The Climatology of the Middle Atmosphere in a Vertically Extended Version of the Met Office’s Climate Model. Part II: Variability

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ABSTRACT

Stratospheric variability is examined in a vertically extended version of the Met Office global climate model. Equatorial variability includes the simulation of an internally generated quasi-biennial oscillation (QBO) and semiannual oscillation (SAO). Polar variability includes an examination of the frequency of sudden stratospheric warmings (SSW) and annular mode variability. Results from two different horizontal resolutions are also compared. Changes in gravity wave filtering at the higher resolution result in a slightly longer QBO that extends deeper into the lower stratosphere. At the higher resolution there is also a reduction in the occurrence rate of sudden stratospheric warmings, in better agreement with observations. This is linked with reduced levels of resolved waves entering the high-latitude stratosphere. Covariability of the tropical and extratropical stratosphere is seen, linking the phase of the QBO with disturbed NH winters, although this linkage is sporadic, in agreement with observations. Finally, tropospheric persistence time scales and seasonal variability for the northern and southern annular modes are significantly improved at the higher resolution, consistent with findings from other studies.

1. Introduction

The ability to reproduce the well-observed structure and variability of the atmosphere is an expectation of modern-day global climate models (Garcia et al. 2007; Lott et al. 2005; Manzini et al. 2006; McLandress and Shepherd 2009; Scinocca et al. 2008). For stratospheric studies, this necessarily includes adequate representation of tropical variability, including such phenomena as the stratospheric quasi-biennial oscillation (QBO) and the semiannual oscillation (SAO) in the stratosphere (Gray 2010) and mesosphere (Richter and Garcia 2006). Outside the tropical stratosphere, the annual cycle of temperature and zonal wind is controlled by seasonal changes in solar irradiance and radiatively significant gases such as ozone, together with the indirect effects of the propagation and dissipation of atmospheric waves on all scales. The ability of general circulation models (GCMs) to reproduce the effects of these is important in the understanding of seasonal and intraseasonal variability and the redistribution of trace gases, as well as reproducing a basic climatology (e.g., the strength of the stratospheric jets).

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Knowledge and understanding of future trends and variability of trace gases such as ozone is especially relevant to understanding and predicting future climate change (Waugh et al. 2009; Morgenstern et al. 2008). This is not only dependent on knowing the constraints on future emissions of greenhouse gases or ozone-destroying substances but is also underpinned by the ability to synthesize this with our knowledge of the physics and dynamics of the atmosphere. Uncertainties in our understanding—or more generally, our representation of physical and dynamical processes in climate models—place important constraints on predicted future climate change. One such uncertainty includes the representation of spatial scales in models.

Studies of the effects following changes in spatial resolution are well documented in the literature (Hamilton 2008, and references therein). In this study, we document the stratospheric variability within the Met Office “high-top” Unified Model (MetUM) and also examine changes in variability resulting from doubling the horizontal resolution. This paper follows on from a companion study describing the climatology (Hardiman et al. 2010, hereafter Part I). We first describe the model’s representation of tropical variability, examining the QBO, SAO, and their interaction with the annual cycle. Extratropical variability is then examined during winter, with emphasis on the occurrence of sudden stratospheric warmings (SSWs) and their relation to the phase of the QBO. We conclude by examining large-scale variability, specifically by calculating persistence time scales and seasonality of the annular modes.

2. Methods

The vertically extended MetUM has 60 vertical levels from the surface to around 84 km, as described in Part I. Unlike previous versions of the vertically extended Unified Model (e.g., Butchart and Austin 1998), this version is height based and utilizes a significantly different numerical core (Davies et al. 2005). The radiation scheme used is that of Edwards and Slingo (1996) with modifications to shortwave spectral irradiance and ozone absorption parameters (Zhong et al. 2008). The effects of subgrid-scale nonorographic gravity waves are parameterized using the Ultra Simple Spectral Parameterization (Scaife et al. 2002; Warner and McIntyre 2001) and a flow blocking scheme is used in the representation of topographic gravity wave effects (Webster et al. 2003). These are employed to improve tropical stratospheric variability (specifically the QBO) and also the extratropical momentum budget.

The study methodology is described in detail by Part I; however, a brief summary is given below. Two sets of ensembles were run to study the differences in model performance resulting from changes in horizontal resolution. A four-member ensemble was run for the period 1975–2000 at $2.5^\circ \times 3.75^\circ$ (N48) latitude–longitude resolution and a two-member ensemble was run from 1980–2000 at $1.25^\circ \times 1.875^\circ$ (N96) resolution. A new monthly-averaged ozone climatology (Dall’Amico et al. 2010) was employed together with monthly-averaged time-varying sea surface temperature, sea ice fields (Rayner et al. 2003), CO$_2$, and methane. For most of the analysis, comparisons are made with data from 1979 to 2002 from the 40-yr European Centre for Medium-Range Weather Forecasts (ECMWF) Re-Analysis (ERA-40; Uppala et al. 2005). Significance levels were evaluated using Student’s $t$ tests, unless specified otherwise.

3. Results

a. Tropical variability

1) QBO

Figure 1 shows the height–time development of monthly-mean, zonal-mean zonal wind between $5^\circ$S and $5^\circ$N from a single member of each of the N48 and N96 ensembles. These runs illustrate typical ensemble behavior and exhibit a spontaneously generated QBO in the stratosphere, with ensemble-averaged periods of 34.0 ± 3.5 months and 36.8 ± 3.3 months at 10 hPa in the N48 and N96 simulations, respectively. The differences between the two periods are statistically significant at some but not all heights. Peak westerlies occur above 32 km (10 hPa), reaching 25 m s$^{-1}$ in places. QBO easterlies peak more strongly and occur at lower levels, reaching amplitudes of at least $-25$ m s$^{-1}$ regularly. However, neither set of ensembles exhibits a QBO signal that penetrates right down to 100 hPa, as seen by the observed downward propagation of westerlies, described by Naujokat (1986). On the other hand, Fig. 2 shows a clearer comparison at 90 hPa with a prominent QBO signal present as far down as 90 hPa in the N96 ensemble that is less evident in the N48 resolution run. Both ensembles show a strong and regular SAO in the mid–upper stratosphere, which extends into the upper mesosphere (not shown).

Figure 3 shows, as a function of height, the time-spectral amplitude of tropical winds and the total wind amplitude, displaying the annual, QBO, and SAO cycles, in each of N48, N96, and ERA-40 ensembles, using monthly data. The methods for constructing the total wind amplitudes are outlined in Pascoe et al. (2005) and represent the amplitude associated with the ratio of an integrated range of Fourier harmonics to the total wind variance. The spectral amplitudes for both the N48 and N96 ensemble averages compare well with the ERA-40...
data. The height and magnitude of the peak QBO amplitude are in good agreement. The period for the QBO is slightly longer than ERA-40 (28 months) and is directly related to the magnitude of the prescribed source of gravity waves in the model. No explicit tuning of the tropical variability was made and the same spectral gravity wave parameter values were employed for both N48 and N96.

Little annual tropical variability is displayed by the high-top MetUM in Fig. 3. In ERA-40, the annual cycle appears linked with the SAO near the stratopause: seasonal modulation of the SAO is manifested by a strengthened annual cycle, with stronger SAO amplitudes in NH winter and spring periods (Gray 2010, see also next section); however, little of this is seen in the model and the SAO amplitude is overestimated, especially at N48. The relative strengths of the SAO and QBO appear linked between the two model resolutions: a weaker SAO at N96 is linked with a stronger QBO, compared with N48, and is consistent with systematic differences in wave filtering. At both resolutions, a notable residual in tropospheric wind variance is seen in the troposphere, which is not explained by annual, semi-annual, or QBO variability. This variability is spread across a broad range of frequencies and so does not appear to be associated with narrowband processes (not shown).

Figure 4 shows the various components of the transformed Eulerian mean (TEM) zonal-mean zonal momentum budget acting on the mean flow at 30 hPa in the tropics during the QBO cycle, split into planetary

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**Fig. 1.** Profile time series of monthly-mean, zonal-mean zonal wind averaged between 5°S and 5°N, from a sample member of the (top) N48 and (bottom) N96 ensembles. Westerlies are shaded and the contour interval is 10 m s⁻¹.

**Fig. 2.** Time series of monthly-mean, zonal-mean zonal wind averaged over 5°S–5°N at 90 hPa, from N96 (thick gray) and two N48 ensemble members (black). Units are m s⁻¹.
wave forcing (DIVF), nonorographic gravity wave forcing (NOGWD), and vertical advection (Andrews et al. 1987). Horizontal advection [see Eq. (1) of Part I] and orographic gravity wave forcing were found to be negligible and are not shown. Superimposed is the value of zonal-mean zonal wind $U$. Diagnostics are averaged from 5$^\circ$S to 5$^\circ$N over all complete QBO periods, with each period scaled to a unit time length. The magnitude of the climatological cycles shown is smaller than that of the individual cycles since they are not perfect sinusoids and so some cancellation will occur at each point within the period.

Under both resolutions NOGWD dominates over DIVF. NOGWD leads $U$ so it can be assumed that it drives the changes in $U$. Compared with a similar figure in Scaife et al. (2000) that employed an earlier version of the MetUM, there is larger parameterized wave forcing (NOGWD) in the current model in the tropics. The time evolution of $U$ is dominated by two terms: the NOGWD drives the wind evolution while the vertical advection opposes it. The vertical advection term consists of the background upwelling associated with the upwelling branch of the Brewer–Dobson circulation, and this is expected to oppose the descent of the QBO phases. In addition, there is a local induced meridional circulation that arises in order to maintain thermal wind balance. In the case of the descending easterly phase, this local circulation consists of equatorial upwelling and thus also opposes the descent of the QBO, while in a descending westerly phase the local induced circulation consists of...
descent at the equator, thus reducing the opposition to the descending QBO. This asymmetry can clearly be seen in Fig. 4, where the (vertical) advection is 3 times as large during the period of easterly acceleration (from $\pi/2$ to $\pi$) compared with the westerly acceleration period ($3\pi/2$–$2\pi$). Also noticeable, at both resolutions, is that the resolved waves (DIVF) contribute primarily to the westerly phase of the QBO and reinforce the forcing due to NOGWD.

Finally, when comparing the two different resolutions, the relative length of time of the westerly phase compared with the easterly phase is clearly longer at N96, consistent with Fig. 1. This is linked to a 10% increase at N96 of the integrated vertical component of horizontal momentum flux density of the eastward NOGWD spectrum at 100 hPa, leading to a reduced easterly phase and longer westerly phase in the lower stratosphere. This increased vertical flux leads to a slightly stronger westerly phase-induced meridional circulation, which can then further aid the vertical descent of the QBO deeper into the lower stratosphere, thus leading to the differences seen at 90 hPa in Fig. 2. These differences in NOGWD fluxes must originate from changes in the tropospheric winds that filter the propagation of the specified non-orographic gravity waves, but investigation of these tropospheric differences is outside the scope of this paper.

These results can be compared to those from the MAECHAM5 model reported by Giorgetta et al. (2006), who showed an insensitivity of the QBO period in MAECHAM5 to modest changes in horizontal resolution but an increase in the strength and duration of the westerly phase. However, they propose that the latter is due to an increase in resolved wave driving. They also report that the QBO in MAECHAM5 is principally driven by resolved waves during the westerly phase and parameterized gravity waves during the easterly phase. While we find that the contribution to the easterly QBO phase is increased at N96 (see Fig. 4), it is still small when compared with the NOGWD term.

2) SAO

Figure 5 shows the annual cycle of equatorial zonally mean zonal wind in ERA-40 and the MetUM ensembles. The SAO is conspicuous in both model and observations, with its easterly phase peaking near 1 hPa. Apparent in the ERA-40 data and other observations (Garcia et al. 1997; Gray 2010) is a strong annual cycle modulation, noticeable as a reduction in peak easterlies at 1 hPa during July compared with December. This feature is less prominent in the model, although the N96 ensemble does show a reduced easterly phase compared to the N48 ensemble during these times. In the observations, it is thought that a weakened Brewer–Dobson circulation during this period (i.e., weakened meridional advection of easterly zonal winds from the summer hemisphere) gives rise to the observed annual cycle at these heights (Hitchman and Leovy 1986; Delisi and Dunkerton 1988; Gray 2010). In addition, the observations also show an annual cycle modulation in the amplitude and descent in time below 1 hPa (see bottom panel of Fig. 5). This is linked with momentum forcing from gravity waves, which are sensitive to the background flow through which they propagate. The peak SAO westerlies in ERA-40 are near 1 hPa, contrasting with both sets of model ensembles, which display peak westerlies nearer 0.1 hPa, one month earlier. This suggests differences in gravity wave forcing in the lower mesosphere, although the reliability of the ERA-40 winds above 1 hPa is questionable, since observational constraints are substantially reduced there (Baldwin and Gray 2005).

To better understand the origin in forcing of the modeled easterly phase of the SAO, Fig. 6 shows the annual cycle of the zonal momentum budget at 2 hPa and 5°S–5°N. At N48 the easterly phase of the SAO...
peaks in February and August. Driving this is the meridional advection of easterlies from the summer hemisphere, peaking one month earlier, although no annual cycle is evident in this forcing. Secondary to this is forcing from DIVF and NOGWD, peaking early during the easterly phase (April and November). The vertical advection term is associated with the locally induced meridional circulation required to maintain thermal wind balance. The term appears in quadrature and is upward (to achieve cooling) and therefore opposes the forcing.

The westerly SAO phase occurs at the equinox and coincides with a reduced magnitude of horizontal zonal wind advection associated with the reversal of the Brewer–Dobson circulation. This phase is primarily driven by NOGWD, peaking in strength in March with a second peak in forcing during September. Its descent is also aided by suppressed vertical upwelling and hence reduced vertical advection during this period. The driving and timing of the easterly–westerly SAO phases by horizontal advection/NOGWD is consistent with a study by Richter and Garcia (2006) using the Whole Atmosphere Community Climate Model, version 2 (WACCM2). Resolved wave driving is also strongest in March and September and peaks less strongly during the early Southern Hemisphere winter as expected. At N96, there is a similar momentum forcing of the easterly phase SAO and the strength and seasonality of the resolved wave driving (see Fig. 6). However, driving of the westerly SAO phase is significantly different during the NH spring, with NOGWD reduced in strength. This is partly offset by marginally stronger westerly driving by vertical advection during these times. The presence of additional parameterized wave forcing at N48 is consistent with changes in wave filtering due to reduced wind shear associated with a weaker OBO at N48.

An examination of the meridional advection of zonal wind in ERA-40 (not shown) exhibits strong horizontal advection of easterlies at 2 hPa during the NH winter but vanishing horizontal advection during the SH winter. This gives rise to the strong annual-cycle modulation in the ERA-40 SAO in Fig. 5, which is absent in the model. Examination of the structure of the zonal wind jets in June–August (JJA) (see Fig. 1 of Part I) shows substantial differences in the equatorial meridional gradient in zonal-mean zonal wind between the high-top MetUM and ERA-40 during JJA in the upper stratosphere, with a peak wind difference of 20 m s⁻¹ near 1 hPa. This is likely to account for the differences in horizontal zonal wind advection in Fig. 5 and thus for the absence of an annual cycle in the upper stratosphere in the MetUM.

b. Extratropical variability

1) SUDDEN STRATOSPHERIC WARMINGS

The left panel of Fig. 7 shows histograms of daily September–October temperature averaged between 60° and 90°S at 50 hPa for the two model resolutions and ERA-40. The MetUM exhibits a warm bias of up to 5 K compared to ERA-40 at both resolutions. The shapes of the temperature distributions are similar between the two ensemble sets and exhibit less spread compared with observations.

The right panel of Fig. 7 shows a histogram of daily December–February (DJF) temperature averaged between 60° and 90°N at 50 hPa. The model distributions again show slightly higher mean temperatures but are in much better agreement. Both model and observational distributions are skewed, weighted slightly to higher temperatures (Pascoe et al. 2006), although at both resolutions the model shows a greater proportion of days having temperatures above 215 K, compared against ERA-40.
This skewness is associated with the occurrence of SSWs and suggests an overestimation of these in the high-top MetUM, compared with the observations. Also, minimum temperatures do not fall much lower than 200 K. Temperatures below 196 K are required for the formation of type-1 polar stratospheric clouds and are a necessary condition for heterogeneous ozone-destroying processes (Solomon 1999). It should, however, be emphasized that the model runs were forced by climatological ozone fields and cannot be directly compared with observations and these polar average temperatures do not preclude the presence of cooler, localized regions.

The top panel of Fig. 8 shows the daily evolution of 10-hPa climatological temperatures, averaged between 60° and 90°N, during the extended NH winter (October–June) and the ensemble spread (i.e., ensemble mean ± one standard deviation). The temperatures show reduced variability during times when radiative forcing is controlling seasonal evolution (i.e., during summertime and early autumn), as expected. Significantly greater variability is seen from late December through mid-March. During this time, the NH winter stratospheric vortex undergoes rapid changes due to vertically propagating (and dissipating) Rossby waves. The bottom panel of Fig. 8 corroborates this fact. Here is shown the relative frequency of SSW events, as defined by Charlton and Polvani (2007). The relative frequency of SSWs is greater in the N48 ensemble than in N96 or ERA-40 during mid- to late winter (January–March), corresponding to a SSW frequency of 8.5 decade⁻¹. The frequency of SSWs in the N96 ensemble is reduced during mid–late winter (6.7 decade⁻¹) and is comparable to ERA-40 (6.6 decade⁻¹). Confidence interval estimates were calculated for the N48 and N96 ensembles employing the methodology set out in Charlton et al. (2007). The incidence of SSWs in N96 (but not N48) was found to be indistinguishable from those in ERA-40, at the 95% confidence level. These results suggest that the higher resolution improves the representation of SSW variability in this model.

The reduced incidence of SSW at N96 is most likely associated with subtle changes in circulation impacting
the propagation of Rossby waves into the stratosphere. Support for this hypothesis comes from the analysis of the climatology of resolved waves and parameterized gravity waves performed in Part I (see Fig. 8), which showed a relative drop in DIVF in the mid–upper stratosphere between N48 and N96. Differences in the gradient of wave refractive index were also found at 35°N; linking upper-troposphere circulation with resolved waves entering the stratosphere. Consequently, it was concluded that more resolved waves are directed toward the equator at N96 than at N48.

Figure 9 shows the relationship between early season (January–February) planetary wave activity, as diagnosed by meridional heat-flux anomalies \( \overline{T^*} \) at 100 hPa, 40°–80°N, and late winter (February–March) polar temperatures at 50 hPa, averaged poleward of 60°N (Newman et al. 2001). For both N48 and N96 ensembles, a clear correlation is seen, comparing well with ERA-40. During the NH winter, both N48 and N96 are found to be statistically indistinguishable from ERA-40 (95%). However, significant temperature and meridional heat-flux biases are seen between the model ensembles and ERA-40 during the SH winter (>99%). These figures are useful for inferring mean polar temperature in the absence of dynamical heating from the resolved circulation (Steil et al. 2003; Manzini et al. 2003). That is, they show up possible differences in diabatic heating or circulation due to nonresolved wave forcing. In the present context we find mean polar temperatures of: 191.5 ± 0.6 K (N48), 191.0 ± 0.8 K (N96), and 189.0 ± 1.1 K (ERA-40). These differences are not statistically significant. Both N48 and N96 values for heat-flux anomalies are consistent with CCMVal model intercomparisons and ERA-40 data (Eyring et al. 2006).

The strength of the wintertime stratospheric vortex is known to be related to the phase of equatorial winds (Holton and Tan 1980, 1982; Lu et al. 2008; Gray 2010). Figure 10 shows the composite differences at N48 of...
zonal-mean zonal wind during times when the lower tropical stratosphere winds are easterly (\(< -10 \text{ m s}^{-1}\)) at 30 hPa compared to when they are westerly (\(> 10 \text{ m s}^{-1}\)). The methodology follows that of Calvo et al. (2007, 2009) and calculates bimonthly averages for early winter (November–December) through to late winter (February–March). The zonal wind threshold separating QBO westerly-phase and easterly-phase composites is applied to mean November–December only and defines those years used for the subsequent bimonthly composites. It is evident from the N48 ensemble that the polar night jet (70°N, 5 hPa) is significantly stronger (\(> 9 \text{ m s}^{-1}, 95\%\)) during times when the tropical lower stratosphere is in a westerly QBO phase (cf. 30 hPa). The positive vortex wind anomaly descends and intensifies throughout the winter. A threefold statistically significant structure is seen at equatorial latitudes, in good agreement with ERA-40 analysis (Pascoe et al. 2005). A region of statistically significant differences observed between 1.0 and 0.1 hPa between 30°S and 30°N is linked with the possible QBO modulation of the SAO and is most evident during the early winter (November–January) when the stratospheric SAO is in an easterly phase.

In the corresponding diagnostic at N96 (Fig. 11) similar behavior is seen, but with notable differences. The positive anomaly showing a strengthened polar night is evident and descends throughout the winter. However, it is significantly stronger in the upper stratosphere during early winter, showing a nonlinear change in strength throughout winter with less statistical significance in late winter, possibly because there are only two ensemble members.

The relative strength and timing of the extratropical stratospheric zonal wind anomaly seen during different phases of the QBO is consistent with the relative number and timing of SSWs observed at N48 and N96. There are more mid–late winter SSWs observed at N48 than at N96, coinciding with a larger extratropical stratospheric wind anomaly in late winter seen in Fig. 10 than in Fig. 11.

Further insight into the significance of the composite differences shown in Figs. 10 and 11 can be achieved by comparing mean January–February time series of equatorial zonal wind \(\bar{u}_{eq}\) and polar-cap temperature \(T_{np}\) at 30 hPa (Fig. 12). The top and middle panels show four combined time series of the N48 ensemble while the bottom panel shows the two at N96. Also shown are the correlations between \(\bar{u}_{eq}\) and \(T_{np}\), for each ensemble member. Nondirectional significances are determined using

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 t = \frac{r}{\sqrt{(1-r^2)/(N-2)}},
\]

where \(r\) represents correlation and \(N\) is the number of years in each run (Clarke and Cooke 1992, 343–344). The confidence level is taken to be 100(1 – \(t\)). No allowance is made for possible degrees-of-freedom issues relating to autocorrelation in the equatorial zonal wind.

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**FIG. 10.** Composite differences in zonal-mean zonal wind between times when QBO is in the easterly phase and times during its westerly phase for the N48 ensemble, during (top left) November–December, (top right) December–January, (bottom left) January–February, and (bottom right) February–March. Contouring is 3 m s\(^{-1}\) and shading denotes the 95% confidence level.
The validity of this was tested using data contained in Figs. 3 and 4 of Lu et al. (2008). Similar levels of confidence were obtained for the higher absolute levels of correlation. A wide spread in correlation values is found for $\bar{u}_{eq}$ and $T_{np}$ within the N48 ensemble. The first (0–25 yr) shows little correlation during the period $[-0.01 (5\%)]$ whereas the third (50–75 yr) shows a noticeable degree of anticorrelation $[-0.52 (99\%)]$. During periods showing strong anticorrelation between equatorial 30-hPa zonal wind and polar temperature, the composite analysis used in Figs. 10 and 11 reveals a strong positive extratropical anomaly, as discussed previously. Superficially, the two N96 integrations show little correlation: $0.06 (20\%)$ and $0.09 (29\%)$, respectively. However, within the first member a noticeable anticorrelation is seen during the latter half of the time series: $-0.61 (96\%)$. Similar sporadic behavior was discussed in Lu et al. (2008). Using the ECMWF ERA-40 and operational analyses, they noted a drop in correlation between tropical stratosphere $\bar{u}$ and high-latitude stratosphere $\bar{u}$ and $\bar{T}$ between 1977 and 1997, with significantly larger (anti)correlations outside this period. Sporadically changing variability such as this complicates the significance and interpretation of composite analyses such as shown in Figs. 10 and 11 and studies such as those of Calvo et al. (2007, 2009). Further exploration of the nature of this variability (e.g., the role of interannual variations of SST) is outside the scope of this study.

2) Annular Modes

The behavior of the annular modes has also been examined, with emphasis on the persistence times of annular mode anomalies. The annular modes were identified as the leading empirical orthogonal functions (EOFs) of daily zonal-mean wintertime geopotential height on individual model levels. Two sets of EOF patterns were constructed from height anomalies poleward of 20°, thus obtaining separate patterns for the southern annular mode (SAM) and northern annular mode (NAM). The EOFs were constructed from deseasonalized and linearly detrended data, which were then smoothed using a 90-day filter. The full daily time series of the original model height anomalies were then projected onto the leading EOF pattern of filtered data, retrieving a principal component time series. An autocorrelation analysis was then performed, employing a Gaussian filter to recover time-of-year decorrelation time scales. These were inferred after fitting the autocorrelation function to an exponential. For this study the methodology is taken from Baldwin and Dunkerton (2001), and for comparison, ECMWF ERA-40 and operational analysis daily zonal-mean geopotential height data were used. Significance tests have been applied to assess differences between the MetUM and ECMWF reanalysis time scales (see the appendix).

Figure 13 shows the time evolution of persistence times of the NAM calculated from the N48 and N96 ensembles and ECMWF. Common features appear in the three sets of data, most conspicuous being the elevated residence times in the wintertime troposphere–lower stratosphere and summertime midstratosphere. Significant differences are seen during late summer between the MetUM and ECMWF, where the former show reduced time scales. Previously, these have been interpreted as being linked
with either longer radiative damping times in the stratosphere (Baldwin and Dunkerton 2001) or ozone anomalies persisting from the previous winter (Fioletov and Shepherd 2003). However, elevated time scales peak well into the mid–upper stratosphere (not shown), so it is not evident that radiative relaxation helps to explain these features at upper levels, as these are on the order of days only. MetUM wintertime tropospheric time scales are in broad agreement with the ECMWF data, although dates showing peak persistence extend out to April at N48 (although they are not statistically significant). These features coincide with similarly elevated tropospheric persistence time scales reported by Baldwin et al. (2003). The coverage and distribution of regions showing significance difference with the ECMWF data suggests little difference between N48 and N96.

Figure 14 shows the corresponding height–time evolution of SAM persistence time scales. Long persistence times occur in the stratosphere during midsummer and midwinter at N48 and are longest during late spring early summer at N96. This behavior is distinct from the NH variability in that the extrema do not occur with a 6-month lag. Furthermore, the peak in tropospheric persistence time scales coincides with the summertime peak in the stratosphere. These tropospheric features are also seen in the ECMWF data but show less persistence. The only significant differences occur in the troposphere during September–June at N48, and during spring and February at N96. This is linked with peak elevated time scales occurring over a longer time at N48 and represents a significant improvement at N96.

These results are consistent with those of Gerber et al. (2008a), who examined persistence time scales in the models of the Intergovernmental Panel on Climate Change Fourth Assessment Report (IPCC AR4). They report similarly timed peaks in persistence time scales in these models, with reduced (inflated) tropospheric time scales during the NH (SH) winter (summer). In this study, the reduction in persistences seen between the N48 and N96 ensembles in the stratosphere is consistent with the conclusions of Gerber et al. (2008b). Using a simplified model, they suggest a reduction in persistence time scales with improved horizontal resolution.
4. Discussion and conclusions

In this paper we report on the stratospheric variability of the vertically extended MetUM and compare against the ECMWF ERA-40. We also report on the effects of changing horizontal resolution. The broad features constituting tropical stratospheric variability are reproduced: an internally driven QBO in the stratosphere and a clear SAO in the upper stratosphere and mesosphere. The strength and frequency of the former is in reasonable agreement with ERA-40. The QBO period is sensitive to changes in horizontal resolution, although the effect is not statistically significant at all levels. This is likely due to a 10% change in the vertical flux of horizontal GW pseudomomentum through 100 hPa at N96 resulting from filtering differences from the underlying tropospheric circulation. This also helps to account for a deeper QBO at N96. The stratospheric SAO compares well with ERA-40, although it does not exhibit strong annual-cycle modulation. The lack of an annual cycle in the SAO is symptomatic of a weak tropical annual cycle throughout the stratosphere and mesosphere, which is little improved by resolution changes. We suggest that this is associated with an overly strong stratospheric jet during the NH summer of up to 20 m s$^{-1}$ rather than a lack of an annual cycle in the Brewer–Dobson circulation itself. Similar biases have been noted in other contemporary GCMs and may be linked with the gravity wave driving during these times (Lott et al. 2005; García et al. 2007; Scinocca et al. 2008). Furthermore, a poor
representation of the tropical annual cycle in the SAO is found in coupled chemistry models and has been linked with the fidelity of the underlying QBO: a well-represented QBO is linked with a poorer representation of the SAO and vice versa (Butchart et al. 2010).

Variability of the polar night jet is well reproduced in the high-top MetUM, as determined from the number of sudden stratospheric warming events. Although significantly more frequent in the N48 ensemble (8.5 decade\(^{-1}\)), the rate of these events in the N96 ensemble (6.7 decade\(^{-1}\)) is statistically equivalent (95%) to that diagnosed in ERA-40 (6.6 decade\(^{-1}\)). The reduction in sudden stratospheric warmings in N96 is consistent with reduced resolved wave forcing. Similar sensitivity of SSW frequency to horizontal resolution changes is reported by Richter et al. (2008). They examined the high-top WACCM model for SSW and found the highest incidence in the coarse-resolution configuration (4° × 5°). They attributed this to increased resolved wave forcing during December–March.

The MetUM displays a sensitivity of wintertime polar temperatures to the phase of the tropical stratospheric wind. The nature of this sensitivity is subtly different between N48 and N96: a statistically stronger mid–late winter polar vortex anomaly is seen at N48 and appears linked with a greater incidence of SSWs when the tropical stratosphere is easterly. However, although less significant, the pattern and timing of sensitivity at N96 is qualitatively similar to that reported elsewhere (Calvo et al. 2007, 2009). Our results suggest that the ability of models to reproduce these effects is dependent on how well they generate large-scale resolved waves and the seasonal distribution of SSWs. The Holton–Tan association in the model appears intermittent in a similar way to observed QBO–vortex correlations (Lu et al. 2008), but the underlying reason for this intermittency requires further study (e.g., to explore the role of interannually varying SSTs). Mean SH wintertime polar temperatures are generally too warm, regardless of changes to horizontal resolution, and are most likely related to an overly strong response of the MetUM to dynamical heating.

Differences between the horizontal resolutions are more apparent in the expression of annular mode variability as diagnosed from the persistence time scales of the NAM and SAM. N96 time scales are generally shorter and more consistent with the ECMWF ERA-40 and operational analyses. Most notably, time scales for the tropospheric SAM are significantly reduced during the southern summer. A recent study by Gerber et al. (2008a) reports a bias in the tropospheric SAM time scale in those models used in the IPCC AR4. A similar study examining the CCMVal-2 chemistry–climate models reports a similar bias (Gerber et al. 2010). This study suggests that increased horizontal resolution will reduce these biases. An important corollary from this study is that larger relative errors are associated with longer time scales. Consequently, time scales in the stratosphere, which are generally longer, also have larger associated errors. In the SH, diagnosed time scales are generally longer, and consequently fewer significant differences are expected. This has direct relevance to the interpretation of time scale features and their significance.

In summary, the Met Office stratosphere and mesosphere resolving “high-top” MetUM effectively reproduces the dominant modes of tropical, extratropical, and annular mode variability. Significant improvements following an increase in horizontal resolution include improved frequency of SSWs, deeper penetration of the QBO into the lower stratosphere, and improved residence times of the tropospheric SAM. Although the MetUM represents tropical variability well, little appreciable improvement is found following increased horizontal resolution. This is largely due to an underrepresentation of resolved wave driving in the tropics, especially of the QBO. A further insensitivity to increased horizontal resolution is found in wintertime SH polar temperatures, which is perhaps linked with circulation induced by the prescribed source of parameterized gravity waves.

It is appreciated that the possible benefits following changes in horizontal resolution must be offset against the added computational expense of running at higher resolution. Added resolution can only improve on the fidelity of the modeled tropospheric climate, storm-track variability, and representations of small-scale mixing processes about the tropopause (Greeves et al. 2007), all of which are important in identifying and better interpreting stratosphere–troposphere interactions.

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APPENDIX

Autocorrelation Confidence Intervals

Confidence intervals are calculated using the estimate of the standard deviation of the persistence time scale used in Gerber et al. (2008b),
\[ s(t_N) \approx k \tau^{3/2} N^{-1/2}, \quad (A1) \]

where \( \tau \) represents the e-folding time scale, \( N \) is the length of the time series, and \( k \) is a constant (\( k \sim 2 \)). Here \( N \) is scaled down to cater for the effect of the Gaussian filter used to calculate the time-of-year time scales (full width at half maximum of 60 days), approximated as \( N_{\text{years}} \times 60 \) days. We establish the following null hypothesis: model time scales are indistinguishable from those for the ECMWF ERA-40 and operational analysis data. Accordingly, a \( t \) statistic is defined,

\[ t = \frac{\frac{\tau - \tau_o}{\sqrt{s^2/N + s_o^2/N_o}}}{\sqrt{s^2/N + s_o^2/N_o}}, \quad (A2) \]

where the subscripts denote observationally specific parameters. The \( t \) statistic is used with a degrees-of-freedom estimate (DoF) to calculate a confidence level (Wilks 1995):

\[ \text{DoF} = \frac{\left( (s^2/N) + (s_o^2/N_o) \right)^2}{(s^2/N)^2((N - 1) + (s_o^2/N_o)^2/(N - 1)).} \quad (A3) \]

These confidence intervals are intended as rough estimates only. Observationally, the annular modes are not autoregressive order-1 processes (AR1), which a presumption of an exponential autocorrelation function implies (Priestley 1983). Furthermore, the sample distribution of the lag-autocorrelation function, from which decorrelation time scales are derived, is generally not normally distributed.

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