Model studies of the interannual variability of the northern-hemisphere stratospheric winter circulation: The role of the quasi-biennial oscillation

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SUMMARY

A series of experiments are described that examine the sensitivity of the northern-hemisphere winter evolution to the equatorial quasi-biennial oscillation (QBO). The prime tool for the experiments is a stratosphere–mesosphere model. The model is integrated over many years with the modelled equatorial winds relaxed towards observed values in order to simulate a realistic QBO. In experiment A the equatorial winds are relaxed towards Singapore radiosonde observations in the height region 16–32 km. In contrast to previous modelling studies, the Holton–Tan relationship (warm/cold winters associated with easterly/westerly QBO winds in the lower stratosphere) is absent. However, in a second experiment (run B) in which the equatorial winds are relaxed towards rocketsonde data over the extended height range 16–58 km, a realistic Holton–Tan relationship is reproduced. A series of further studies are described that explore in more detail the sensitivity to various equatorial height regions and to the bottom-boundary forcing. The experiments suggest that the evolution of the northern-hemisphere winter circulation is sensitive to equatorial winds throughout the whole depth of the stratosphere and not just to the lower-stratospheric wind direction as previously assumed.

KEYWORDS: Interannual variability Northern-hemisphere winter Quasi-biennial oscillation Semi-annual oscillation Stratosphere Stratospheric warming

1. INTRODUCTION

A quasi-biennial oscillation (QBO) influence on the extratropical middle atmosphere has been observed for many years. It was highlighted by Holton and Tan (1980, 1982) and has been followed up by many more studies (e.g. Baldwin and Dunkerton 1991, 1998; Naito and Hirota 1997; Baldwin et al. 2001). In general, observations suggest that the stratospheric polar vortex is weaker, warmer and more disturbed by planetary waves when the lower-stratospheric equatorial winds are in the easterly phase of the QBO than when they are in the westerly phase. Baldwin and Dunkerton (1998) found that the difference between east and west QBO phase composites in zonally averaged zonal wind reached 14 m s⁻¹ at high latitudes in the northern hemisphere (NH) in January, with stronger westerlies when the equatorial QBO was in a westerly phase. They found that the differences were greatest in NH winter when the composites were compiled on the basis of the direction of equatorial wind in the lower stratosphere (near 40 hPa, ~22 km).

Modelling studies (e.g. Dameris and Ebel 1990; O’Sullivan and Salby 1990; Balachandran et al. 1991; Holton and Austin 1991; O’Sullivan and Young 1992; O’Sullivan and Dunkerton 1994; Balachandran and Rind 1995; Hamilton 1995, 1998; Niwano and Takahashi 1998) have supported a mechanism for planetary-wave–mean-flow interaction involving the influence of equatorial QBO winds in the lower stratosphere on the propagation of planetary waves. More specifically, the phase of the equatorial QBO in the lower stratosphere determines the position of the zero wind line in the subtropics which then acts as a critical line for planetary waves as they propagate through this lower part of the atmosphere. In easterly years, with the critical line in the northern subtropics, the planetary waves are more confined to the higher northern latitudes than in the westerly phase. This leads to more effective poleward transfer of heat and momentum into
the polar-vortex region, resulting in a weaker vortex. Current theory is thus based on the premise that it is the background state of the lower equatorial stratosphere, as the waves propagate vertically into the stratosphere, that influences events at higher latitudes, such as the nature and frequency of stratospheric sudden warmings. For example, O'Sullivan and Young (1992) used a stratosphere–mesosphere model to test the sensitivity of the NH polar vortex to the equatorial winds over various height ranges. They concluded that there was particular sensitivity to the tropical winds in the lower equatorial stratosphere.

However, it is also well known that in reality this relationship of warm-disturbed/cold-undisturbed winters when the lower-stratospheric equatorial winds are in an easterly/westerly QBO phase is by no means the rule. For example, while there is a tendency for major mid-winter warmings to occur preferentially in the east phase, there are examples of major warmings in westerly phase years, e.g. 1967/68, 1969/70, 1978/79, 1980/81, 1987/88, 1988/89, and 1990/91. Indeed, the WMO definition of a major warming is rather arbitrary and there are many cases of westerly QBO phase years in which a significant mid-winter warming occurs but, nevertheless, does not technically qualify as a major warming. Hamilton (1998), on the basis of a 48-year general-circulation model (GCM) simulation, concluded that there is nothing approaching a one-to-one relationship between the tropical QBO phase and the state of the winter-mean NH polar vortex or the occurrence of sudden warmings. However, while polar warmings occurred in his model with almost any tropical-mean wind profile, warmings were significantly more probable when the lower-stratospheric equatorial wind was easterly.

In a data study using equatorial wind observations from rocketsondes and stratospheric analyses of polar winter temperatures, Gray et al. (2001) found a greater sensitivity of polar temperatures to equatorial wind distributions in the region of the stratopause (≈50 km) than in the lower-stratospheric region (≈25 km). In this paper we use a stratosphere–mesosphere model (SMM) to investigate this unexpected result. We simulate the QBO by relaxing the modelled zonally averaged equatorial zonal winds (\(\bar{u}_{eq}\)) towards observations. Two primary experiments are described. In the first, the modelled equatorial winds are relaxed towards radiosonde observations that extend to 32 km (≈10 hPa); above this level, they are allowed to evolve freely. In the second experiment, the modelled equatorial winds are relaxed towards rocketsonde data that extend to 58 km (≈0.5 hPa). Using these two experiments we examine the sensitivity of the NH winter evolution to the height range of the equatorial QBO winds. In particular we are able to examine whether the winds above 32 km have any influence on the high-latitude winter evolution. This is the region of the equatorial atmosphere dominated by the semi-annual oscillation but in which there is nevertheless a QBO influence (Kennaugh et al. 1997). The paper is arranged as follows. The model is described in detail in section 2. Results from the two runs are described in sections 3 and 4, including an analysis of the model variability, and results of various composite and correlation studies. In section 5 a series of sensitivity studies is described. The main conclusions of the study are summarized in section 6.

2. The model

The model employed in this study is the UK Met Office (UKMO) SMM. It is a global three-dimensional primitive-equation model of the middle atmosphere (Fisher 1987) with horizontal resolution of 5 degrees latitude, 5 degrees longitude and 32 vertical levels equally spaced in log-pressure giving approximately a 2 km vertical resolution. It is a mechanistic model with temperature and horizontal winds as prognostic variables and a lower boundary located near the tropopause at which 100 hPa (approximately
16 km) geopotential-height fields are specified as the bottom-boundary condition (see below).

Radiative heating and cooling rates are computed in the model using the MIDRAD scheme (Shine 1987; Shine and Rickaby 1988) with a prescribed annual-mean, global-mean carbon dioxide amount and a zonal-mean, height-resolved, monthly mean ozone climatology. In this respect, the model is more sophisticated than the simple scheme employed in the model of O'Sullivan and Young (1992) and O'Sullivan and Dunkerton (1994). A leapfrog integration scheme is used with fourth-order accuracy in the horizontal and second-order accuracy in the vertical. A time step of 240 seconds is employed. A simple (Rayleigh friction) relaxation scheme is used to simulate the effect of gravity-wave breaking. The zonally averaged zonal winds are relaxed towards zero on a time-scale varying from about 100 days in the lower stratosphere to 2 days in the mesosphere.

The model has previously been used extensively for dynamical studies of the stratosphere, e.g. Butchart et al. (1982), O'Neill and Pope (1988), Fairlie et al. (1990), Rosier and Lawrence (1999), Gray (2000), and Scaife and James (2001). These studies have confirmed that the model is capable of producing a realistic annual evolution of the middle atmosphere, including simulations of sudden stratospheric warmings.

In all experiments (see Table 1) the model was initialized with a realistic November wind and temperature distribution using UKMO assimilated data (Swinbank and O'Neill 1994). In the region of the upper mesosphere for which assimilated data were not available, data from the CIRA (COSPAR (Committee on Space Research) International Reference Atmosphere) dataset (Fleming et al. 1986) were merged with the assimilated dataset. The bottom-boundary forcing employed geopotential-height fields from the UKMO assimilated dataset. The model runs were forced with the UKMO daily geopotential-height fields from two different years: 1979/80 and 1991/92. Data for the period 7 November to 6 November of the following year were used. To continue into the second and subsequent years of the integrations, the same annual cycle of geopotential-height data was used, with a smoothed transition back to the first year data so that there were no sudden steps in the forcing data to disturb the model. This repeated cycling of the lower-boundary forcing data was carried out for each year of the model integration so that the interannual variability of the lower-boundary forcing was minimized. (Although the forcing of geopotential height in this way does not completely constrain the Eliassen–Palm (E–P) fluxes to be identical in each year, we found no significant correlation of the NH winter polar temperatures with the lower-boundary E–P fluxes.)

The QBO was included in the model by relaxing $\overline{u}_{eq}$ towards monthly averaged equatorial wind observations. The relaxation time-scale was 5 days at the equator and this time-scale was linearly increased with latitude (to 20 days at 17.5 degrees latitude) to produce a realistic QBO zonal wind. Similarly, above the upper limit of the forcing the relaxation time-scale was increased in order to achieve a smooth transition from the forced zonal winds below to the modelled winds above. The latitudinal width of the forcing by this method was very similar to that of Hamilton (1998). A sensitivity experiment, in which the width of this forcing was doubled, showed little sensitivity of the results to the width of this forcing.

Two observational data sources were used in the equatorial-wind relaxation scheme. In model runs A and D, radiosonde data extending to 32 km were employed (Naujokat 1986). In model runs B, C, and E (see Table 1) rocketsonde observations extending to 58 km from Ascension Island (8°S) and Kwajalein (8°N) were employed (Dunkerton and Delisi 1997; Gray et al. 2001). The rocketsonde data from Ascension Island
and Kwajalein were combined to produce a single, unbroken time series of ‘proxy-equatorial’ winds in the following manner. Data from the two stations were first de-seasonalized by calculating and subtracting the climatological monthly mean at each station. Gaps in data of up to 3 months were filled in by linear interpolation. For those months and heights where data were available from both stations, the average of the two stations was used. Where data from only one station was available, those data were used. In this way, an unbroken time series of de-seasonalized ‘proxy equatorial’ wind was established. Using this technique a time series spanning the period autumn 1964–spring 1987 was achieved. Although data were actually available until 1990, they became increasingly intermittent during this period and required significant interpolation over missing data, so the last three years were not used. In order to reintroduce the annual cycle, the climatological values for each month at each of the stations were averaged together. This value was then added to each corresponding month of the time series.

3. Model run A: QBO relaxation below 32 km

(a) The model run

In model run A, a 33-year integration was performed in which the modelled equatorial winds were relaxed towards observed winds from the Singapore radiosonde data (Naujokat 1986) in the region 16–32 km for the period autumn 1964–spring 1997 (i.e. 33 winters). Above 32 km, the modelled equatorial winds were unconstrained and were, therefore, weakly easterly throughout the year because there was no parametrization of the effects of wave forcing, e.g. from gravity or Kelvin waves included in the model. Annually repeating geopotential-height fields from 1991/92 were used as the bottom-boundary forcing of the model (see section 2). In the following sections, the years of the model run are referred to as 1964–97. Note that these years, therefore, refer only to the phase of the QBO and not to the bottom-boundary planetary-wave forcing.

(b) Model variability

As an indication of the variability of the NH winters in the 33-year model run A, Fig. 1 shows the time series of area-weighted temperature north of 62.5°N at 32 km (approximately 10 hPa) for the months November through March in each year of the model integration. There is little interannual variation early in the winter until mid December. However, in mid and late winter, there is significant variability, the amplitude of which compares favourably with observations (e.g. Manzini and Bengtsson 1996). The majority of the winters follow the same broad pattern with a series of minor warmings that increase in strength during the winter. Despite the broad similarities in the evolution of the winters there is, nonetheless, significant variability from year to year from around mid December onwards. Some winters display a minor warming in January while others remain substantially colder and less disturbed than average.
throughout January. Similarly, in February there is substantial variability. In some years the high-latitude zonal winds in February decrease, in association with the warming, but still remain greater than 20 m s$^{-1}$ (hence only a minor warming), in other years the winds reduce and became easterly, but only just qualify as a major warming, and in the remaining years there is a much more substantial warming with winds decreasing to at least $-20$ m s$^{-1}$.

(c) QBO composites

In order to examine the influence of the QBO on the evolution of all the winters, QBO ‘composites’ were derived. The modelled winters were separated into an easterly composite and a westerly composite based on the amplitude and sign of $\overline{u}_{eq}$ at 22 km (40 hPa) in December of each winter. 40 hPa was chosen as the criteria to aid comparison with the National Centers for Environmental Prediction (NCEP) data study of Baldwin and Dunkerton (1998). ‘Transition’ years, in which the QBO phase was changing from easterly to westerly or vice versa during the course of the winter, were excluded. This was done using a simple test that excluded winters where $\overline{u}_{eq}$ at 22 km in December was within the range ±5 m s$^{-1}$. Different threshold values for the definition of the transition phase (e.g. ±10 m s$^{-1}$) were tested to check whether the definition of transition biased the results in any way, but all results were found to be robust.

Figure 2(a) shows the time evolution of zonally averaged, area-weighted temperature north of 62.5°N at 32 km for the two QBO composites. In December and February, there is no obvious difference in the pattern of the evolution between the two QBO phases. Both distributions display substantial variability. For example, both composites contain years with significant minor warmings in January followed by major warmings in February and, equally, both contain years with relatively undisturbed flow. Of the 23 years that exhibited major or near-major warmings in February, 11 were in easterly
phase years and 12 were in westerly phase years. Of the years with no major warming in February, 4 were in the west phase, 3 in the east phase and 3 were classified as transition years. There is, therefore, no apparent bias towards one phase or the other. In January, on the other hand, there is a marked separation between the average behaviour of the two composites. However, the differences are not consistent with a Holton–Tan relationship (hereafter referred to as the H–T relationship), since the westerly composite is, on average, warmer than the easterly composite.

In a study of NCEP data (Kalnay et al. 1996) for the period 1978–96, Baldwin and Dunkerton (1998) derived composites of the monthly averaged, zonally averaged zonal winds for each phase of the QBO and showed the west-minus-east composite anomalies for each month. In the northern mid and high latitudes the NCEP analysis
shows a positive anomaly throughout the winter, with a maximum anomaly of 14 m s\(^{-1}\) in January. This positive anomaly, with stronger high-latitude westerlies on average in the west QBO phase than in the east QBO phase provides evidence of colder, less disturbed winters in the west phase than in the east phase, in agreement with the H–T theory.

For comparison, we have derived identical diagnostics from model run A. Figure 3 shows the west-minus-east QBO composite differences in zonally averaged zonal winds for each month from November to February. The QBO signal in the lower equatorial stratosphere is clearly evident in all the fields. There is a positive (westerly) anomaly of around 20 m s\(^{-1}\) centred at 22 km over the equator in all months, with a negative (easterly) anomaly of around \(-20\) m s\(^{-1}\) just above it, centred at 32 km (10 hPa). At mid and high latitudes in November and December there is a dipole pattern with negative (easterly) anomalies at mid latitudes and positive (westerly) anomalies at polar
Figure 3. Model run A (see text): west-phase-minus-east-phase composite of monthly averaged zonally averaged zonal wind (m s$^{-1}$) for November to February. Contour values are ±2, ±4, ±6, ±8, ±10, ±12, ±14, ±16, ±18, ±20, ±25, and ±30. Solid contours denote positive (westerly) anomalies, dotted contours denote negative (easterly) anomalies, the dashed line is the zero contour. Shading indicates values are significant at the 99% confidence level using a Student’s t-test.

latitudes. This pattern agrees well with the NCEP anomalies, although the polar anomaly in December is rather weaker and further poleward than the NCEP data. In January, however, the model has a strong negative (easterly) anomaly at mid and high latitudes which is opposite in sign to the NCEP data results. In February, there is very little difference at high latitudes between the two QBO composites.

The amplitudes and distributions of the composite anomalies in Fig. 3 therefore agree relatively well with NCEP analyses only in the early winter period (November and December) but in mid to late winter (January and February), when the significant warming events occur, the results do not compare well. However, we note here the lack of shading at mid and high latitudes in Fig. 3, which indicates that even the relatively large differences in January are not statistically significant.

(d) Correlations

To determine the extent of the correlation between the evolution of the NH winter and the phase of the equatorial QBO in the lower stratosphere, Fig. 4 shows the rank correlation coefficient of $u_{eq}$ at 22 km with $\bar{u}$ at all other latitudes and heights for each month from November to February. All correlations were carried out using monthly averaged data. As in the case of the wind composite anomalies discussed earlier, the correlations in early winter (November, December) are in good agreement with predictions of behaviour using the H–T relationship. There is a positive correlation...
at high latitudes reaching 0.6 and 0.5 in November and December, respectively. This confirms that in early winter the polar-vortex westerlies are stronger when $\tilde{u}_{eq}$ in the lower stratosphere is westerly. However, in mid to late winter (January, February), during the period of the significant warming events, there is no discernible correlation with the phase of the QBO in the equatorial lower stratosphere.

Figure 5(a) shows the correlation over the 33 winters of the zonally averaged, area-weighted, January/February (JF) averaged polar temperature north of 62.5°N at 24 km in each modelled year with $\tilde{u}_{eq}$ at all heights in each of the previous 13 months (i.e. from January of the previous year to February of the same year). Gray et al. (2001) carried out a corresponding correlation analysis of observed polar temperatures from the Berlin stratospheric analyses and equatorial winds from radiosonde and rocketsonde data. The results of the rocketsonde analysis are reproduced in Fig. 5(b). This observational analysis shows a negative correlation in the lower stratosphere, with the region of maximum negative correlation descending through the atmosphere, corresponding to the descent of the equatorial wind QBO. This pattern is, therefore, consistent with the H–T mechanism, with easterly (negative) wind anomalies associated with warm (positive) polar temperature anomalies. Model run A (Fig. 5(a)), on the other hand, shows the opposite situation, with a small positive correlation between 20 and 40 km. However, this correlation is extremely small and is not statistically significant.
Figure 5. Height-time distribution of correlation coefficients between area-weighted, January-to-February average polar temperatures north of 62.5°N at 24 km and monthly averaged $\tau_{eq}$ in the previous 13 months (i.e. from January of the previous year to February of the same year). Dotted contours denote negative values, the dashed contour is zero. Light and dark shading indicate values are significant at the 95% and 99% confidence levels, respectively. See appendix for details of the significance test applied. (a) Model run A, (b) observations using rocketsonde equatorial winds and polar temperatures from the Berlin analyses, (c) model run B, (d) model run C, (e) model run D, (f) model run E. See text for further details.
Figure 5. Continued.
Figure 6. Model run B (see text): west-phase-minus-east-phase composite of monthly averaged zonally averaged zonal wind (m s\(^{-1}\)) for November to February. Contour values are ±2, ±4, ±6, ±8, ±10, ±12, ±16, ±18, ±20, ±25, and ±30. Solid contours denote positive (westerly) anomalies, dotted contours denote negative (easterly) anomalies, dashed line is the zero contour. Shading indicates values are significant at the 99% confidence level using a Student's t-test.

4. Model run B: QBO relaxation up to 58 km

(a) The model run

In model run B, a 23-year integration was performed in which the modelled equatorial winds were relaxed in the region 16–58 km towards observed winds from the combined Kwajalein and Ascension Island rocketsonde dataset (see section 2). The upper limit of the height range over which the modelled winds are relaxed towards observations has thus been extended from 32 km in model run A to 58 km in model run B. This enables an examination of the influence of the upper-stratospheric equatorial wind on the evolution of the winter vortex. Rocketsonde data for the period autumn 1964–spring 1987 (i.e. 23 winters) were employed. The data record is, therefore, significantly shorter than for the radiosonde data and hence the statistical analysis of run B is inferior to that of run A. We note here, however, that all diagnostics of model run A have been repeated using only the 23 years to check that results were not biased by the length of the integration. Similarly, a detailed comparison of the radiosonde and rocketsonde data in the region 16–32 km was carried out to ensure that the different data sources for equatorial winds in this height region did not bias the results. The two sets of observations were virtually identical and this possibility was, therefore, ruled out.

In the following sections, the years of the model run are referred to as 1964–87. Note that these years refer only to the phase of the QBO and do not refer to the planetary-wave forcing at the lower boundary, since the 1991/92 annual cycle of geopotential height was
used repeatedly throughout the runs as the bottom-boundary forcing field, as in run A (see section 2).

(b) QBO composites

QBO composites were derived for model run B based on the sign and amplitude of \( \bar{u}_{eq} \) at 22 km in December as before (see section 3(c)). Figure 2(b) shows the average time evolution (± one standard deviation) of the zonally averaged, area-weighted temperature north of 62.5°N at 32 km for each of the composites and can be compared directly with the corresponding plot from run A (Fig. 2(a)). The broad evolution of the winters is similar to model run A, with a series of warmings and increasing variability as the winter proceeds. In contrast to run A, however, there is a separation between the composite difference in December through February which is now consistent with the H−T relationship, namely the QBO east-phase composite is consistently warmer than the westerly phase composite. For example, at the peak of the warming in January the difference between the composite-average temperatures at 32 km in Fig. 2(b) is nearly 9 K and at 40 km it is nearly 20 K (not shown). This difference is also evident in Fig. 6 which shows the latitude−height distribution of the west-minus-east QBO composite difference in zonally averaged zonal winds for each month November−February. This can be compared directly with the results of model run A shown in Fig. 3. There is a consistently stronger vortex (i.e. more westerly) during the west phase of the lower-stratospheric QBO than during the east QBO phase. The composite difference
Figure 8. Model run B (see text): latitude–height distribution of correlation coefficients over 23 years between $u_{eq}$ at 22 km in the previous September and $u$ at all other latitudes and heights for November to February. Contour interval is 0.2. Dotted contours denote negative values. Only contours with absolute values of 0.3 and greater have been plotted. Shading indicates values are significant at the 99% confidence level. See appendix for details of the significance testing.

is 14 m s$^{-1}$ in both December and February and is significant at the 99% level. A negative anomaly at subtropical latitudes also strengthens from December to February and reaches 8 m s$^{-1}$ at $\sim$30$^\circ$N and 40 km in February. This latter anomaly is also significant at the 99% level.

(c) Correlations

Figure 7 shows the correlation coefficient over 23 years of $u_{eq}$ at 22 km with $u$ at all other latitudes and heights in each month November to February. It can be compared directly with the equivalent plot from run A shown in Fig. 4. Interestingly, the correlations in January and February from run B do not show much improvement over those from run A even though the composites (Fig. 6) were markedly improved (but see below). Figure 5(c) shows the correlation over 23 years of the zonally averaged, area-weighted, JF-averaged temperatures north of 62.5$^\circ$N at 24 km with $u_{eq}$ at all heights in the previous 13 months. This shows a marked improvement compared with the corresponding diagnostic from run A (Fig. 5(a)). There is a region of negative correlation centred at around 32–34 km in January of the previous year that gradually descends with time, with a maximum correlation of $-0.5$ at the 95% level at 24 km in September/October. Similarly, a descending region of high positive correlation is present at higher levels, descending from around 40–45 km in January of the previous year to around 32–34 km in January/February of the same year. The correlation pattern compares well with the observational study (Fig. 5(b)). Note, however, that the strong
correlation in Fig. 5(b) at ~52 km in September is missing from the model analysis. This may be due to the lack of a realistic semi-annual oscillation in the modelled mesosphere since the equatorial relaxation towards rocketsonde data extends only to 58 km. Furthermore, the model does not include a representation of solar-cycle effects, which may influence this height region.

The reason for the lack of improvement of the correlations in individual months (Fig. 7) is now evident from Fig. 5(c). The correlation analysis shown in Fig. 7 was carried out using $\overline{u}_{eq}$ from the same month, i.e. November to February. Figure 5(c) shows that the correlation of polar temperatures with equatorial winds at around 22 km is declining from November through February. The peak amplitude of the correlation is, in fact, in September/October. Repeating the correlation analysis of Fig. 7 but using $\overline{u}_{eq}$ at 22 km in the previous September results in a larger, more significant correlation (see Fig. 8).

5. Sensitivity studies

From the results shown in the previous sections it is evident that the H–T relationship is present in the model when the equatorial winds are relaxed towards observations throughout the height region 16–58 km (Fig. 5(c)) but is absent when they are relaxed towards the more limited range of 16–32 km (Fig. 5(a)), the part of the atmosphere traditionally regarded as the region of primary influence. In order to further test the sensitivity of the H–T relationship to the height region over which the equatorial winds are relaxed, three additional experiments were carried out. In model run C, a 23-year model integration (1964–87) was performed in which the equatorial winds were relaxed towards the rocketsonde data over the height 16–40 km. This model run, therefore, extends the upper limit of the relaxation region of model run A by a further 8 km. This run is probably closest to the experiment of Hamilton (1998), whose upper limit of significant relaxation was approximately 38 km. Figure 5(d) shows the correlation coefficients over the 23 years between the JF-averaged, area-weighted temperatures north of 62.5°N at 24 km and $\overline{u}_{eq}$ at all heights in the previous 13 months. In contrast to model run A (Fig. 5(a)), model run C shows an H–T relationship with a descending negative correlation between the polar temperatures and equatorial winds at 20–30 km and a descending positive correlation overlying it in the region 30–40 km. However, this intermediate run fails to capture the observed structure of the correlation above 40 km (not surprisingly) and the correlation below 30 km is smaller and less significant than in run B (Fig. 5(c)).

In two further 23-year (1964–87) experiments the equatorial relaxation towards the rocketsonde data was included only between the height regions 40–58 km and 30–58 km, in order to test whether these upper regions could be solely responsible for the strong H–T relationship displayed by model run B. Neither displayed the strong correlation pattern of run B, suggesting that the whole depth of the equatorial atmosphere in the region 16–58 km is influential in producing the H–T relationship shown in Fig. 5(c).

Finally, two sensitivity tests, runs D and E were carried out to check whether the results obtained were specific only to experiments using the 1991/92 tropospheric-forcing boundary conditions. The two primary experiments (runs A and B, respectively) were repeated except that 100 hPa geopotential heights from 1979/80, the first available year of the UKMO TOVS (TIROS (Television Infra-Red Observation Satellite) Operational Vertical Sounder) analyses (Bailey et al. 1993), were used as the bottom-boundary forcing (see Table 1). The general evolution of the area-weighted temperature north of
62.5°N at 32 km (Figs. 2(c) and (d)) for the east and west QBO composites (derived as in section 3(c)) in these two experiments is different in nature to runs A and B, reflecting the different amplitude and timing of the tropospheric wave forcing. However, the results are generally consistent with those from runs A and B. The polar-averaged temperatures in run D (Fig. 2(c)) show the westerly phase years to be slightly warmer on average than the easterly phase years (as in run A), in contrast to the H–T relationship. In run E however (Fig. 2(d)), this relationship is reversed in December and January, with warmer winters in the easterly QBO phase on average, in agreement with the H–T relationship. In February, inclusion of equatorial forcing in the upper stratosphere appears to have had little impact and in mid February the westerly phase years are on average slightly warmer than the easterly phase years. This may be a result of the stronger warmings in the easterly phase in the January: weaker westerly winds following a warming can inhibit further wave forcing in the following days/weeks, resulting in a colder than average period immediately following the warming event. This effect is also evident in Fig. 2(b) from run B: in January and February the easterly QBO phase winters are, on average, warmer than the westerly phase winters but in March they are significantly colder.

Because of the lack of improvement in February of run E compared with run D, correlations of averaged polar temperatures with equatorial winds in the previous 13 months using both January and February (averaged together) show no significant correlations in run E. However, if only the January polar temperatures are used, the pattern of correlations from runs D and E are similar to those from runs A and B. Run D (Fig. 5(e)) shows a positive correlation in the lower stratosphere while run E (Fig. 5(f)) shows a negative correlation that is in closer agreement with the observations in Fig. 5(b). Note also that the region of positive correlation in run E extends higher in the atmosphere than in run B, with maximum values of 0.5 in the region 50–60 km in June. Hence, while the overall patterns of the correlations are similar in runs B and E there are, nevertheless, significant variations, perhaps unsurprisingly, given the different nature of the tropospheric forcing in the two runs. Although outside the scope of this current report, an extension of this comparison to include tropospheric forcing from all years for which tropospheric forcing data are available would be valuable.

6. Conclusions

Two primary-model experiments have been carried out using an SMM that extends from 16 km to 80 km. In model run A (section 3) the modelled zonally averaged equatorial winds were relaxed towards radiosonde observations over the height region 16–32 km. In model run B (section 4) they were relaxed towards rocketsonde observations over the height region 16–58 km. The prime aim of the study was to examine the influence of the equatorial wind QBO on the evolution of the NH winter evolution. In particular, each model integration was tested for evidence of an H–T relationship between equatorial winds and lower-stratospheric polar temperatures. An H–T relationship was determined to be present if the polar vortex was statistically colder and more westerly during a west-phase equatorial QBO and warmer and more easterly during an east-phase equatorial QBO.

Both model integrations were run for many years (model run A: 33 years, model run B: 23 years) in order that the results could be tested for statistical significance. In this respect, the experiments contrast with many previous model studies of the QBO, in which only a single winter in each QBO phase was considered. In order to minimize the
interannual variability associated with variations in tropospheric planetary-wave forcing, the model was forced at the lower boundary using a cycle of observed geopotential heights from the same year. These were recycled and used repeatedly for each year of the model integration.

The prime diagnostics employed to analyse the model results were composites and correlation studies. The analysis of model run A (equatorial QBO relaxation over 16–32 km) showed some evidence of an H–T relationship in early winter (November, December) but no relationship during late winter (January, February) when the major warming events occurred. This contrasts with results from earlier studies, e.g. O'Sullivan and Young (1992). On the other hand, model run B (equatorial QBO relaxation over 16–58 km) showed a strong H–T relationship and compared well with a similar analysis using observational data (Fig. 5(b), see also Gray et al. (2001)).

An intermediate model run C (equatorial QBO relaxation over 16–40 km) showed a weak H–T relationship that, not surprisingly, did not capture the observed structure of the correlations above 40 km and had a weaker, less significant correlation below 30 km. Hence, as the upper limit of the equatorial QBO wind relaxation was increased from 32 km (run A), to 40 km (run C) through to 58 km (run B), the H–T relationship displayed by the model was improved and strengthened.

The results summarized above show that the presence of an H–T relationship in late winter in this model requires realistic equatorial winds over an extended height region that includes the upper stratosphere as well as the lower stratosphere. The study, therefore, suggests that equatorial winds in the upper equatorial stratosphere (i.e. above 35 km) are influential in determining the state of the polar vortex in late winter. This result is consistent with conclusions of Gray et al. (2001) who found a significant correlation between late-winter NH polar temperatures from the Berlin stratospheric analyses and upper-stratospheric equatorial wind observations from rocketsondes.

In a sensitivity study, two model integrations were carried out in which the equatorial winds were relaxed over the height regions 40–58 km and 30–58 km. These were to investigate whether an H–T relationship would arise if only the upper-stratospheric equatorial winds were imposed from observations. Although it seems unlikely that the H–T relationship would arise solely as a result of relaxing $\overline{u}_{eq}$ towards observed winds in the upper stratosphere, it is important to rule out this possibility. Neither of these two model runs showed any indication of an H–T relationship, thus confirming that the upper equatorial winds cannot, by themselves, reproduce the H–T relationship. This reinforces the conclusion of the study, that the equatorial influence on the evolution of the NH winter polar vortex involves the whole depth of the equatorial stratosphere. Sensitivity experiments, identical to runs A and B except that tropospheric forcing from a different year was used, showed that the above results were not specific to one particular year.

While the results of this model study are unambiguous, the study is nevertheless idealistic and care is required in applying the results to the real atmosphere. The model was constrained in a number of ways in order to simplify the interpretation, and the lack of several important feedback mechanisms must be recognized. The model does not simulate the troposphere and there may be feedback between the equatorial and polar stratosphere through this route. The imposed tropospheric forcing was taken from two years only (1991/92 and 1979/80). Although the sensitivity test carried out using tropospheric forcing from 1979/80 supports the conclusions of the main 1991/92 study, more extensive tests with many different forcings are required to fully test the robustness of the result. It is unlikely that all years would show a similar degree of sensitivity to the upper equatorial stratosphere. Also, the model does not include a solar cycle and its interaction with the QBO is, thus, not represented. Finally, the model does not
include an interaction between the imposed QBO dynamics and the ozone (and hence heating) distributions. Any of these missing feedbacks may influence the relevance of these model results to the real atmosphere.

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APPENDIX

The significance of each point on the correlation plots was tested by calculating the value of the statistic \( t = r \sqrt{(N_c - 2)/(1 - r^2)} \), where \( r \) is the correlation between the two time series and \( N_c \) is the number of independent data points in the time series. In the null case of no correlation, this statistic is distributed like the Student’s t-distribution with \( N_c - 2 \) degrees of freedom (Press et al. 1992). In order to compensate for the autocorrelation of the wind data, the effective sample size \( N_c = N \), where \( N \) is the actual number of data points in the time series, unless \( \rho_1 > 0 \) where \( \rho_1 \) is the lag-one autocorrelation coefficient, in which case \( N_c = N(({1 - \rho_1})/(1 + \rho_1)) \) (Wilks 1995).

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