causal link from low to high latitudes, with modulation of planetary wave activity being integral to this link. As described in section 3, the range of models that have verified some form of coupling between the QBO and the polar vortex is encouragingly robust.

Yet the detailed nature of QBO–vortex coupling remains unclear. While the HT80 hypothesis (section 1) is plausible, the QBO level of greatest influence remains uncertain. More generally it is unclear what type of $\Delta\overline{u}_{\mathrm{EQ}}$ most effectively induces a vortex response and whether changes in the location of the $\overline{u}(\phi, z) = 0$ line are an important factor (section 2.3; Garfinkel *et al.*, 2012), and if so at what altitudes (e.g. Naoe and Shibata, 2010; Yamashita *et al.*, 2011). Observations and GCMs show that the response varies intraseasonally and differs between the NH and SH winters (Figure 3). This indicates that the high-latitude influence of the QBO depends on the seasonal development of the winter polar vortex, which is very different in the two hemispheres; such differences presumably underly the fact that different QBO levels yield optimal responses in each hemisphere (e.g. 40 hPa for the NH and 25 hPa for the SH, as given by Baldwin and Dunkerton (1998)), although differences in the types of large-scale waves that dominate in each hemisphere may also play a role. Modelling studies have shown, mostly using mechanistic models, that a variety of types of $\Delta\overline{u}_{\mathrm{EQ}}$ may influence the vortex. Many of these studies have tested the sensitivity of the response to the vertical structure of $\Delta\overline{u}_{\mathrm{EQ}}$, but there is not yet a consensus on what vertical structure yields the strongest response. While particular single QBO levels may be used to maximize the response in the observed record, the signal thus defined has been interpreted as exhibiting decadal variability. Evidence from observations and models suggests that QDV in the vortex response might be caused by the 11-year solar cycle, ENSO, or internal stratospheric variability.

As noted in section 1, it is not well known from first principles – beyond the fact that planetary waves are important and that they break in a broad surf zone – the mechanism by which influence extends from low to high latitudes. It is through such a mechanism that the interannual 'memory' of low-latitude winds would be communicated to higher latitudes, where the seasonal transition between easterly and westerly winds erases the memory of the previous winter. Low-latitude memory in $\overline{u}$ need not be due only to the QBO: it might arise due to the weak radiative constraint on $\overline{u}$ in the Tropics (Scott and Haynes, 1998) or due to 11-year solar cycle modulation of the subtropical meridional temperature gradient, and hence the strength of the subtropical westerly jet, near the stratopause (Kodera and Kuroda, 2002). It is unlikely that the effects of different influences add linearly; rather, observations (and some modelling results) suggest that once the vortex becomes disturbed due to one of these factors, its evolution becomes less sensitive to the others (Camp and Tung, 2007b). Such behaviour is reminiscent of the vortex response to $\Delta\overline{u}_{\mathrm{EQ}}$ found in mechanistic modelling studies, where sensitivity to $\Delta\overline{u}_{\mathrm{EQ}}$ occurs only for an intermediate range of Φ_0, and extreme vortex states – either quiescent or very disturbed – are insensitive to $\Delta\overline{u}_{\mathrm{EQ}}$ (e.g. Holton and Austin, 1991; O'Sullivan and Dunkerton, 1994). The vortex

[‡‡]Other avenues for QBO influence on the extratropical troposphere may exist. It is outside the scope of this review to address these possibilities in detail, but we note that Collimore *et al.* (2003) and Liess and Geller (2012) give evidence for a QBO influence on tropical deep convection, while Garfinkel and Hartmann (2011) consider the direct response of tropospheric eddies to QBO-induced subtropical wind anomalies. Baldwin *et al.* (2001), section 6.3, indicate some other possible relationships.

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response to the QBO thus resembles an internal mode of extratropical variability (Scott and Polvani, 2004) that may be excited, but not uniquely, by the QBO (Dunkerton and Baldwin, 1992).

While no definitive picture of the mechanism of QBO–vortex coupling has yet emerged, a summary of the detailed results of observational and modelling studies of this phenomenon is pertinent to the ongoing process of GCM development and to model–measurement comparisons. Current chemistry–climate models vary considerably in their ability to represent winter polar stratospheric variability, exhibiting considerable spread in the frequency distribution of SSWs (Butchart *et al.*, 2011) and the seasonality and time-scales of annular mode variability (Gerber *et al.*, 2010). The lack of a QBO in most of these GCMs may contribute to these deficiencies. It is not obvious whether the effect of the QBO is only to 'organize' the interannual variability – i.e. to create a tendency for SSWs to occur roughly every two years – or also to affect the climatological vortex state, which itself affects the variability (Gerber *et al.*, 2010). Conversely, a biased climatological vortex state (a common GCM problem) is likely to affect the vortex response to the QBO. That the response should depend on the vortex state is suggested both by the contrast between observed NH and SH responses as well as the finding from mechanistic studies, noted above, that sensitivity to $\Delta\overline{u}_{\mathrm{EQ}}$ seems to occur for intermediate rather than extreme vortex states. For this reason, the QBO index yielding the strongest observed response might not do likewise in a model. The issue is further complicated when $\Delta\overline{u}_{\mathrm{EQ}}$ in a model does not realistically represent phases of the real QBO (which, as noted in section 3, is often true). Since it is not known which part of $\Delta\overline{u}_{\mathrm{EQ}}$ is most important for the vortex response – i.e. from what levels causality originates – it is not known which unrealistic features of $\Delta\overline{u}_{\mathrm{EQ}}$ will adversely affect the response. Hence some care must be taken when defining a QBO index (e.g. a single-level wind or a phase angle) to diagnose QBO–vortex coupling in a model: differences in the optimal index between model and observations could indicate model errors in the QBO, in the vortex, or in both. It may be useful, when making model–measurement comparisons, to account for errors in $\Delta\overline{u}_{\mathrm{EQ}}$ so as to attempt to distinguish these two types of problems; we hope that the summary of the observational studies given here (section 2) will help to facilitate such comparisons.

Based on roughly half a century of observations as well as the cumulative evidence from a variety of models, it seems very likely that there is a causal effect of the QBO on the extratropical winter stratosphere. That this coupling interacts with other sources of interannual variability has become increasingly apparent as the observational record has lengthened since Holton and Tan (1980). We have reviewed the literature on the Holton–Tan effect with several aims in mind: compiling a useful source of references; determining the consistencies between these studies (i.e. which results seem robust and which do not); presenting updated versions of some key diagnostics; and better enabling detailed comparisons between future work and the large number (and variety) of observational and modelling results that already exist. Particularly with respect to this last aim, it is hoped that the review may contribute toward clarifying the mechanisms responsible for QBO–vortex coupling.

Acknowledgements

This work has been supported by the Canadian Foundation for Climate and Atmospheric Sciences (CFCAS) and the Natural Sciences and Engineering Research Council (NSERC) of Canada. We acknowledge the European Centre for Medium-range Weather Forecasts (ECMWF) for our use of ERA-40 and ERA-Interim reanalysis data. Daily 10.7 cm solar flux data were obtained from the website of the National Research Council (NRC) of Canada Dominion Radio Astrophysical Observatory in Penticton, British Columbia. The authors would like to thank Peter Watson and two anonymous reviewers for insightful and constructive comments on the manuscript.

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