INFLUENCE OF THE QUASI-BIENNIAL OSCILLATION ON INTERANNUAL VARIABILITY IN THE NORTHERN HEMISPHERE WINTER STRATOSPHERE

by

James A. Anstey

A thesis submitted in conformity with the requirements for the degree of Doctor of Philosophy
Graduate Department of Physics
University of Toronto

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James A. Anstey
Doctor of Philosophy, 2009
Department of Physics, University of Toronto

ABSTRACT

Observations show that the interannual variability of the Northern Hemisphere (NH) extratropical winter stratosphere is strongly correlated with the quasi-biennial oscillation (QBO) of tropical stratospheric winds, particularly during early winter. Most current general circulation models (GCMs) do not exhibit a QBO and therefore do not represent this important mode of tropical-extratropical interaction. In this study we examine the QBO-extratropical correlation using a 150-year GCM simulation in which a QBO occurs.

Since no external forcings or interannual variations in sea surface temperatures are imposed, the modelled tropical-extratropical interactions represent an internal mode of atmospheric variability. The QBO itself is spontaneously forced by a combination of resolved and parameterized waves. The effects of this QBO on the climatological mean state and its interannual variability are considered, both by comparison with a control simulation (also 150 years in length, but with no QBO) and by compositing winters according to the phase of the QBO. Careful attention is given to the definition of QBO phase. Comparisons of the model results with observations (reanalysis data) are also made.

QBO-induced changes in the climatological state of the model are found to have high statistical significance above the tropopause. In the extratropical winter stratosphere, these mean-state changes arise predominantly from the influence of the QBO on the propagation and dissipation of planetary-scale waves. This behaviour is shown to depend on the seasonal cycle, which argues for the usefulness of considering tropical-extratropical interactions in a GCM context. QBO influence on the interannual variability of the extratropical winter stratosphere is also seasonal, and
the tropical-extratropical interaction is sensitive to the phase alignment of the QBO with respect to the annual cycle. This phase alignment is strongly affected by the seasonality of QBO phase transitions, which – due to the QBO being spontaneously generated, rather than having an imposed period – is somewhat realistic in the model. This leads to fluctuations in the strength of the modelled tropical-extratropical interaction occurring on a decadal timescale as an internal mode of atmospheric variability.
Acknowledgements and dedication

My supervisor, Theodore G. Shepherd, has provided invaluable guidance from the inception of this project through to its completion, for which I am deeply grateful. Many others have also given generously of their time and expertise. David Sankey, Mike Lazare, Mike Berkley, Fouad Majaess, Stephen Beagley, Kirill Semeniuk, John Scinocca, and Charles McLandress all provided help with the technical aspects of running the model. For valuable scientific discussions and for fielding my questions, I would like to thank Tiffany Shaw, Thomas Birner, Kirill Semeniuk, John Scinocca, and especially Charles McLandress. My committee members, Kimberly Strong and Paul J. Kushner, provided valuable feedback as the project evolved. And I would like to thank my external examiner, Matthew H. Hitchman, for his thorough and insightful reading of the thesis.

I would like to acknowledge John Scinocca of the Canadian Centre for Climate Modelling and Analysis (CCCma) of Environment Canada for initially obtaining a QBO in CMAM and thereby providing the impetus for me to extend the model runs and examine the extratropical impact of the modelled QBO.

I am grateful to the National Sciences and Engineering Research Council of Canada (NSERC) for financial support during my time as a graduate student.

Finally, I would like to thank my parents. And I dedicate this work to my fiancée, Megan Haggerty, whose patient encouragement has helped see it to completion.
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Glossary of acronyms

3D  Three-dimensional
AMIP  Atmospheric Model Intercomparison Project
AR(1)  First-order autoregressive process
BDC  Brewer-Dobson Circulation
CCCma  Canadian Centre for Climate Modelling and Analysis
CMAM  Canadian Middle Atmosphere Model
DJF  December-January-February average
E  Easterly (from the east, i.e. westward)
ECMWF  European Centre for Medium-range Weather Forecasts
ENSO  El Nino-Southern Oscillation
EOF  Empirical Orthogonal Function
EP flux  Eliassen-Palm flux
ERA-40  ECMWF 40 Year Reanalysis Data
FM  February-March average
GCM  General Circulation Model
GHG  Greenhouse Gas
HT80  Holton and Tan [1980]
HTE  Holton-Tan Effect
JF  January-February average
JFM  January-February-March average
JJA  June-July-August average
L08  Lu et al. [2008]
MAM  March-April-May average
MMC  Mean Meridional Circulation
NCAR  National (USA) Center for Atmospheric Research
NCEP  National (USA) Centers for Environmental Prediction
ND  November-December average
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<td>September-October-November average</td>
</tr>
<tr>
<td>SST</td>
<td>Sea Surface Temperature</td>
</tr>
<tr>
<td>SSW</td>
<td>Stratospheric Sudden Warming</td>
</tr>
<tr>
<td>TEM</td>
<td>Transformed Eulerian Mean</td>
</tr>
<tr>
<td>W</td>
<td>Westerly (from the west, i.e. eastward)</td>
</tr>
<tr>
<td>W-E</td>
<td>Westerly-minus-easterly composite difference</td>
</tr>
</tbody>
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List of symbols

\( a \) \quad \text{Radius of Earth}
\n\( \nabla \cdot \mathbf{F} \) \quad \text{EP flux divergence}
\n\( F(z) \) \quad \text{Vertical component of EP flux}
\n\( F(\phi) \) \quad \text{Meridional component of EP flux}
\n\( F \) \quad \text{Total zonal forcing}
\n\( f \) \quad \text{Coriolis parameter}
\n\( H \) \quad \text{Log-pressure scale height}
\n\( \bar{m} \) \quad \text{Zonal-mean zonal angular momentum}
\n\( P_y \) \quad \text{Meridional gradient of potential vorticity}
\n\( p \) \quad \text{Pressure}
\n\( R \) \quad \text{Ideal gas constant for dry air}
\n\( T \) \quad \text{Zonal-mean temperature}
\n\( t \) \quad \text{Time, or } t\text{-statistic (depending on context)}
\n\( \bar{u} \) \quad \text{Zonal-mean zonal wind}
\n\( \Delta \bar{u}_{EQ} \) \quad \text{Change in tropical stratospheric } \bar{u}
\n\( \bar{v} \) \quad \text{Zonal-mean meridional wind}
\n\( \bar{v}^* \) \quad \text{TEM meridional wind}
\n\( \bar{w} \) \quad \text{Zonal-mean vertical wind}
\n\( \bar{w}^* \) \quad \text{TEM vertical wind}
\n\( w_{ph} \) \quad \text{QBO vertical phase velocity}
\n\( \bar{X} \) \quad \text{Zonal forcing by parameterized gravity waves}
\n\( z \) \quad \text{Log-pressure altitude}
\n\( \theta \) \quad \text{Potential temperature}
\n\( \kappa \) \quad \frac{R}{c_p}, \text{ where } c_p = \text{specific heat capacity at constant pressure}
\n\( \rho_0 \) \quad \text{Basic state density}
\n\( \tau \) \quad \text{Lag with respect to January (Chap. 4)}
\n\( \Phi \) \quad \text{Geopotential}
<table>
<thead>
<tr>
<th>Symbol</th>
<th>Definition</th>
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<tr>
<td>φ</td>
<td>Latitude</td>
</tr>
<tr>
<td>Ψ∗</td>
<td>TEM meridional streamfunction</td>
</tr>
<tr>
<td>ψ</td>
<td>QBO phase angle defined from EOFs (Chap. 4)</td>
</tr>
<tr>
<td>ψ\text{G}</td>
<td>A range of ψ values (Chap. 4)</td>
</tr>
<tr>
<td>\tilde{ψ}\text{G}</td>
<td>An “opposing” range of ψ values (Chap. 4)</td>
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<tr>
<td>ψ\text{G}₁</td>
<td>Lower boundary of ψ-bin (Chap. 4)</td>
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<td>ψ\text{G}₂</td>
<td>Upper boundary of ψ-bin (Chap. 4)</td>
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<tr>
<td>Δψ\text{G}</td>
<td>ψ\text{G}₂ − ψ\text{G}₁, i.e. width of ψ-bin (Chap. 4)</td>
</tr>
<tr>
<td>δψ\text{G}₁</td>
<td>ψ-increment (Chap. 4)</td>
</tr>
<tr>
<td>(−)</td>
<td>Zonal mean</td>
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<tr>
<td>( )′</td>
<td>Eddy (departure from zonal mean)</td>
</tr>
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</table>
Chapter 1

Introduction

1.1 Stratospheric interannual variability

This thesis concerns the interannual variability of the stratospheric circulation during Northern Hemisphere (NH) winter. The strength of this circulation varies widely from year to year, due in large part to the occurrence of stratospheric sudden warmings (SSWs), which are forced by large-amplitude planetary waves that propagate upward from the troposphere to the stratosphere during winter. Dissipation of these waves in the stratosphere disturbs the circulation away from the quiescent radiative equilibrium state of a strong, circumpolar vortex. The NH winter stratospheric polar vortex exhibits disturbed states much more often than does the corresponding vortex in the Southern Hemisphere (SH) winter stratosphere, due to higher topography and greater land-sea contrasts in the NH being capable of forcing larger planetary-wave amplitudes than are found in the SH; consequently the NH exhibits much greater interannual variability than does the SH. In the summer, both hemispheres are comparatively quiescent due to the lack of planetary-wave propagation in the summertime easterly winds.

A number of factors influence both the strength of planetary-wave forcing in the troposphere and the propagation and dissipation of planetary waves in the winter stratosphere. Wave sources in the troposphere may be affected by changes in sea surface temperatures (SSTs) due to external forcings with interannual timescales (such
as radiative forcing trends, e.g. greenhouse gas changes) or to low-frequency modes of natural interannual climate variability such as the El Niño Southern Oscillation (ENSO). Wave propagation in the winter stratosphere is affected by the state of the zonal-mean zonal wind $\bar{u}$, which may be thought of as providing a waveguide for the waves in the sense that it determines their direction of travel and manner of dissipation. This “background” distribution of $\bar{u}$ in the winter stratosphere can, like the tropospheric wave sources, change under the influence of long-term external radiative forcings. It is also affected by internal stratospheric low-frequency variability, such as the quasi-biennial oscillation (QBO) of tropical stratospheric winds.

In this study we examine the influence of the tropical QBO on the interannual variability of the stratospheric polar vortex during NH winter. While the QBO is not the only factor affecting the interannual variability of the NH winter extratropics (as noted above), it correlates with a substantial fraction of that variability and thus appears to be an important mode of natural – i.e., not externally forced – variability. Accurate detection and prediction of the climate (i.e., long-term) response to external forcings requires an understanding of the natural variability of the climate system (e.g., Palmer [1999]). The interaction between the QBO and the polar vortex may be thought of as internal atmospheric variability in the sense that no external forcings or interannual variations in SSTs are required in order for it to exist.

A practical motivation for understanding the interannual variability of the NH stratospheric polar vortex is the problem of predicting ozone depletion, since cold NH winters may potentially create conditions favourable for ozone destruction similar to those seen in the SH polar vortex – which, unlike the NH vortex, is almost always very cold and undisturbed – for as long as the concentration of ozone-depleting substances remains high. Proper quantification of stratospheric variability may also be

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1 E.g., Dunkerton and Baldwin [1991] found a correlation coefficient of 0.64 between 40 hPa QBO phase and midwinter polar vortex strength using 1964-1988 data, indicating that roughly 40% of vortex variability in their 25-year record is accounted for by the QBO (in a linear regression sense [Wilks, 2006]).

2 “Internal” and “natural” variability are interchangeable terms if we are concerned only with the atmosphere, in isolation from other parts of the climate system. The term “natural variability”, however, may be more general in that it may also refer to interactions between different parts of the climate system (e.g. atmosphere-ocean coupling, solar cycle effects, etc.).
important for intraseasonal prediction, as well as predictions of climate on decadal timescales [Keenlyside et al., 2008]. Recent evidence from observations and models has indicated the importance of annular modes that couple together stratospheric and tropospheric variability [Baldwin and Dunkerton, 1999, 2001], and which may be poorly represented in models that do not include an adequate representation of the stratosphere [Norton, 2003]. Reliable modelling of stratospheric variability may therefore be an important part of successfully capturing the natural variability of the tropospheric circulation.

As stated above, this thesis concerns the influence of the QBO in the tropical stratosphere on the interannual variability of the stratospheric polar vortex during NH winter. Here in Chap. 1, we begin with a review of the literature concerning the QBO and the observed QBO-vortex interaction, which is often referred to as the Holton-Tan effect (HTE). This is followed by an overview of relevant modelling studies, both with general circulation models (GCMs) as well as simpler mechanistic models. An outline of the rest of the thesis is then given, followed by a description of the datasets that are employed.

1.2 The Quasi-Biennial Oscillation and the observed Holton-Tan effect

The QBO in the tropical stratosphere consists of alternating regimes of westerly (W) and easterly (E) zonal-mean zonal wind, $\bar{u}$, that propagate downward at a rate of $\approx 1$ km/month [Baldwin et al., 2001]. Its discovery in 1960 [Reed et al., 1961; Ebdon, 1960] was followed less than a decade later by a theory that explained the downward propagation as being caused by wave-mean flow interactions between $\bar{u}$ and upward-propagating tropical waves that dissipate in the stratosphere [Lindzen and Holton, 1968]. Unsuccessful earlier attempts to explain the tropical QBO had looked for an extratropical cause, being motivated by the fact that an extratropical QBO signal in stratospheric $\bar{u}$ is also observed [Baldwin et al., 2001]. Noting the apparent success
of the Lindzen-Holton theory in explaining the tropical QBO in terms of forcing
by tropical waves, Holton and Tan [1980; HT80] suggested that the problem of the
extratropical QBO should be posed in reverse, and asked whether the extratropical
QBO signal might be explained as a consequence of the tropical QBO. (Hereinafter
we adhere to modern terminology and refer to the tropical QBO simply as the QBO.)

HT80 hypothesized that the modulation by the QBO of the position of the low-
latitude $\bar{u} = 0$ line would affect the propagation and dissipation of extratropical
planetary waves in the winter stratosphere in such a way as to generate the extratrop-
cal QBO signal. While the summer stratosphere contains easterlies ($\bar{u} < 0$), in
which stationary planetary waves$^3$ are evanescent, the seasonal changeover to win-
ter westerlies allows stationary planetary waves to propagate upward into the deep
stratosphere [Charney and Drazin, 1961]. The propagation and dissipation of plan-
etary waves is strongly affected by the background mean flow profile, $\bar{u}(\phi, z)$ (where
$\phi$ is the latitude and $z$ the altitude), on which the waves propagate [Matsuno, 1970;
Karoly and Hoskins, 1982]. The QBO, in turn, modulates the structure of $\bar{u}(\phi, z)$,
and therefore modulates the position of the low-latitude $\bar{u} = 0$ line, which separates
the “winter” and “summer” regions of the stratosphere. Stationary planetary waves
are confined to the winter hemisphere by the $\bar{u} = 0$ line. Hence the QBO may,
HT80 suggested, act to modulate the meridional extent of the region in which these
planetary waves are able to propagate and dissipate.

The hypothesis of HT80 was motivated by their analysis of 16 years of National
Meteorological Center (NMC) data (1962-1977), in which they presented statistical
evidence of an association between the QBO and the extratropical stratospheric flow.
By compositing the data according to the phase of the QBO at 50 hPa, they showed
the westerly phase of the QBO (QBO-W) to be associated with anomalously low 50
hPa geopotential height at high latitudes – corresponding to a stronger polar vortex
– in both early (Nov-Dec) and late (Jan-Mar) NH winter. Anomalies of the oppo-

$^3$Stationary planetary waves are Rossby waves of planetary scale that have phase speed $c = 0$
with respect to the ground. In the NH they are the dominant component of stratospheric variabil-
ity because their forcing mechanisms (topography and land-sea thermal contrasts) are themselves
stationary.
site sense were associated with the easterly QBO phase (QBO-E). This correlation between QBO phase and the stratospheric polar vortex has subsequently been referred to as the Holton-Tan effect (HTE). Since the strength of the polar vortex is modulated by planetary wave forcing originating from the troposphere (as described at the beginning of Sec. 1.1), HT80 supposed that the effect of the QBO on planetary waves (as described in the preceding paragraph) constituted the causal linkage between QBO phase and polar vortex strength.

Given that planetary waves provide the forcing for SSWs [Matsuno, 1971], which contribute strongly to the interannual variability of the NH winter stratosphere, many studies have focused attention on the link between QBO phase and the occurrence of SSWs. Labitzke [1982] examined a 26-year record of 30 hPa North Pole temperatures and found midwinter SSWs to be less common during QBO-W, although periods of solar maxima (i.e., maxima of the 11-year solar 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 $\bar{u} = 0$ line appeared to accompany SSW development. McIntyre [1982] stressed the importance of a “preconditioning” of the stratospheric flow 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. Wallace and Chang [1982] noted, however, 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 pointed out that the HTE need not be manifest only through SSWs, noting that HT80 had attempted some separation of SSW and non-SSW effects by considering early (Nov-Dec) and late (Jan-Mar) winter separately. (Major SSWs tend to occur only in Dec or later.)

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 when the QBO was in a deep W 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. Labitzke [1987] and Labitzke and van Loon [1988] cast further doubt on the robustness of the QBO-SSW association by showing that the 11-year solar cycle appeared to modulate the HTE: during QBO-W winters, a strong positive correlation was found between north pole temperature and 10.7 cm radio flux (which is a proxy for solar irradiance variations due to the 11-year sunspot cycle), while a weaker, negative, correlation was found during QBO-E. However, the statistical significance of this association was questioned by Baldwin and Dunkerton [1989].

HT80’s analysis was updated using a longer record of NMC analyses (1964-1988) by Dunkerton and Baldwin [1991]. They showed that while the HTE appeared to weaken in the longer record, its statistical significance remained high even when they considered the possibility of the vortex exhibiting a lag-1 year anticorrelation unrelated to the QBO. They found that defining QBO phase according to the 40 hPa equatorial wind gave the strongest correlations with the extratropics. The 1979-1988 data, which incorporated satellite measurements, allowed Dunkerton and Baldwin [1991] to diagnose the HTE at altitudes up to 1 hPa. They showed that while Eliassen-Palm flux (hereinafter “EP flux”) convergence anomalies in the W and E categories were consistent with the sign of the extratropical $\bar{u}$ anomalies, planetary wave amplitudes differed only slightly between the two categories. A further update of HT80 was provided by Naito and Hirota [1997] using Freie Universität Berlin data for 1957-1994, who found that the early winter (Nov-Dec) HTE appeared to be unrelated to solar cycle variations while the late winter (Jan-Feb) HTE obeyed the modulation described by Labitzke and van Loon [1988]. Naito and Hirota [1997] argued that the reason why HT80 found a strong HTE was that their analysis period (1962-1977) contained two solar minima and one solar maximum. They suggested that the polar vortex is influenced strongly 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], who analyzed 12 years of NH extratropical NMC data and concluded that the four different combinations of QBO and solar cycle phase\(^4\) appeared to induce different patterns of anomalous stratospheric \(\bar{u}\) in early winter, which then converged to become similar patterns in late winter. He suggested that the early winter influence of those forcings at low latitudes in the upper stratosphere created a polar signal by means of downward and poleward propagation of the \(\bar{u}\) anomaly due to wave-mean flow interaction. He characterized this as a natural mode of extratropical stratospheric variability, analogous to behaviour seen in idealized stratospheric models [Holton and Mass, 1976], that could be excited by forcings such as the QBO or solar cycle. A similar point of 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 in the layer above 10 hPa, extending from NH polar latitudes to \(30^\circ\)S, and also stressed the vertical coherence of the NH signal (which extended over the whole depth of the extratropical stratosphere). Baldwin and Dunkerton [1991] pointed out that the strong signal at high altitudes may or may not indicate a causal role for the flow at those altitudes, and suggested that the HTE 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) diagnosis of NH \(\bar{u}\) and zonal-mean temperature \(\bar{T}\) and showed that the leading mode of variability coincided well with the spatial structure of correlations between the QBO and the extratropics during Dec, Jan and Feb, but less well with the solar cycle correlations in the same months. They noted that most of the interannual variance (of DJF-averaged \(\bar{T}\)) at high latitudes could be represented by two signals: the QBO, and a term representing the modulation of the QBO influence by “quasi-decadal variability” (QDV). The QDV-modulation terms

\(^4\)That is, the combinations W/max, W/min, E/max and E/min.
1. Introduction

also yielded a spatial correlation pattern, in Feb, similar to that of the leading EOF of $\bar{T}$. This led Dunkerton and Baldwin [1992] 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. A longer record, Dunkerton and Baldwin [1992] argued, would be required for determination of the true role of the solar cycle. Kodera [1993] further clarified the behaviour of the QDV by calculating the running correlation between north-polar temperature and 45 hPa QBO wind. He showed that decadal variability in the strength of the HTE was not a spurious effect, as had previously been argued by Salby and Shea [1991]. The time variation of the decadal-varying HTE modulation was diagnosed by demodulation of the polar temperature (Kodera [1993], Fig. 3(c)) without any reference to the solar cycle. Thus while the existence of a QDV modulation of the HTE was succinctly demonstrated, its explicit connection to the solar cycle remained (and remains) an open question.

Since a stratospheric polar vortex also occurs during winter in the SH, there is no particular reason to expect the variability characteristic of the HTE to occur only in the NH. The HTE in the SH was examined by Baldwin and Dunkerton [1998] using NCEP analyses spanning 1978-1996 and pressure levels from 1000 hPa to 1 hPa. They represented the QBO in terms of a “phase angle” defined from the two leading EOFs of the tropical radiosonde data [Wallace et al., 1993] and used this succinct representation of the QBO state to optimize the extratropical HTE signals in both hemispheres. The optimal NH signal was found to be equivalent to defining QBO phase by the 40 hPa equatorial wind (consistent with Dunkerton and Baldwin [1991]), while the 25 hPa equatorial wind was found to optimize the SH signal. The timing and magnitude of the QBO-vortex interaction during SH winter was found to differ markedly from that of the NH winter: QBO-W (at 25 hPa) coincided with anomalous westerlies only in midlatitudes (rather than polar latitudes) during May-Oct, followed by a strengthened polar vortex – i.e., a delay in the final warming – in Nov. Naito [2002] also considered the HTE in the SH, finding similar results to Baldwin and Dunkerton [1998]. Analyzing the influence of
the QBO on EP fluxes, Naito [2002] argued that QBO modulation of upward wave fluxes (in the extratropics) 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.

Although particular single levels are found to optimize the correlations between the extratropics and the QBO (40 hPa for the NH and 25 hPa for the SH), why this is so remains an open question, and it has been argued on physical grounds that the tropical winds over a deep layer should be more influential than the winds at any single level [McIntyre, 1982; Dunkerton et al., 1988]. Using rocketsonde data extending to 58 km, Gray et al. [2001b] demonstrated high correlations between Jan-Feb north-polar temperature at 30 hPa and equatorial winds near the stratopause during the preceding Aug-Sep. The focus on equatorial winds at lower altitudes by earlier authors was no doubt 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 ERA-40 data, the QBO has recently been demonstrated to extend across the whole depth of the stratosphere, with a three-cell vertical structure [Pascoe et al., 2005], although its magnitude in the 40-50 km layer is diminished in comparison with the 20-40 km layer\(^5\).

In summary, the observed HTE in both the NH and SH involves vertically deep fluctuations in \( \bar{u}, \bar{T} \) or \( \Phi \) that are of hemispheric scale, and which may represent an internal mode of stratospheric variability that is excited by the QBO. It is possible to optimize the HTE signal by choosing a specific vertical level to define QBO phase, but the downward propagation of the QBO signal results in more than one level appearing to have a degree of influence over the extratropics, and the magnitude and the timing of this influence varies with the choice of level. It is not obvious on physical grounds which level, or range of levels, should be most important. This question cannot be resolved from observations alone, but the key point to take from the observations is that the QBO-vortex interaction is characterized by coherent variability across the depth

\[^5\]The W-E composite differences at the equator yield a peak-to-peak QBO strength of 30 m/s at 23 km, 25 m/s at 35 km, and 10 m/s at 48 km [Pascoe et al., 2005].
of the stratosphere in both the tropics and extratropics. The vertical coherence of the signal in the extratropical winter stratosphere suggests that the QBO excites a deep hemispheric mode of natural variability. This mode may not necessarily be isolated to the stratosphere, since it has been shown to exhibit coherence with tropospheric variability [Baldwin and Dunkerton, 1999; Thompson and Wallace, 2000]. Through this mode, a connection between the QBO and NH winter weather comparable in strength to the influence of ENSO has been demonstrated [Thompson et al., 2002].

In closing our review of the observational studies, it should be noted that the statistical significance of the observed correlation between equatorial winds and NH stratospheric polar vortex strength, originally discovered by HT80 in 16 years of data, has been mostly retained in the longer records provided by global reanalysis datasets. However, there is recent evidence that the strength of the HTE may fluctuate on decadal timescales [Lu et al., 2008]. Whether or not these fluctuations are physically related to the 11-year solar cycle remains an open question. As noted above, previous studies (Baldwin and Dunkerton [1991], Dunkerton and Baldwin [1992], Kodera [1993]) have argued that the HTE represents the QBO projection onto an intrinsic mode of extratropical variability, and that other influences might also project onto this mode. While the solar cycle may be one such influence, a longer observational record is required before its role can be conclusively established.

1.3 Modelling studies of the Holton-Tan effect

Sec. 1.2 described the observed correlations between the QBO and the winter stratospheric polar vortex, and the hypothesis that this correlation is caused by the QBO modulating the characteristics of the stratospheric planetary waveguide (in particular, the position of the low-latitude $\bar{u} = 0$ line). Testing this causal mechanism is a task for modelling studies. Most such studies have utilized so-called mechanistic models, which are designed to capture the essential mechanisms governing the large-scale dynamics of the middle atmosphere in a simplified manner. A smaller number of studies have addressed the HTE in a GCM context.
1.3.1 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 (PE) 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 amplitude $\Phi_0 = 350$ m and zonal wavenumbers $k = 1$ or 2. Running the model under conditions representative of NH winter, with imposed deep ($\approx 40$ km, over $z \approx 10-50$ km) tropical $\bar{u}$ anomalies (hereinafter $\Delta \bar{u}_{EQ}$) of opposing sign, 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 3D PE model with a fixed $k = 1$ perturbation at the model’s 100 hPa lower boundary, run under Jan-Mar radiative conditions, with imposed ($\approx 15$ km deep) W and E $\Delta \bar{u}_{EQ}$ in the lower stratosphere ($z \approx 20-35$ km). Holton and Austin [1991] found that for strong wave forcing ($\Phi_0 \geq 400$ m) there was no apparent relation between SSW occurrence and the sign of $\Delta \bar{u}_{EQ}$, similar to Bridger [1984]’s results. However, by varying the forcing amplitude $\Phi_0$, Holton and Austin [1991] found SSWs to be sensitive to $\Delta \bar{u}_{EQ}$ for an intermediate range of $\Phi_0$. For small wave forcing amplitudes ($\Phi_0 \leq 200$ m) no SSWs occurred at all, irrespective of $\Delta \bar{u}_{EQ}$. Thus the results of Holton and Austin [1991] suggested that while very quiescent or very disturbed NH winters are insensitive to tropical winds, there is an intermediate range of forcing amplitudes for which a sensitivity to QBO phase does exist. Holton and Austin [1991] found that $\Delta \bar{u}_{EQ} < 0$ tended to hasten the onset of SSWs, apparently in agreement with the sign of the observed HTE.

Subsequent and similar mechanistic modelling studies have tended to support Holton and Austin [1991]’s results. O’Sullivan and Young [1992] found that positioning $\Delta \bar{u}_{EQ}$ in the lower stratosphere ($z \approx 20-35$ km) yielded a stronger HTE than when $\Delta \bar{u}_{EQ}$ was shifted upward by 9 km; they also found the HTE strength to increase when the latitudinal width of $\Delta \bar{u}_{EQ}$ was increased. O’Sullivan and Young [1992] emphasized the importance of the development of a nonlinear Rossby wave crit-
ical layer – often referred to as the “stratospheric surf zone” [McIntyre and Palmer, 1983] – in communicating the influence of $\Delta \bar{u}_{\text{EQ}}$ to high latitudes, arguing that the case of vanishingly small $\Phi_0$ 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 $\Delta \bar{u}_{\text{EQ}}$ 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 (Jan-Feb) Holton and Austin [1991] had avoided the period during which the observed HTE 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 $\Phi_0$ behaved as a bifurcation parameter and that the value of $\Phi_0$ separating quiescent from disturbed winters was lower when $\Delta \bar{u}_{\text{EQ}} < 0$ than when $\Delta \bar{u}_{\text{EQ}} > 0$ (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 of the bifurcation point (in terms of $\Phi_0$) became somewhat smeared out. O’Sullivan and Dunkerton [1994] also showed that the inclusion of parameterized gravity wave drag tended to reduce the HTE strength to realistic levels$^6$. A failing of their model, O’Sullivan and Dunkerton [1994] noted, was that it produced a HTE only in Dec-Feb, whereas the observed HTE also occurs earlier in winter – a deficiency shared by previous studies (Holton and Austin [1991], O’Sullivan and Young [1992]).

The HTE has also been the subject of studies using single-layer models, which are intended to represent a deep, vertically averaged stratospheric layer. Such models have been shown to give useful insight into the behaviour of nonlinear Rossby wave critical layers [Juckes and McIntyre, 1987] which have been argued (e.g. O’Sullivan and Young [1992]) to communicate the influence of $\Delta \bar{u}_{\text{EQ}}$ to high latitudes. O’Sullivan and Salby [1990] used a one-layer model with realistic radiative damping under perpetual January conditions to investigate the effect of $\Delta \bar{u}_{\text{EQ}}$ on the behaviour of the crit-

$^6$Without the parameterized drag, the HTE in their model reflected the contrast between a disturbed polar vortex and a vortex in radiative equilibrium, yielding an extratropical $\bar{u}$ difference far in excess of the observed HTE strength.
ical layer. They argued that the large latitudinal scale of such a critical layer, which may span tens of degrees of latitude under realistic NH winter forcing conditions, provides a plausible means of coupling between low and high latitudes. They also noted that the tendency of Rossby wave breaking (RWB) to displace and distort the polar vortex from zonal symmetry was opposed by radiative damping, which acts to relax PV gradients towards a zonally symmetric radiative equilibrium state [Butchart and Remsberg, 1986] and thus maintains wavelike eddy motion (as opposed to wave breaking). They showed that the ability of RWB to overcome the radiative drive and break up the vortex was more pronounced (for a given $\Phi_0$) when the $\bar{u} = 0$ line occurred further poleward in the NH.

Neglecting radiative effects, the response of high latitudes to $\Delta \bar{u}_{EQ}$ was studied in adiabatic one-layer models by Chen [1996] and O’Sullivan [1997]. Chen [1996] found that the strength of the subtropical $\bar{u}$ 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 $\Delta \bar{u}_{EQ}$. 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 seems, however, to depend 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. Polvani et al. [1995] did not specifically consider the HTE; however, they showed that when their shallow water model was initialized with a $\bar{u}(\phi)$ profile typical of NH winter mid-stratospheric conditions, with the $\bar{u} = 0$ 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 $\phi \approx 55^\circ N$ (see Polvani et al. [1995], Fig. 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. Hence the evidence from one-layer model studies
suggests that while RWB provides the basic mechanism for communication between low and high latitudes, the effects of radiation in opposing the erosion of PV gradients by RWB cannot be neglected. The importance of radiative effects suggests that a full explanation of the HTE may require consideration of the seasonal cycle, which has usually been neglected in mechanistic studies (O'Sullivan and Dunkerton [1994] being an exception).

As well as idealized radiative driving, most mechanistic studies have also considered only idealized forms for $\Delta \bar{u}_{EQ}$. The 3D studies cited above (Bridger [1984], Holton and Austin [1991], O’Sullivan and Young [1992], O’Sullivan and Dunkerton [1994]) all imposed $\Delta \bar{u}_{EQ}$ as a single, deep layer of either W or E winds, neglecting the fact the real QBO typically exhibits oppositely-signed $\bar{u}$ at lower and mid-stratospheric levels\(^7\). While this is a useful idealization for studying the first-order effect of $\Delta \bar{u}_{EQ}$ on the extratropics, it makes the connection between these studies and the observed HTE ambiguous. Such studies also neglect the downward propagation of QBO phase. In a more recent study, Gray \textit{et al.} [2001a] relaxed a 3D mechanistic model towards observed tropical radiosonde (over a 33-year period) and tropical rocketsonde winds (over a 23-year period). Radiosonde data does not extend higher than $\approx 32$ km (10 hPa), while rocketsondes reach as high as $\approx 58$ km (0.3 hPa); hence Gray \textit{et al.} [2001a]’s study contrasted the HTE for cases of shallow ($z \approx 16$-32 km) and deep ($z \approx 16$-58 km) $\Delta \bar{u}_{EQ}$ – and, additionally, incorporated realistic QBO phase propagation. They found the HTE to be stronger for the case of deep $\Delta \bar{u}_{EQ}$, suggesting that the lowermost equatorial stratosphere may not be the most important region from which low-latitude influence extends to higher latitudes. Gray \textit{et al.} [2001a] ruled out the possibility that the HTE could be caused only by $\Delta \bar{u}_{EQ}$ at higher altitudes, finding that no HTE 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 [1992]’s results). Thus they demonstrated the importance of $\Delta \bar{u}_{EQ}$ over

\(^7\)Holton and Austin [1991] did prescribe upper and lower $\Delta \bar{u}_{EQ}$ of opposite sign, but their upper cell was weak and did not induce a shift of $\bar{u} = 0$ between hemispheres; hence their $\Delta \bar{u}_{EQ}$ effectively constituted a single-layer perturbation.
the whole depth of the stratosphere.

Further exploring the effect of a very deep $\Delta \bar{u}_{EQ}$, Gray et al. [2003] imposed a single-layer $\Delta \bar{u}_{EQ}$ 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 $\Delta \bar{u}_{EQ}$ for an intermediate range of $\Phi_0$ values (with $\Delta \bar{u}_{EQ} < 0$ leading to more disturbed winters). They emphasized that planetary waves have deep vertical wavelengths, roughly the depth of the stratosphere, and hence should be sensitive to stratospheric $\bar{u}$, including $\Delta \bar{u}_{EQ}$, over the whole stratosphere. In another study using the same model, Gray [2003] contrasted the effects of $\Delta \bar{u}_{EQ}$ 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) $\Delta \bar{u}_{EQ}$, while the lower $\Delta \bar{u}_{EQ}$ mainly affected the early winter evolution. Gray et al. [2004] considered yet another form of $\Delta \bar{u}_{EQ}$, imposing E anomalies of $\approx 10$ km depth$^8$ at a series of altitudes that were moved upward in 5 km increments: 20-30, 25-35, 30-40, 35-45, and 40-50 km. The most pronounced effect occurred in the 30-40 km experiment, with the poleward-shifted $\bar{u} = 0$ 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 (see Gray et al. [2004], Fig. 4); hence the major effect of $\Delta \bar{u}_{EQ}$ 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 $\Delta \bar{u}_{EQ}$ cases).

One issue that complicates the study of the HTE using observations, as well as GCMs, is the fact that stable statistics of polar vortex interannual variability may require very long records. In this regard, the lower computational cost of mechanistic models gives them an advantage over GCMs. Naito et al. [2003] used a 3D PE model, including a simplified troposphere, with $k = 1$ surface topography and performed 12,000-day integrations under nine different idealized $\Delta \bar{u}_{EQ}$ cases. The vertical structure of their $\Delta \bar{u}_{EQ}$ was single-celled and strongest in the $z \approx 20-30$ km

$^8$The anomaly was strongest in a 10 km layer; tapering above and below this layer led to the actual layer of imposed E winds occupying a $\approx 14$ km depth.
layer. They found that a weaker vortex occurred for $\Delta \bar{u}_{EQ} < 0$, as well as for extremely strong ($> 40$ m/s) positive $\Delta \bar{u}_{EQ}$, and explained their results using index of refraction arguments. Naito and Yoden [2006], also using very long integrations, examined the effect of $\Delta \bar{u}_{EQ}$ on planetary wave structures associated with SSWs. They considered eight $\Delta \bar{u}_{EQ}$ cases, with the vertical structure of $\Delta \bar{u}_{EQ}$ in the different cases being somewhat representative of the QBO at different stages of its downward propagation (although their $\Delta \bar{u}_{EQ}$ was strongest in the lower stratosphere and extended to an unrealistically low altitude, occupying the layer between 200 hPa and 5 hPa). They found equatorward planetary wave propagation to be enhanced for the $\Delta \bar{u}_{EQ} > 0$ 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 $\Delta \bar{u}_{EQ} > 0$ cases). Naito and Yoden [2006] also highlighted a statistically significant difference in upward EP flux in the troposphere at $\phi \approx 45^\circ$N: both before and after SSW events, greater upward flux occurred for the $\Delta \bar{u}_{EQ} > 0$ 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]: because of 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)$^9$. The planetary wave response to a given $\Delta \bar{u}_{EQ}$ evidently can involve changes in wave structure (hence, in EP flux) over a deep vertical scale encompassing both the stratosphere and troposphere.

The only mechanistic modelling study to consider a downward-propagating QBO is that of Hampson and Haynes [2006], who specified $\Delta \bar{u}_{EQ}$ 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 $\Delta \bar{u}_{EQ}$ on the HTE. They found the strength of the HTE to be strongly affected by QBO phase alignment, and

$^9$Gray et al. [2003] pointed out that this assumption is probably a good one for waves of small vertical wavelength such as gravity waves, but may be less applicable to planetary waves.
also noted that their model obtained a strong HTE with $\Delta \bar{u}_{EQ}$ specified only in the lower stratosphere, in contrast to the results of Gray and collaborators.

In summary, the range of mechanistic modelling studies to date appear to demonstrate that a range of HTE behaviours are possible. Precise comparisons of the different studies are difficult because of the varied nature of choices made for $\Delta \bar{u}_{EQ}$. However, all studies agree that a poleward-shifted $\bar{u} = 0$ line may affect the polar vortex when the planetary wave forcing is of intermediate strength. Although the effect of poleward-shifted $\bar{u} = 0$ 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 $\bar{u} = 0$ 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 $\bar{u} = 0$ line is located in the NH. Mechanistic studies have also shown that the strength and timing of the influence of $\Delta \bar{u}_{EQ}$ on the vortex is likely to be sensitive to the vertical structure of $\Delta \bar{u}_{EQ}$ over the whole depth of the stratosphere.

1.3.2 General circulation models

Two common deficiencies suffered by many GCMs are a poor representation of polar vortex variability and the absence of a QBO. The possibility that these problems are connected provides a motivation for GCM studies of the HTE. There exist a number of studies in which a spontaneous QBO\(^{10}\) was obtained in a GCM, but the HTE was not examined [Takahashi, 1996; Horinouchi and Yoden, 1998; Takahashi, 1999; Hamilton et al., 1999; Scaife et al., 2000; Hamilton et al., 2001; Giorgetta et al., 2002, 2006]. Here we focus on those studies that do consider the HTE.

Two early GCM studies were performed by Hamilton [1998] and Niwano and Takahashi [1998]. Hamilton [1998] used an imposed zonal momentum forcing to

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\(^{10}\)By “spontaneous”, it is meant that some combination of resolved and parameterized tropical waves were able to induce the GCM to exhibit a QBO. The state of the art in GCM simulation of QBO-like oscillations is such that spontaneously forced QBOs often possess unrealistic features.
generate a QBO with properties closely matching those of the observed QBO, while Niwano and Takahashi [1998] utilized a spontaneously forced – and hence somewhat less realistic – QBO\textsuperscript{11}. Both studies found the polar vortex to be weaker during QBO-E, although not during early winter (Hamilton [1998] showed a DJF effect while Niwano and Takahashi [1998] showed a JFM effect). Hamilton [1998] obtained more reliable statistics by employing a 48-year model run as opposed to Niwano and Takahashi [1998]’s 15-year run, and emphasized the fact that a long record may be required for stable statistics of the HTE to be obtained\textsuperscript{12}.

Besides Hamilton [1998] and Niwano and Takahashi [1998], only two other GCM studies have focused solely on the HTE as an internal mode of atmospheric variability (i.e. without considering external forcings). Such studies are important because they allow for unambiguous detection of the HTE and its QDV. Pascoe et al. [2006] used an imposed forcing, similar to Hamilton [1998], to relax equatorial winds towards specified semiannual oscillation (SAO), shallow QBO (with $\Delta \bar{u}_{EQ}$ occupying the range $z \approx 15-35$ km), and deep QBO (with $\Delta \bar{u}_{EQ}$ occupying the range $z \approx 15-65$ km) states, performing a 28-year run for each case. They found the timing of midwinter SSWs to be advanced in the deep QBO case, while the imposed shallow QBO affected only their early winter (Nov) warmings. Hence Pascoe et al. [2006] emphasized that realistic upper stratospheric tropical wind variability may be required for GCMs to properly simulate polar variability. Calvo et al. [2007] examined the HTE in a GCM with a spontaneous QBO [Giorgetta et al., 2006]. Using a 50-year model run containing 20 complete QBO cycles, they found the NH winter polar vortex to be stronger (weaker) when the QBO at 30 hPa was W (E). Similar to earlier GCM studies (Hamilton [1998], Niwano and Takahashi [1998]), the HTE in Calvo et al. [2007] occurred only from Dec-onwards.

Other GCM studies of the HTE have focused on the combined effects of the QBO

\textsuperscript{11}Niwano and Takahashi [1998]’s QBO, which was achieved in a GCM with 500 m vertical resolution, had a 1.3 year period and a W bias that was particularly strong in the 50-80 hPa layer.

\textsuperscript{12}See Hamilton [1998]’s Fig. 10, which shows that the spatial structure of correlations between the QBO ($\bar{u}$ at 40 hPa) and the extratropics ($\bar{u}$ at all other gridpoints) differs between the first 23 years and the last 22 years of the 45-year analysis period.
and 11-year solar cycle. Kodera et al. [1991] performed ten 60-day runs under conditions of two different QBO phases with somewhat realistic structure (imposed by the same method used by Hamilton [1998]) and five different solar forcing conditions. They obtained an effect similar to that observed by Labitzke and van Loon [1988], but the magnitudes of their solar forcings were unrealistically large\textsuperscript{13}. Matthes et al. [2004] also used an imposed QBO, with realistic two-celled structure (covering the range $z \approx 20$-$40$ km), and imposed solar cycle radiative forcing of realistic magnitude\textsuperscript{14}. Performing one 15-year run for each of the four QBO/solar cases (i.e. QBO-W + SC-max, QBO-W + SC-min, etc.), they also found behaviour reminiscent of Labitzke and van Loon [1988]'s results. Palmer and Gray [2005] used a GCM with a spontaneous QBO [Scaife et al., 2000] and imposed solar cycle forcing and, with four 28-year runs, found similar results. However, like Kodera et al. [1991] they used unrealistically large solar forcing. Matthes et al. [2004] is, to date, the only GCM study to examine combined QBO and solar cycle influences on the polar vortex using realistic solar irradiance changes.

To summarize our review of GCM studies, the HTE has been examined both on its own (as a mode of internal atmospheric variability) as well as in conjunction with the solar cycle. The only studies with spontaneous QBOs that have considered the HTE are Niwano and Takahashi [1998] (with only a 15-year run), Palmer and Gray [2005] (which focused on solar cycle effects) and Calvo et al. [2007]. Of the studies that have used imposed QBOs, only Hamilton [1998] used a downward-propagating QBO.

\textsuperscript{13}Kodera [1993] modified the solar UV heating rates in their GCM by values ranging from 70\% to 110\%.

\textsuperscript{14}Matthes et al. [2004] give the observed difference in total solar irradiance (TSI) between solar cycle maxima and minima as 0.1\%, and the difference at UV wavelengths 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\%. Both the Matthes et al. [2004] and Baldwin et al. [2001] figures are smaller than the solar forcing magnitudes employed by Kodera et al. [1991].
1.4 Thesis outline

This thesis considers the behaviour of the HTE in two long (150-year) runs of the Canadian Middle Atmosphere Model (CMAM), a GCM. In one of the runs, a spontaneous QBO is forced by a combination of resolved and parameterized waves, while time-mean equatorial easterlies prevail in the other (control) run. Since no externally imposed sources of interannual variability are prescribed in the model runs (e.g. the solar cycle, or interannually varying SSTs), this study represents an attempt to quantify the QBO-vortex interaction that occurs as an internal mode of variability of an atmospheric GCM.

Only two previous GCM studies with spontaneously occurring QBOs [Niwano and Takahashi, 1998; Calvo et al., 2007] have considered the HTE as an internal mode of atmospheric variability, and both used substantially shorter runs (15 and 50 years, respectively) than are considered here. Like those studies, we wish to determine whether a HTE occurs in our model (CMAM), and whether its timing and magnitude are similar to that of the observed HTE. We also wish to determine, by comparing the QBO and control runs, whether the addition of a QBO to the model affects its climatological state. Finally, the length of our QBO run (150 years) allows us to ask whether decadal variability in the strength of the HTE occurs as an internal mode of variability in our model. It has been noted, as described in Sec. 1.2, that the strength of the observed HTE may perhaps vary on a decadal timescale, but it remains unclear whether this behaviour results from external forcings (such as the solar cycle) or from internal atmospheric variability.

This chapter has reviewed the relevant background from observational and modelling studies. Chapter 2 describes the QBO that occurs in the CMAM, with comparisons made to observations (ERA-40 and tropical radiosonde data). An analysis of the seasonality of QBO phase transitions, in both CMAM and in observations, is also made. In Chapter 3 we contrast the climate of the QBO and control runs and compare their differences to the observed contrast between QBO-W and QBO-E composites in ERA-40 (i.e. to the observed HTE). The forcing of these differences by
resolved and parameterized waves is considered in some detail. Chapter 4 considers
the timing and magnitude of the HTE within the QBO run, and examines the nature
of the decadal variability in HTE strength that occurs in the CMAM simulation as
well as in observations. Concluding remarks can be found in Chapter 5. Sec. 4.3.1
and some of Sec. 4.6.2 are published as Anstey and Shepherd [2008].

Before proceeding to the body of the thesis, the datasets to be examined are
described in Sec. 1.5.

1.5 Datasets

The CMAM data used in this thesis consists of two 150-year integrations, hereinafter
referred to as the “QBO” and “control” runs. For comparison of the CMAM results
with observations, the ERA-40 reanalysis [Uppala et al., 2005] is employed. The long
record of tropical radiosonde data compiled at the Freie Universität Berlin (commonly
referred to as the “Singapore winds”) is also used. Details of these datasets are given
below.

1.5.1 Canadian Middle Atmosphere Model (CMAM)

CMAM is a spectral GCM developed through a collaboration between Canadian
universities and the Canadian Centre for Climate Modelling and Analysis (CCCma)
of Environment Canada [Beagley et al., 1997; Scinocca et al., 2008]. The integrations
(“runs”) employed in this study use a horizontal T47 resolution and 98 vertical levels
extending from the surface up to 0.01 hPa, or \( z \approx 75 \) km. The vertical resolution varies
throughout the model domain: from the surface to \( z \approx 2.5 \) km there are 12 vertical
levels, for which the vertical level spacing increases (in the upward direction) from 0.1
km to 0.5 km; between \( z \approx 2.5 \) km and \( z \approx 32 \) km (10 hPa), the spacing is roughly
0.5 km; above \( z \approx 32 \) km, the spacing increases smoothly from 0.5 km to become 2.7
km near the model lid. (These values of \( z \) for the level spacings are approximate, as
the model employs a terrain-following vertical coordinate.) Sea surface temperatures
(SSTs) are prescribed from the climatology of AMIP-II observations by interpolating
the climatological monthly means to generate a smooth time series of daily values.
Hence the SSTs vary seasonally but possess no interannual variability. Other sources
of interannual variability, such as long-timescale radiative forcings due to increases
in greenhouse gas concentrations or 11-year solar cycle variations, are also absent.
Prognostic chemistry is not employed, and ozone concentrations are prescribed from
a seasonally varying climatology. Longwave radiation in the middle atmosphere is
parameterized using the scheme of Fomichev and Blanchet [1995]. The effects of
sub-gridscale gravity waves, both orographic [McFarlane and Scinocca, 2000] and
non-orographic [Scinocca, 2003], are parameterized.

The QBO in CMAM was obtained by Dr. J. F. Scinocca of CCCma. The vertical
resolution used in the two CMAM integrations is finer than that normally employed in
CMAM or in most other GCMs that resolve the middle atmosphere. It was increased
in order to obtain a QBO in the model [J. F. Scinocca, personal communication].
Similarly increased vertical resolutions have been employed in other GCMs in order
to obtain a QBO [Baldwin et al., 2001]. Increased vertical resolution alone, however,
proved insufficient to generate a stable QBO in CMAM. In addition, Scinocca found
it necessary to increase the magnitude of the source spectrum for the parameterized
nonorographic gravity waves by approximately a factor of three at tropical latitudes.
The increase was imposed only at tropical latitudes, tapering off smoothly so that
no enhancement of the source spectrum was imposed poleward of 40° in either hemi-
sphere. A QBO in CMAM resulted when both of these modifications – enhanced
vertical resolution and enhanced parameterized nonorographic gravity wave forcing –
were implemented [J. F. Scinocca, personal communication].

There is no obvious physical justification for the increased magnitude of the pa-
parameterized nonorographic gravity wave drag that catalyzes the QBO in CMAM. As
will be noted in Sec. 2.1, however, there are few observational constraints on the mag-
nitude of this parameterized forcing. Thus the source spectrum magnitude cannot
be justified on the basis of observations for either the unamplified case (the control
run) or for the amplified case (the QBO run). Given this ambiguity, we regard the
amplified source spectrum simply as an expedient that allows us to conduct a GCM
experiment involving two strongly contrasting cases of tropical stratospheric winds. Since it is not the goal of this study to examine the forcing of the QBO itself in any detail, this justification will be regarded as adequate for our purposes. It should be noted, however, that it is now generally accepted that some level of gravity-wave forcing is required to drive the real QBO [Dunkerton, 1997].

As stated above, this study employs two 150-year CMAM runs that are hereinafter referred to as the QBO and control runs. The only difference between the two runs is that the source spectrum for the parameterized nonorographic gravity wave forcing is amplified at tropical latitudes in the QBO run (as described in the two previous paragraphs, above). The vertical resolution, although finer than that usually employed in CMAM, is identical in the QBO and control runs.

1.5.2 ERA-40 reanalysis

ERA-40 is a reanalysis dataset spanning the 44-year period from Sep 1957-Aug 2002, and is used here to provide an observational dataset representing global-scale features of the stratospheric circulation. Reanalysis involves the combination, by means of data assimilation, of an atmospheric GCM with a diverse collection of meteorological observations. The observations are used to constrain the GCM to follow closely the evolution of the observed atmosphere, while the GCM extends the information from the observations by providing increased (i.e., global) spatial coverage in both the horizontal and vertical dimensions. Details of the data assimilation methodology used in ERA-40, as well as the spatial extent and timing of observations, are given by Uppala et al. [2005].

ERA-40 was produced by the European Centre for Medium-Range Weather Forecasts (ECMWF). Data assimilation is used routinely to produce analyses that are used for weather forecasting, and for this purpose the assimilation system is run at a very high resolution (e.g. T799 with 91 vertical levels up to 0.01 hPa, in a recent version [Lu et al., 2008]). The assimilation system naturally changes over time as improvements to it are made (e.g. increasing the horizontal resolution when increased computing power becomes available). The intention of the ERA-40 project,
in contrast, is to provide a dataset that is useful for studies of long-term trends and variability in the atmosphere [Uppala et al., 2005]. For this purpose it is desirable that the assimilation system be held in a fixed state, so that only the number and type of observations changes over time. This is done so that any long-term changes in climate that are seen in the reanalysis data may be ascribed either to real changes in climate or to changes in the number and type of observations (but not to a changing computational methodology). The horizontal and vertical resolution of the GCM used to produce the ERA-40 reanalysis is therefore fixed; it is T159, with 60 vertical levels extending from the surface up 0.1 hPa. Computational feasibility dictates that the resolution must be lower than that of the operational system, since 44 years of reanalysis data must be produced at a rate somewhat quicker than in real time.

Due to its high model lid (0.1 hPa), ERA-40 represents the whole stratosphere and a portion of the mesosphere. The upper stratosphere (5-2 hPa layer) suffers from a cold bias of $\approx 5$ K in the later part of the record, as well as a large cold bias of $\approx 10$ K during winter and spring in the Antarctic lower stratosphere during the early years of the record [Randel et al., 2004; Uppala et al., 2005]. Since we do not examine trends in this study, it is the quality of NH interannual variability in ERA-40 that is of primary importance here. Uppala et al. [2005] argue that due to extensive NH radiosonde coverage over the whole reanalysis period, ERA-40 gives an accurate representation of SSW events in both the pre-satellite and post-satellite eras. The ERA-40 representation of the QBO is also of high quality, at least up to 10 hPa, due to the assimilation of tropical radiosonde data [Randel et al., 2004; Baldwin and Gray, 2005].

In this study we employ all 44 years of the publicly available ERA-40 data, which was obtained from the ECMWF website (http://data.ecmwf.int/data/d/era40_daily/). This data is given on 23 pressure levels spanning the range from 1000 hPa up to 1 hPa, at $2.5^\circ \times 2.5^\circ$ resolution in latitude and longitude.

It should be noted that the NCEP/NCAR reanalysis [Kalnay et al., 1996; Kistler et al., 2001] also provides a dataset of length similar to that of ERA-40. For this study it is felt that ERA-40 is more appropriate for comparison with CMAM, due to
ERA-40 having a greater vertical extent (the NCEP model lid is at 10 hPa) as well as a better representation of the QBO.

### 1.5.3 Tropical radiosonde winds

Since 1953, radiosonde observations from near-equatorial stations have provided a valuable long-term record of the QBO winds. The data were compiled into a contiguous record at the Freie Universität Berlin, using observations from three stations: Canton Island (3°S, 172°W; Jan 1953-Aug 1967), Gan/Maledives (1°S, 73°E; Sep 1967-Dec 1975) and Singapore (1°N, 104°E; Jan 1976-present) [Naujokat, 1986]. The data are provided as monthly means at seven vertical levels: 70, 50, 40, 30, 20, 15, and 10 hPa. The 10 hPa data begin only in Jan 1956. The number of daily observations contributing to the monthly means decreases with increasing altitude due to the fact that the altitude reached by any given radiosonde ascent before the balloon bursts is variable. Although the observations exist only at particular longitudes, they are assumed to provide an adequate representation of the QBO on the basis that zonal asymmetries in the lower stratospheric flow are believed to be small in the monthly means, a view which is supported by GCM modelling results [Hamilton et al., 2004].

It should be noted that in various places in the thesis we alternate between usage of the ERA-40 equatorial winds and the tropical radiosonde winds when referring to the observed QBO. The two datasets give almost identical representations of the QBO at altitudes below 10 hPa, due to the fact that ERA-40 assimilates the radiosonde measurements [Baldwin and Gray, 2005].
Chapter 2

The QBO in CMAM

2.1 Introduction

In this chapter the QBO as obtained in CMAM is characterized. While a detailed analysis of the CMAM QBO is not the focus of this study, establishing its basic properties provides a necessary groundwork for examining the effects of the QBO on the extratropical circulation. Previous modelling work, discussed in Sec. 1.3, indicates that the strength and timing of the extratropical response depends on the type of QBO – or more generally, the type of tropical wind perturbation – that occurs in a model. Hence it is necessary to establish the properties of the CMAM QBO before proceeding to analyze the extratropical response.

We begin with a summary of some relevant QBO theory and describe the reasons why many GCMs do not accurately represent the QBO. According to the currently accepted theory [Lindzen and Holton, 1968; Holton and Lindzen, 1972], the QBO is driven by tropical waves that propagate upward from the troposphere. A variety of wave types are believed to be important, including planetary-scale Kelvin and Rossby-gravity waves as well as inertio-gravity waves and a continuous spectrum of smaller-scale gravity waves [Baldwin et al., 2001]. These waves are forced in the troposphere by convective heating.

Large-scale tropical waves are well resolved by GCMs such as CMAM, which typically have horizontal grid spacings of $O(100 \text{ km})$. However, the forcing of these waves
depends on the nature of the convective parametrization employed. Such parameterizations are often tuned to yield realistic mean precipitation values, but may not necessarily generate realistic amounts of upward-propagating tropical wave activity at resolved scales [Horinouchi et al., 2003]. Hence it is an open question whether the spectrum of resolved waves in many GCMs, including CMAM, provides the correct forcing for the QBO. Small-scale gravity waves that are not resolved at all in GCMs are also important for driving the QBO [Dunkerton, 1997]. The momentum transport due to these waves must be parameterized if their effects are to be taken into account. Unfortunately, there is a paucity of observational constraints on this process. Hence inaccurate representation of both the convective forcing of tropical waves as well as the momentum forcing by small-scale gravity waves may hinder the ability of a GCM to exhibit a QBO. Additionally, observations of tropical waves are not sufficiently extensive that the partitioning of QBO forcing by different wave types can be regarded as well understood [Baldwin et al., 2001]. This fact also impairs efforts to properly model the QBO in GCMs, since it is not clear what a realistic spectrum of waves that drive the QBO would look like.

However, uncertainty regarding the exact partitioning is not regarded as grounds to call the Lindzen-Holton theory of the QBO into question. This is because the theory’s basic mechanism is robust in the sense that it is able to incorporate a variety of wave types, so long as the waves propagate upward and are somehow induced to deposit momentum within the flow [Holton and Lindzen, 1972]. The key requirement is that there exist waves with both signs of zonal phase speed, $c$. The simplest possible case was described by Plumb [1977] and involves two waves with opposite values of $c$ and equal amplitudes, interacting with the mean flow due to radiative damping of the waves. One may imagine an initially stationary ($\bar{u} = 0$) stratospheric flow into which the two waves propagate. Radiative damping of each wave as it travels upward will induce, for each wave, a zonal force in the direction of $c$ that is inversely proportional to the intrinsic phase speed of the wave, $|\bar{u} - c|$. Since the two waves have equal amplitude and since $\bar{u} = 0$, the two oppositely-directed forces cancel and there is no acceleration of $\bar{u}$. This situation is, however, an unstable equilibrium, because any
imbalance between the wave-induced forces will induce a feedback that strengthens their imbalance. This feedback works as follows: given an imbalance between the two forces at an altitude \( z \), such that \( \partial \bar{u} / \partial t > 0 \) at \( z \), \( \bar{u} \) at \( z \) will increase and hence so will the positive force, since it is inversely proportional to \( |\bar{u} - c| \) (and \( c \), which is positive at \( z \), is unchanged). The negative force, which is due to the \( c < 0 \) wave, will decrease at \( z \). This implies a concurrent increase of the negative force at an altitude above \( z \), since a smaller fraction of the wave’s momentum flux is absorbed by the mean flow at \( z \). Hence a net negative force will yield \( \partial \bar{u} / \partial t < 0 \) at the higher altitude. Thus the overall effect on \( \bar{u} \) is that opposite-signed forces occur at different altitudes, forcing zonal jets of opposite signs, due to the filtering effect of \( \bar{u} \) on the waves. Because the waves propagate upward, the filtering effect will cause the resulting shear zones – i.e. regions of nonzero \( \partial \bar{u} / \partial z \) that separate the zonal jets – to descend in time. A notable consequence of this mechanism is that the period of the resulting oscillation depends only on the amount of wave forcing: stronger waves will deposit more momentum in the flow and cause the shear zones to descend more quickly.

In reality, a variety of tropical wave types – of varying phase speeds, amplitudes and wavelengths – exist and drive the QBO via the above mechanism. Waves that force the QBO will either have critical levels within the range of altitudes spanned by the QBO, or will have vertical group velocities that are sufficiently slow for radiative damping to cause them to deposit appreciable amounts of momentum in the QBO flow. The particular combination of waves that provide the forcing need not be the same for both phases of the oscillation. Kelvin waves are generally believed to contribute to the westerly forcing, Rossby-gravity waves to the easterly forcing, and inertia-gravity and smaller-scale gravity waves to both phases. Meridionally propagating planetary waves from the winter hemispheres may also contribute to the forcing of the easterly phase [Dunkerton, 1983]. Differences between the types of wave that contribute to each phase of the QBO should lead to asymmetries between the two QBO phases, and observations indicate that the two phases do indeed differ somewhat in their amplitude, duration and the rate at which they descend.

It is also expected that the sources of the waves that drive the QBO may vary in
The seasonal cycle in tropical convection has been shown to contribute significantly to the time variation of Kelvin wave and Rossby-gravity wave fluxes in the lower stratosphere, with the Rossby-gravity wave being particularly affected [Maruyama, 1991]. Interannual variations in SSTs (e.g., ENSO) may also be important [Geller et al., 1997]. Since the rate of QBO phase descent (and hence the QBO period) depends on the strength of the wave forcing, time variations in the sources of the waves will most likely create irregularities in the QBO from cycle to cycle, and possibly contribute to the observed asymmetries between the two phases. Because a GCM is able to represent all of the relevant wave types – albeit only by parametrization for the smaller-scale waves – a realistic GCM simulation of the QBO may be expected to display irregularities from cycle to cycle due to the variation in time of the wave sources. However, exactly how large should be the irregularities that are driven by variability of the wave sources is unknown.

Variations in wave forcing, however, are not the only mechanism that affects the descent of QBO phases; the effects of the Earth’s rotation and of the seasonal cycle are also important. Rotation introduces an asymmetry between the descent rates of easterly and westerly phases, due to the nonzero vertical velocity, $\bar{w}$, that is required to maintain thermal wind balance at the equator in the presence of nonzero vertical shear $\partial \bar{u} / \partial z$. Thermal wind balance holds at the equator according to [Andrews et al., 1987]

$$\frac{\partial \bar{u}}{\partial z} = -\frac{R}{H \beta a^2} \frac{1}{\partial \phi^2} \frac{\partial T}{\partial \phi}$$

(2.1)

where $R$ is the ideal gas constant for dry air, $H$ is the scale height, $a$ the radius of the Earth, and $\beta = a^{-1} df/d\phi$ where $f$ is the Coriolis parameter. Eq. 2.1 shows that equatorial shear anomalies, due to mechanical forcing by waves, must be associated with equatorial $\bar{T}$ anomalies if thermal wind balance is to be maintained. This is

### Footnote

1In the CMAM runs considered here, the source spectrum for the parameterized gravity waves does not vary in time. But the sources of resolved waves may vary. The imposed SSTs have no interannual variability, but the seasonal cycle in SSTs may contribute to time variations of tropical wave sources. There is also the possibility that internal atmospheric interannual variability of the tropical troposphere may affect the QBO, due to filtering of the wave fluxes entering the stratosphere.
accomplished by a mean meridional circulation [Eliassen, 1951], whereby \( \bar{w} \) forces departures from radiative equilibrium via adiabatic ascent or descent, thus inducing a \( \bar{T} \) anomaly that is in thermal wind balance with \( \partial \bar{u} / \partial z \). Since \( \partial \bar{u} / \partial z \neq 0 \), however, \( \bar{w} \neq 0 \) also creates a zonal momentum forcing due to vertical advection. The sign of the contribution is the sign of \( -\bar{w} \partial \bar{u} / \partial z \). Hence \( \bar{w} \) acts to speed up (slow down) the descent of the QBO-W (QBO-E) phase [Lindzen and Holton, 1968; Plumb and Bell, 1982]. This is consistent with the observed asymmetry of mean QBO phase descent rates [Baldwin et al., 2001]. Since GCM simulations respect thermal wind balance at the equator, this aspect of QBO behaviour should occur in a GCM simulation of the QBO.

A seasonally-varying factor that may contribute to the irregularity of QBO cycles and the asymmetry between phases is the Brewer-Dobson circulation (BDC) [Dunkerton, 1991, 1997]. The entire tropical stratosphere is slowly moving upward, and it has been argued that this upward motion is due to remote forcing by wave-mean flow interactions that occur in the extratropics during winter in either hemisphere, with the largest pull being exerted during NH winter due to the wave forcing being stronger in NH winter than in SH winter [Holton et al., 1995]. The additional advective contribution of the BDC to \( \partial \bar{u} / \partial t \) will always act to resist the descent of a shear zone since it advects wind of opposite sign to that of the descending QBO phase (i.e., it pulls upward instead of downward). Because the rate of BDC upwelling is of the same order as the rate of QBO phase descent (approximately 1 km/month), it is likely that the QBO is sensitive to the strength of the BDC. That is, if we suppose that there is a natural QBO descent rate that is determined by the strength of the total tropical wave forcing, then the QBO would be insensitive to BDC upwelling if the latter were an order of magnitude smaller than the QBO’s natural descent rate. Conversely, much stronger BDC upwelling (than the natural QBO descent rate) would likely prevent the QBO from occurring at all.

Interestingly, the QBO descent rate that is inferred from momentum fluxes due to observed Kelvin and Rossby-gravity waves alone is too weak to drive the QBO in the presence of the BDC upwelling. This argues for the importance of gravity waves
as additional contributors [Dunkerton, 1991]. Sensitivity to the strength of the BDC upwelling is a likely explanation for why stallings of the QBO-E descent occur predominantly during NH winter [Dunkerton and Delisi, 1985], although sometimes they begin during SH winter and persist through NH winter. While the BDC upwelling should act to slow the descent of both QBO phases, the descent of QBO-E phase is also inhibited by the advective forcing that is required to maintain thermal wind balance, as discussed earlier. The combined influence of the two effects may be sufficient to temporarily halt the descent of the easterlies until the BDC upwelling has subsided in NH spring. Hypothesizing the BDC as the cause of QBO-E stallings also suggests a reason why not all QBO-E phases undergo stalling: due to the large interannual variability of NH winter dynamics, the strength of the BDC upwelling will vary from year to year. Though the exact spatial structure of tropical upwelling is still open to question, results from modelling studies [Kinnersley and Pawson, 1996; Hampson and Haynes, 2004] lend support to the hypothesis that the BDC strongly affects the descent of QBO phases.

A successful GCM simulation of the QBO requires sufficient amounts – and, hopefully, the correct types – of tropical wave forcings. A GCM can also be expected to represent seasonal effects such as the variability of tropical wave sources and the BDC upwelling. Just as the QBO appears to influence the extratropical atmosphere both in observations and in models, the extratropics may also influence the QBO. The influence of the extratropics on the QBO may be manifested in the irregularity of QBO cycles as well as in their modulation by the seasonal cycle, although these might be considered secondary effects that act as perturbations to the overall, robust, QBO mechanism. We now consider how these influences are manifest in the CMAM QBO, comparing the model to observations as appropriate.

2.2 Characteristics of the CMAM QBO

Fig. 2.1 shows the complete 150-year time series of the CMAM QBO. For comparison, Fig. 2.2 shows the observed QBO as represented by the tropical radiosonde winds
Figure 2.1: Monthly average $\bar{u}$, $2^\circ$S-$2^\circ$N average, for the 150-year CMAM QBO run. Contour spacing is 5 m/s, red (blue) indicates westerlies (easterlies), and the $\bar{u}=0$ line is shown in black.
2. The QBO in CMAM

Figure 2.2: The QBO in the real atmosphere, as represented by monthly average tropical radiosonde winds for 1953-2004 (data at 10 hPa begins in 1956). Contours and colours as in Fig. 2.1.

(Se. 1.5.3). Two features stand out in Fig. 2.1 as being unrealistic compared with the observed QBO. First, the CMAM QBO has an unrealistically long period, averaging roughly 35 months, whereas the period of the observed QBO averages roughly 28 months. (See Figs. 2.10 and 2.11 for more precise values of the observed and modelled QBO periods.) Second, there is a large westerly bias at the lowermost altitudes reached by the CMAM QBO. Easterly phases generally do not penetrate below about 23 km (roughly, 40 hPa), and hence the wind between 18 km and 23 km (80 hPa to 40 hPa) is almost always westerly. There is nevertheless an oscillation at these altitudes, albeit between weak and strong westerlies rather than between easterlies and westerlies; this is seen in Fig. 2.3, which shows deseasonalized $\bar{u}$ for a 10-year excerpt from the CMAM QBO run. (The deseasonalized $\bar{u}$ is defined by subtracting the climatological annual cycle of the model run.)

A somewhat realistic feature of the CMAM QBO, apparent in both Figs. 2.1 and 2.3, is the irregularity of the oscillation from one cycle to the next. That is, QBO cycles in CMAM differ from one another in their amplitude, duration and timing. The observed QBO also displays substantial variability from cycle to cycle (Fig. 2.2). Such variability may be due to a number of factors, as discussed in Sec. 2.1, and will
be considered further in Sec. 2.3, below.

Fig. 2.3 also shows the full vertical range of CMAM. It is seen that coherent QBO phases appear to initiate somewhere in the 35-50 km region, although downward-propagating $\bar{u}$ anomalies – characteristic of interactions between $\bar{u}$ and tropical waves, as described in Sec. 2.1 – are apparent at higher altitudes as well. An observed QBO signal has been shown to occur as high as the stratopause [Pascoe et al., 2005], while a mesospheric QBO also appears to exist [Baldwin et al., 2001]. Because the version of CMAM used in this study does not resolve the whole mesosphere, we do not here address the existence of QBO signals above the stratopause. Such mesospheric QBO signals as might occur in the model are likely to be highly unrealistic, due to the truncated mesosphere and the proximity of the nonzonal sponge layer (the influence of which may be felt as far as two scale heights from the model lid).

It is obvious from Figs. 2.1, 2.2 and 2.3 that the observed QBO and the CMAM QBO both constitute large variability about the time-mean state of the tropical winds. This behaviour may be contrasted with variability of tropical $\bar{u}$ in the CMAM control
Figure 2.4: Monthly average $\bar{u}$, 2°S-2°N average, for the 150-year CMAM control run. Contours and colours as in Fig. 2.1.
2. The QBO in CMAM

run, shown in Fig. 2.4. The control run predominantly exhibits easterlies in the lower equatorial stratosphere, which may provide a very different “boundary condition” for the extratropical flow than is the case when a QBO occurs.

Fig. 2.4 shows that forcing of \( \bar{u} \) by resolved tropical waves and non-enhanced parameterized nonorographic gravity waves in CMAM is insufficient to establish a QBO. It may be noted in passing, however, that in the upper reaches of the range of altitudes spanned by the CMAM QBO, Fig. 2.4 shows that the control run exhibits westerly wind regimes that propagate erratically downward on a very long timescale. This behaviour is reminiscent of that seen by Hamilton et al. [2001] in perpetual-equinox GCM integrations in which a QBO-like oscillation first appeared and then mysteriously stopped: instead of a regular oscillation, their simulation exhibited vertically stacked (though erratically varying) jets of alternating sign in the tropics\(^2\). The CMAM control run appears to behave similarly, but more erratically. The irregularly descending westerly regimes may be forced by Kelvin waves or gravity waves (resolved or parameterized) with westerly phase speeds that are, on their own, of insufficient strength to induce the westerlies to propagate below a certain altitude. Thus Fig. 2.4 makes it apparent that in CMAM, an additional source of wave momentum flux is required to induce a QBO. This is provided in the CMAM QBO run by the increased parameterized nonorographic gravity wave drag (Sec. 1.5.1).

The variability of tropical \( \bar{u} \) in both the QBO and control runs (Figs. 2.1 and 2.4) may be contrasted with the climatological tropical \( \bar{u} \) of both runs, and the CMAM climatologies may also be contrasted with the climatologies of observed tropical winds from ERA-40 and the tropical radiosonde winds. The four climatologies are shown in Fig. 2.5. The enhancement of the parameterized nonorographic gravity wave drag in CMAM (Sec. 1.5.1) changes the lower stratospheric climatological winds from weak easterlies of \( \approx 2 \) m/s in the control run (Fig. 2.5(b)) to westerlies of \( \approx 12 \) m/s in the QBO run (Fig. 2.5(a)). Substantial changes in the semianual oscillation (SAO)

\(^2\)Hamilton et al. [2001] used the SKYHI model with 80 vertical levels between the surface and 0.01 hPa, i.e. a vertical resolution similar to that of CMAM (Sec. 1.5.1). Unlike CMAM, no parameterized subgrid-scale gravity wave drag was used in SKYHI.
are seen also to occur, particularly in the enhanced downward penetration of the first SAO-W phase of the calendar year. In comparison to the ERA-40 climatology (Fig. 2.5(c)), the CMAM QBO run (Fig. 2.5(a)) seems to have an incorrect asymmetry between SAO-E phases, with the SH winter phase being stronger than the NH winter phase. However, as ERA-40 assimilates no equatorial wind observations above 10 hPa, it is not obvious that ERA-40 exhibits realistic SAO phases either. In the troposphere, climatological winds in the QBO and control runs appear identical, consistent with the fact that the only difference between the two runs is the enhancement
Figure 2.6: Composites of deseasonalized $\bar{u}$ lagged with respect to the initiation of QBO phases at 20 hPa: (a) ERA-40 QBO-W, (b) ERA-40 QBO-E, (c) CMAM QBO-W, (d) CMAM QBO-E. Contour interval is 5 m/s. Colours as in Fig. 2.1.

A composite picture of the evolution of QBO winds in both CMAM and ERA-40 is shown in Fig. 2.6. Lag zero in these composites corresponds to the time of QBO phase onset at 20 hPa ($z \approx 27$ km). Deseasonalized winds are composited, and hence the westerly bias of the CMAM QBO is not apparent in these plots. In comparison to ERA-40, the CMAM QBO is seen to have realistic amplitude, but a somewhat reduced vertical extent both above and below the altitudes at which the oscillation is most prominent. The slower average descent rate of the QBO-E phase relative to the QBO-W phase (which was discussed in Sec. 2.1) is apparent in both ERA-40 and in...
Fig. 2.7 shows zonal cross sections of the QBO in both ERA-40 and CMAM (corresponding to the lag zero time in Fig. 2.6), in which the reduced vertical extent of the CMAM QBO is again apparent. Fig. 2.7 also shows that the CMAM QBO has unrealistically narrow latitudinal width in the lowermost stratosphere. While the half width (i.e., latitude of half-maximum $\bar{u}$) of the ERA-40 QBO is roughly $12^\circ$ for the lower jet in both ERA-40 composites (Fig. 2.7(a,b)), the half widths in CMAM are $8^\circ$ and $7^\circ$ for the lower QBO-E and QBO-W jets in Fig. 2.7(c,d), respectively. This difference of nearly a factor of two may affect the influence of the QBO on the extratropics if the off-equatorial winds in the lowermost tropical stratosphere are important. The latitudinal width of the CMAM QBO at upper levels is realistic.
Finally, Figs. 2.6 and 2.7 indicate that the CMAM QBO has realistic amplitude.

### 2.3 Seasonality of QBO phase transitions

It has been known for some time that the period of the QBO varies on decadal timescales and that the timing of phase transitions is not independent of the seasonal cycle, but rather displays a weak seasonality [Dunkerton and Delisi, 1985]. One realistic aspect of the CMAM QBO is that the timing of its phase transitions displays a seasonal dependence not unlike that seen in observations.

The seasonality of phase transitions for the observed QBO, represented by the 1953-2004 deseasonalized tropical radiosonde winds (Sec. 1.5.3), is shown in Fig. 2.8.
Figure 2.9: As in Fig. 2.8, but for the CMAM QBO ($2^\circ$S-$2^\circ$N averaged $\bar{u}$) at the model vertical levels corresponding most closely to the tropical radiosonde levels. Unlike in Fig. 2.8, the ordinate spans 150 years (the full length of the CMAM QBO run), rather than 52 years (the length of the observed record).
for six of the seven vertical levels available (70 hPa is excluded because it is extremely noisy). We use the radiosonde winds rather than ERA-40 so as to consider a slightly longer record. (As noted in Sec. 1.5.3, there is excellent correspondence between the two datasets over the altitude range spanned by the radiosonde data.) The seasonality of QBO phase transitions in CMAM, at the six pressure levels closest to those used in Fig. 2.8, is shown in Fig. 2.9. Both the CMAM and radiosonde winds are first deseasonalized and then smoothed with a 5-month running mean before the months in which phase transitions occur are determined. Neither of these filterings substantially affects the patterns of behaviour seen in Figs. 2.8 and 2.9. The purpose of the 5-month running mean is simply to prevent transient equatorial wind fluctuations on timescales of less than ≈ 5 months being misclassified as persistent QBO phases.

The histograms shown in Fig. 2.8 (upper panels) update Fig. 4 of Dunkerton [1990]. (See also Pascoe et al. [2005], Fig. 7.) They indicate a tendency for QBO phase initiations at all levels to cluster seasonally, with different levels favouring different seasons. The clustering appears to be strongest at the lowest altitudes (40 and 50 hPa). The histograms in Fig. 2.9 (upper panels) indicate that similar behaviour occurs in CMAM. The modulation by the seasonal cycle of QBO phase transition timing is a result of the CMAM QBO being spontaneously forced by resolved and parameterized waves, rather than being forced by an imposed zonal momentum source with a specified period (e.g., as was used in Hamilton [1998] and in other modelling studies).

The lower panels in Figs. 2.8 and 2.9 show the same data as the histograms in the upper panels, but in a form that reveals the temporal variation of QBO phase transitions. These plots show, for both radiosonde winds and the CMAM QBO, that some parts of the time series seem to favour the seasonal clustering of phase transitions more than others. Hence the clustering of phase transitions appears to vary on a decadal timescale.

The somewhat erratic variation of phase transition seasonality seen in the lower panels of Figs. 2.8 and 2.9 implies contemporaneous variations in the period of the
2. The QBO in CMAM

Figure 2.10: Durations of QBO phases for the deseasonalized tropical radiosonde winds. Durations correspond simply to the elapsed time intervals between the events shown in Fig. 2.8. Red (blue) indicates W (E) QBO phase durations. Upper panels show the probability distribution of individual phase durations; the numbers in the upper right of these panels give (top) the mean and (bottom) the standard deviation of the phase durations, in months. Lower panels show the temporal variation of QBO phase durations. Dots indicate the duration of individual phases, while crosses represent the duration of whole QBO cycles counting either from the beginning of W phases (red crosses) or E phases (blue crosses).

QBO\(^3\). Probability distributions for the durations of W and E QBO phases are shown in the upper panels of Figs. 2.10 and 2.11 for radiosonde winds and the CMAM QBO, respectively. The lower panels show the temporal variation of W and E phase durations, as well as the temporal variation of the duration of complete QBO cycles.

\(^3\)Note that it is hypothetically possible for QBO phase transitions to favour certain times during the seasonal cycle and still give rise to a QBO with constant period. For example, suppose that the QBO had a fixed period of 28 months, and that a QBO-W initiation occurred in Mar. Then the QBO-W initiations for the next three cycles of the oscillation would occur in Jul, Nov, and then Mar again. However, erratic variations in the seasonality of phase transitions imply a variable QBO period.
Figure 2.11: As in Fig. 2.10, but for CMAM QBO phase durations.
(i.e. the QBO period, as measured for each individual cycle of the oscillation). These plots confirm that the QBO period in both datasets does indeed change over time. The fact that this behaviour occurs in CMAM demonstrates that a variable QBO period may result from the internal dynamics of the atmosphere.

In Chap. 4, we will show that variations in the seasonality of QBO phase transitions have a discernible impact on the strength of the interaction between the QBO and the NH winter stratospheric polar vortex, in both observations and in CMAM. Motivated by this fact, as well as by a desire to better understand the internal atmospheric variability that occurs in CMAM, we now digress briefly to consider the underlying causes of the changing seasonality of QBO phase transitions.

\section*{2.3.1 Mechanisms for interaction between the QBO and the seasonal cycle}

It is not known with certainty what causes the seasonal modulation of QBO phase transition timing shown in Figs. 2.8 and 2.9. At the higher altitudes (e.g. 10 hPa) it is likely that the SAO plays a role in initiating QBO phases [Dunkerton and Delisi, 1985], because for waves undergoing critical level filtering, a source of shear is required to initiate the downward descent of a shear zone [Lindzen and Holton, 1968; Campbell and Shepherd, 2005]. For waves undergoing radiative damping, however, this is not required [Holton and Lindzen, 1972]. Since the exact nature of the combination of waves contributing to the observed QBO remains an open question (Sec. 2.1), the importance of the SAO in synchronizing QBO phase initiations with the annual cycle is unclear. What is clear, however, is that other effects must also contribute to variations in the rate of descent of QBO phase at altitudes below those directly influenced by the SAO, because if this were not the case then the QBO period would always be some multiple of six months - which Figs. 2.10 and 2.11 demonstrate is not the case (although, interestingly, this does occur in some GCMs).

One possible mechanism for the seasonality of QBO phase transitions is the BDC, as described in Sec. 2.1. This would manifest itself as a tendency for QBO-E
phase initiations to avoid occurring during winter, particularly NH winter, as is observed [Dunkerton and Delisi, 1985]. Fig. 2.8 shows, however, that E phase initiations do occasionally occur during either NH or SH winter. But if the BDC is the correct mechanism then we should not expect perfect avoidance, because the large interannual variability of NH winters may result in tropical upwelling that is stronger in some years and weaker in others, which would impede QBO phase descent to varying degrees.\(^4\)

A potential problem with the BDC mechanism is that it may apply less well to QBO-W phase initiations. Fig. 2.8 shows that these also display a seasonal preference, and yet they should be less susceptible to BDC-induced stalling (Sec. 2.1). However, given a seasonal preference for the initiation of the E phase, we may expect a similar tendency to be manifest in W phase transitions, due to the knock-on effect of W initiations lagging the timing of E initiations [Dunkerton, 1990]. The coherence of the knock-on effect will depend on the variability of E phase durations. Since the E phase at lower QBO levels (40 hPa, 50 hPa) typically lasts one year, with a standard deviation of about 1-2 months (making it less variable than W phase durations at these levels, as shown in Fig. 2.10), the seasonality of W phase initiations at 50 hPa seen in Fig. 2.8 is consistent with its being due to this knock-on effect. At higher levels, the E phase duration becomes as variable as the W phase duration and the knock-on effect will become smeared out; this is also consistent with the behaviour seen in Fig. 2.8. Thus the BDC upwelling need only exert a significant influence on the descent of the E phase in order to induce a seasonal clustering of phase transitions by both phases.

However, the BDC is not the only factor that may cause variations in the rate of QBO phase descent. Variations in convective heating in the tropics, either seasonal or interannual (e.g. due to ENSO), may affect the magnitude of the waves that drive the QBO. Observed increases in Kelvin and Rossby-gravity wave activity during April

\(^4\)One might note that the tendency for QBO-E transitions to avoid SH winter is a little cleaner in Fig. 2.8, as not a single E initiation occurs in Aug-Sep at 20, 30 and 40 hPa. This may be consistent with the fact that the SH winter exhibits less interannual variability than the NH winter.
have been argued to result from the annual cycle in tropical SSTs [Maruyama, 1991]. In a two-dimensional QBO model, annual modulation of the QBO wave forcing has been shown to cause a seasonality in 50 hPa QBO phase transitions similar to that observed, although a similar effect was achieved in the same model by modulating the strength of BDC upwelling [Dunkerton, 1990]. In fact, virtually any effect that leads to an annual or semianual variation in the QBO phase descent rate may impart a seasonal preference to the timing of QBO phase initiations. It is therefore difficult to establish which of the above mechanisms – BDC upwelling, or variations in the sources of the waves that drive the QBO – is most important, or if both are required, without modelling studies.

We do not undertake detailed modelling studies of the QBO itself in this thesis. However, we may still take a diagnostic approach to clarifying the role of seasonal variation in the QBO phase descent rate. We denote the QBO vertical phase velocity by $w_{ph}$; hence the QBO phase descent rate is $-w_{ph}$. Figs. 2.12 and 2.13 show the seasonal cycle of $w_{ph}$ diagnosed from observations (radiosonde winds) and the CMAM QBO, respectively. $w_{ph}$ was calculated as the time derivative of the altitude of the $u = 0$ line ($\bar{u}$ for CMAM) of the deseasonalized winds. Unlike in Figs. 2.8, 2.9, 2.10 and 2.11, no smoothing was applied to the time series of the winds. $w_{ph}$ values have been grouped into different altitude bins, indicated in Figs. 2.12 and 2.13, showing that the seasonal modulation of $w_{ph}$ varies with altitude. Both annual and semianual variations are present to varying degrees at different altitudes. For observed QBO-W and E phases, and for CMAM QBO-W phases, the prominence of the semianual variation appears to increase at the higher altitudes. $w_{ph}$ also shows interannual variability, which is stronger in observations than in CMAM. $w_{ph}$ for observed QBO-E phases shows the most variability (Fig. 2.12), consistent with the occurrence of stallings.

The vertical variation of the semianual and annual components of $w_{ph}$ gives a clue as to which of the aforementioned mechanisms – the BDC, or wave source variations – is more important for determining the seasonality of QBO phase transitions. The fact that semianual variation appears more prominently at upper QBO levels (around 10
2. The QBO in CMAM

Figure 2.12: QBO vertical phase velocity $w_{ph}$ (km/month) of tropical radiosonde winds, in three altitude bins. Red (blue) corresponds to westerly (easterly) initiations. Solid line is the mean $w_{ph}$, dashed lines are $+/−1$ standard deviation and the dots show individual data points.

hPa), while annual variation is relatively more prominent at lower levels (around 50 hPa) suggests that the BDC may contribute more at the upper levels, while annual variation of wave fluxes entering the stratosphere may be more important at the lower levels. However, annual variations at the lower levels may also be consistent with BDC causality, since it has been argued that the annual cycle in $\bar{T}$ at these altitudes is due to interhemispheric asymmetry in the strength of extratropical winter wave drag causing an annual cycle in tropical upwelling [Yulaeva et al., 1994; Rosenlof, 1995].

Another aspect of the seasonal cycle of $w_{ph}$ that is apparent in Fig. 2.12, and which has implications for the seasonal clustering of QBO phase transitions, is that
Figure 2.13: As in Fig. 2.12 but for CMAM, in six altitude bins. It is possible to show more altitude bins than in Fig. 2.12 because of the higher vertical resolution of CMAM. Colours and lines as in Fig. 2.12.
during April-June the observed QBO-E phase is seen to descend faster, on average, than the QBO-W phase. This is the opposite sense of the asymmetry in $w_{ph}$ that is expected due to the QBO mean meridional circulation (Sec. 2.1). The BDC effect is unable to explain this feature, since upwelling should slow the descent of both phases equally. But it may be consistent with an increase in Rossby-gravity wave forcing due to the seasonal cycle of tropical convection, provided that this effect modulates the Rossby-gravity wave activity more than the Kelvin wave activity, as suggested by observations [Maruyama, 1991]$^5$. This view was argued by Wallace et al. [1993], who defined a rate of QBO phase progression $\partial \psi / \partial t$, with $\psi$ being a measure of the QBO state, based on an EOF analysis of the tropical radiosonde winds$^6$. Wallace et al. [1993] showed that the strongest values of $\partial \psi / \partial t$ occurred during spring and early summer in the NH, and noted further that descent of the QBO-E phase during this time appeared to be associated with slightly stronger values of $\partial \psi / \partial t$ than were found to occur during the descent of the QBO-W phase. Fig. 2.12 shows that the diagnosed phase descent rates for both the W and E observed QBO phases are consistent with Wallace et al. [1993]'s results$^7$.

As noted above, in a diagnostic context it is not possible to resolve the question of what dominant factor causes the behaviour shown in Figs. 2.12 and 2.13. We may, however, verify that the diagnosed $w_{ph}$ values are consistent with the seasonal clustering of QBO phase transitions shown in Figs. 2.8 and 2.9. Consider the case of QBO-E phases that initiate at some time between May and August at 10 hPa (28-33 km). Fig. 2.12 shows that climatological $w_{ph}$ for QBO-E (i.e., the solid line in Fig. 2.12) is decreasing during May-Aug in the 28-33 km bin. This will make phases that initiate earlier at 10 hPa (e.g., in May) arrive more quickly at lower levels than

$^5$Note that a seasonal modulation of the relative strengths of the E (e.g. Rossby-gravity wave) and W (e.g. Kelvin wave) forcing is required to explain this feature, since it is only during a part of the year that the QBO-E phase descends faster than the QBO-W phase. During the rest of the year, the QBO-W phase descends faster (Fig. 2.12).

$^6$We employ Wallace et al. [1993]'s representation of the QBO in Sec. 4.4 for the purpose of diagnosing the Holton-Tan effect; details of the definition of $\psi$ may be found in Sec. 4.4.1.

$^7$It should be noted that vertical variations in $w_{ph}$ are obscured by Wallace et al. [1993]'s method, but since Fig. 2.12 shows that increased $w_{ph}$ during April-June occurs for QBO-E over the whole altitude range of the radiosonde data, this fact is unlikely to affect their conclusion.
phases that initiate later at 10 hPa (e.g., in August), assuming that these phases propagate at the climatological \( w_{ph} \) (interannual variability in \( w_{ph} \) can be ignored for the moment so as to illustrate the effect of the seasonal cycle in \( w_{ph} \)). Hence an ensemble of QBO phases that initiate at 10 hPa between May and August will have arrival times at lower levels that are more spread out in time than were their initiations at 10 hPa. Conversely, an ensemble of QBO phases that initiate during a time of year when the climatological \( w_{ph} \) increases with time (e.g., Jan-Apr or Aug-Oct for the E phase at 28-33 km in Fig. 2.12) will tend to cluster in time when they arrive at the next lowest level. Because Fig. 2.12 shows that seasonal variation of \( w_{ph} \) occurs at all altitudes, descending phases will propagate through regions of alternately enhanced and reduced \( w_{ph} \).

The above argument can be made more precise. Consider a QBO phase that initiates at time \( t \) at 10 hPa. Having diagnosed climatological \( w_{ph} \), the altitude \( z(t) \) of the \( \bar{u} = 0 \) line can be found simply by integrating \( dz/dt = w_{ph}(\tau, z(t)) \) in time, where \( \tau \) denotes the month (i.e., \( \tau \) is \( t \) modulo one year). That is, the position of the \( \bar{u} = 0 \) line is advected as if it were a particle in a flow. We consider a collection of 12 phases, starting in each of the calendar months at 10 hPa, and use the CMAM
2. The QBO in CMAM

QBO-E $w_{ph}$ values (Fig. 2.13), which are chosen because the long CMAM time series (150 years), high vertical resolution (0.5 km) and fairly low noise (i.e., effects of interannual variability) makes the vertical and seasonal variation of these $w_{ph}$ values better defined than the observed values (Fig. 2.12), and hence allow the calculation to be done with the least possible ambiguity.

The result of the calculation is shown in Fig. 2.14. The lines in this figure show the altitude of $\bar{u} = 0$, which indicates the initiation of a QBO phase, as a function of time. Fig. 2.14 illustrates that a collection of phases at 10 hPa that are initially spaced evenly in time\(^8\) will exhibit seasonal clustering by the time they reach the lower levels, and that this leads to a tendency for E phase initiations at the lower levels to occur predominantly during NH spring/early summer. Similar calculations for the CMAM westerly phase, as well as for observations, lead to qualitatively similar results, but the higher noise levels in observations make the result more ambiguous, as $w_{ph}$ values are less well estimated in those cases. In all of the calculations, however, the tendency for phase initiations at the lower levels to avoid NH winter is fairly pronounced (although more so for the QBO-E phase than for the QBO-W phase, as expected). Hence Fig. 2.14 suggests that a tendency for seasonal synchronization of QBO phase initiations at lower levels may occur solely as a result of the seasonal cycle in $w_{ph}$.

In both observations (Fig. 2.8) and CMAM (Fig. 2.9), the timing of QBO phase initiations at 10 hPa are not spread evenly throughout the year, and Figs. 2.12 and 2.13 show that $w_{ph}$ exhibits interannual variability. The effect of these two factors can be estimated by applying a small amount of random noise to $w_{ph}$ (so as to mimic the interannual variability) and by using a realistic distribution of upper level phase initiations (rather than the uniform distribution employed in Fig. 2.14). Repeating the above calculation under these assumptions yields histograms of phase initiations.

\(^8\)Note that although the phases initiating during JFM at the top level in Fig. 2.14 appear to be clustering at that time, they are in fact initiating at evenly spaced times. Their clustering near the top level is due to the QBO-E phase descent rate being slow during Jan and Feb in the 30-32 km layer (Fig. 2.13). Also note that the top level, i.e. the level at which phases initiate, is actually set as 31 km for this calculation, since 31 km is the middle of the 30-32 km layer for which the top-level phase descent rate is defined (Fig. 2.13).
at 50 hPa that resemble the actual data (not shown). Thus while a seasonal preference for phase initiations at the upper levels may contribute to the seasonality of initiations at lower levels, it is not required, and it does not destroy the effect of the seasonal cycle in $w_{ph}$.

It is interesting to compare the Fig. 2.14 result with results from Hampson and Haynes [2004], who used a mechanistic model to study the effect of the BDC on the seasonality of QBO phase transitions. A QBO was forced in their model by parameterized tropical waves of a specified strength $F_0$, and the strength of extratropical wave forcing was specified by controlling the amplitude of a geopotential height perturbation, $\Phi_0$, at the model’s lower boundary in the winter extratropics. This yielded a two-parameter ($F_0, \Phi_0$) system in which the effect of the extratropical wave forcing on the period of the QBO could be explored across a range of parameter space. It was found that nonzero $\Phi_0$ induced a seasonal synchronization of QBO phase transitions similar to that seen in observations. Moreover, an attempt was made to separate the effects of the seasonal cycle of tropical upwelling from the interannual variability by setting $\Phi_0 = 0$ and instead applying the model’s climatological extratropical wave forcing repeatedly in each year of the integration. In this case, seasonal clustering of phase initiations still resulted (and was in fact slightly stronger than in the original, interannually varying case), consistent with the notion that the seasonal cycle in tropical upwelling is sufficient to induce this behaviour.

That the seasonal cycle alone seems able to produce seasonal clustering of QBO phase initiations suggests that this behaviour is likely to persist in the real atmosphere, just as it does in the 150-year CMAM run. This suggests that the observed seasonal clustering of phase transitions is not an artefact of the short (roughly 50-year) observational record.

Of course, consistency between $w_{ph}$ and the timing of phase initiations is guaranteed; in itself, it tells us nothing about the mechanisms that are responsible for this behaviour (at least, not beyond the clues already suggested earlier). But Figs. 2.10 and 2.11 show that decadal variations in QBO period are associated with the seasonal clustering of phase initiations, which we have argued may be explained to first or-
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It may seem trivial to point out that a seasonal clustering is due to the seasonal cycle\textsuperscript{9}. However, the point is that interannual variability seems not to be large enough to obscure the seasonal effect. If it were large enough to obscure the seasonal effect, then the observed seasonal clustering would simply be due to luck, and phase initiations would tend towards a more uniform (i.e., aseasonal) distribution over time. The persistence of seasonal clustering in the 150-year CMAM QBO run (containing 50 QBO cycles) suggests that this is not the case.

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Chapter 3

Comparison of the CMAM QBO and control runs

3.1 Introduction

As discussed in Sec. 2.1, most GCMs that represent the middle atmosphere are unable to resolve a realistic spectrum of vertically propagating tropical waves and hence are unable to simulate the QBO and sometimes also the SAO. The absence of these tropical wind oscillations means that at no time are the simulated tropical winds in a realistic state, in spite of the fact that their climatological state may bear some resemblance to the observed climatology (cf. panels (b) and (c) of Fig. 2.5). This deficiency may be expected to impact the extratropical circulation, due to the fact that a correlation between low and high latitude circulations is observed, as discussed in Sec. 1.2.

As stated in Secs. 1.1 and 1.4, this thesis considers the effects on the extratropics of changes in the tropical stratospheric winds in CMAM. In this chapter, differences between the composites of the CMAM QBO and control runs (hereinafter referred to as QBO-control differences) are the main tool used to assess the extratropical changes. ERA-40 differences between composite mean QBO-W and QBO-E states (hereinafter referred to as W-E differences) are provided for comparison with the QBO-control differences. This comparison should not be expected to yield exact correspondence
3. Comparison of the CMAM QBO and control runs

for at least one reason – namely, that the CMAM QBO-control tropical $\bar{u}$ difference does not precisely resemble the W-E tropical $\bar{u}$ difference. However, both difference patterns (QBO-control and W-E) do exhibit strong westerly differences in the lowermost tropical stratosphere. Hence any similarities between the two extratropical responses may be taken to suggest that tropical lower stratospheric winds exert an important influence over the extratropics.

For the rest of this chapter, tropical $\bar{u}$ differences will be denoted by the symbol $\Delta \bar{u}_{EQ}$ (as in Chap. 1), and for brevity’s sake we will adopt the language that extratropical QBO-control or W-E differences represent a “response” to $\Delta \bar{u}_{EQ}$. Our choice of the symbol $\Delta \bar{u}_{EQ}$ carries the connotation that the extratropical response is caused by differences in the time-mean wind conditions that prevail in the tropical stratosphere. It should be borne in mind, however, that the extratropical response to $\Delta \bar{u}_{EQ}$ could occur for a variety of reasons. For example, the QBO-control $\Delta \bar{u}_{EQ}$ is westerly in the lowermost stratosphere, and it was shown in Sec. 2.2 that easterlies virtually never occur in the QBO run at these altitudes; a simple interpretation of these facts is that the extratropical QBO-control response is due to the occurrence of time-mean westerlies in the lowermost tropical stratosphere, as suggested above (and also in the previous paragraph). However, this is not the only possible interpretation of the response. It is also conceivable that $\Delta \bar{u}_{EQ}$ at other altitudes (e.g., closer to the stratopause) are causal for the extratropical response, or that the oscillatory nature of the QBO is important (i.e., our QBO run is not the same as if we had simply imposed a time-invariant westerly $\Delta \bar{u}_{EQ}$ in the lowermost tropical stratosphere).

It should also be borne in mind that the QBO-control difference characterizes the difference in mean climates between two model runs, whereas the W-E difference characterizes interannual variability. The fact that observations also contain the effects of secular trends and the decadal variability associated with the solar cycle and volcanic eruptions may also be important. Nevertheless, since nature is unable to provide us with observations of an atmosphere in which the QBO is absent, the observed W-E differences are taken here as the best available point of departure.
3.2 Zonal wind and temperature composites

Seasonal composites of zonal average zonal wind $\bar{u}$ are shown in Fig. 3.1 for the CMAM QBO run, CMAM control run, and ERA-40 reanalysis. Each CMAM composite is an average of 149 years, while the ERA-40 composites use all 44 available years (1958-2001). Fig. 3.1 shows that the CMAM climatology generally compares well to ERA-40 in the troposphere and stratosphere. The comparison cannot be made in the mesosphere because ERA-40’s highest vertical level is at the stratopause. As noted in Sec. 1.5.1, the lid for this particular version of CMAM is at $z \approx 75$ km, so only a portion of the mesosphere is resolved; additionally, the influence of the nonzonal sponge layer may extend as far as roughly two pressure scale heights ($\approx 14$ km) from the lid and hence may impact the realism of the simulated mesosphere. In this study we focus on the stratosphere.

As already noted in Sec. 2.2, the tropical stratospheric winds in Fig. 3.1 differ between the three climatologies. In the lowermost tropical stratosphere, the CMAM control run is more similar to ERA-40 than is the CMAM QBO run. The QBO run climatology shows a westerly bias of $\approx 10$ m/s in this region, due to the westerly bias of the QBO there (Sec. 2.2). At higher altitudes in the tropical stratosphere it is less clear which CMAM run has a more realistic climatology, but (as already noted in Sec. 2.2) we cannot be confident that ERA-40 exhibits realistic winds in this region either. We reiterate from Sec. 1.5.1 that the QBO-control comparison should be interpreted as a perturbation experiment comparing the climates of two GCM runs with different tropical wind conditions, neither of which is wholly realistic.

Fig. 3.1 shows that CMAM and ERA-40 differ in the strengths of their NH and SH winter stratospheric polar vortices. The NH vortex in CMAM is typically too weak (in SON and DJF) by $\approx 10$ m/s, and the SH vortex too strong (in JJA) by $\approx 10$ m/s, and the SH vortex too strong (in JJA) by

---

1. Each CMAM run is 150 years long, but the first year of each run is excluded from the composites so as to eliminate dependence on the initial conditions, which are the same for both runs. This procedure is followed for all subsequent CMAM composites.

2. As noted in Sec. 1.5.3, ERA-40 is constrained to closely match tropical radiosonde data up to an altitude of 10 hPa. Hence it is meaningful to assess the realism of CMAM $\bar{u}$ against ERA-40 $\bar{u}$ in the tropical stratosphere at altitudes up to 10 hPa ($\approx 32$ km).
Figure 3.1: Seasonal average zonal average zonal wind $\bar{u}$ in (left to right) ERA-40 reanalysis, CMAM QBO run and CMAM control run. CMAM runs are 149-year composites. Contour interval: 5 m/s. Red is positive and blue negative, with the zero line in thick black.
$\approx 20$ m/s, compared to ERA-40. These biases occur in both QBO and control runs. Nevertheless, the change in tropical winds between the QBO and control runs ($\Delta \bar{u}_{EQ}$) does affect the extratropical stratosphere. Seasonal (three-month average) QBO-control differences in $\bar{T}$ and $\bar{u}$ (hereinafter denoted as $\Delta \bar{T}$ and $\Delta \bar{u}$) are shown in Fig. 3.2. Shading in Fig. 3.2 indicates that the difference is statistically significant at least at the 95% level by the standard $t$-test. The differences shown in Fig. 3.1 support rejection of the null hypothesis that the extratropical circulation is the same in the two simulations. However, Fig. 3.2 shows that the QBO-control $\Delta \bar{u}_{EQ}$ does not induce extratropical changes of sufficient magnitude, or of the correct spatial distribution, to correct the NH and SH polar vortex biases in CMAM.

The timing of the strongest extratropical $\Delta \bar{u}$ and $\Delta \bar{T}$ responses shown in Fig. 3.2 is consistent with the Holton-Tan hypothesis (Sec. 1.2) that the propagation of stationary planetary waves in the winter stratosphere should be affected by changes in the position of the low-latitude $\bar{u} = 0$ line. Fig. 3.2 shows that the strongest DJF and JJA differences occur in the winter hemispheres, consistent with this mechanism\(^3\). Fig. 3.2 also shows that the winter stratosphere in the QBO run is cooler than in the control run, accompanied by warming at higher altitudes. This is also consistent with the Holton-Tan hypothesis, in that tropical westerlies are expected to lead to enhanced propagation of stationary planetary wave activity into the tropics, thus reducing the amount of wave-driven downwelling that maintains $\bar{T}$ above radiative equilibrium values at higher latitudes in the stratosphere. The warming at higher altitudes, in turn, is consistent with the expected gravity wave feedback due to $\bar{u}$ changes at lower altitudes [Holton, 1983].

Are the QBO-control differences similar to the ERA-40 W-E differences? The QBO run is characterized by stronger westerlies in the tropical lower stratosphere than are found in the control run. Even though these winds are oscillatory (Fig. 2.3), they vary between strong and weak westerlies and thus remain westerly on average.

\(^3\)Statistically significant differences also occur in the summer hemispheres, but they are physically small. Their high statistical significance is due to the relative quiescence (i.e., low variability) of the summer stratosphere and mesosphere and the large number of years in the dataset.
Figure 3.2: Seasonal QBO-control differences for 149 year-composite zonal average temperature $\bar{T}$ (left) and zonal average zonal wind $\bar{u}$ (right). Contour intervals are 1 K for $\bar{T}$ and 2 m/s for $\bar{u}$. Red is positive and blue negative, with the zero line in thick black. Shading indicates differences that are at least 95% significant by the $t$-test (see text for details).
3. Comparison of the CMAM QBO and control runs

Figure 3.3: ERA-40 composite W-E differences for (a) $\overline{T}$ and (b) $\overline{u}$ during DJF. Contours and shading as in Fig. 3.2. QBO phase is defined according to the sign of the deseasonalized $\overline{u}$ at the equator, using the mass-weighted average of the 30 hPa and 50 hPa levels.

(Fig. 2.5), whereas the control run exhibits weak easterlies in this region (Figs. 2.4 and 2.5). If the time-mean state of $\overline{u}$ in the tropical lower stratosphere is a major determinant of the extratropical mean state during winter, we might expect the extratropical QBO-control differences to show strong similarities to the ERA-40 W-E differences.

ERA-40 W-E differences are shown in Fig. 3.3. The W and E composites are determined by using the sign of deseasonalized equatorial $\overline{u}$, using the mass-weighted average of $\overline{u}$ at the 30 hPa and 50 hPa levels to define QBO phase\(^4\). Comparison of Figs. 3.3 and 3.2 shows that the DJF extratropical CMAM QBO-control differences deviate from the DJF extratropical ERA-40 W-E differences in several respects. For $\Delta \overline{T}$, the W-E case shows cooling in the lower stratosphere that maximizes at the pole and shrinks to zero at 50°N, consistent with Dunkerton and Baldwin [1991]. The QBO-control case shows stratospheric cooling that extends almost to the pole but maximizes in midlatitudes (between 40°N and 50°N; Fig. 3.2), rather than at the pole. The peak magnitude of the QBO-control cooling, 4 K, is similar to the peak magnitude of W-E cooling. Both CMAM and ERA-40 display a quadrupole pattern.

\(^4\)Dunkerton and Baldwin [1991] found that the 40 hPa QBO yielded the highest correlation with extratropical $\overline{u}$ and $\overline{T}$ during NH winter. The public ERA-40 data excludes the 40 hPa level, so we used the mass-weighted average of the 30 and 50 hPa levels.
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Figure 3.4: Latitudinal profiles of $\bar{T}$ in ERA-40 at 50 hPa (left) and CMAM at 10 hPa (right). Red corresponds to QBO-W for the ERA-40 plot (with QBO phase defined as in Fig. 3.3) and the CMAM QBO run for the CMAM plot. Black corresponds to QBO-E for the ERA-40 plot and the CMAM control run for the CMAM plot.

That the strongest cooling in the QBO-control case is found to be shifted upward and equatorward with respect to the ERA-40 pattern.

In summary, the strongest extratropical QBO-control $\Delta \bar{u}$ and $\Delta \bar{T}$ responses occur during winter, consistent with the Holton-Tan hypothesis that coupling between high and low latitudes in the stratosphere is mediated by stationary planetary waves. However, the QBO-control and W-E $\Delta \bar{T}$ and $\Delta \bar{u}$ signals for NH winter are not very
3. Comparison of the CMAM QBO and control runs

Fig. 3.2 shows that QBO-control $\Delta u_{EQ}$ in the tropical lower stratosphere is strongly westerly; hence one possible interpretation of the extratropical $\Delta \bar{u}$ and $\Delta \bar{T}$ responses is that the introduction of persistent westerlies in the tropical lower stratosphere is insufficient to rectify CMAM’s biased representation of the winter stratospheric polar vortices.

3.3 Residual circulation and momentum budget

The QBO-control $\Delta \bar{u}$ and $\Delta \bar{T}$ signals reflect the different eddy forcings in the two runs. Since the QBO-control $\Delta u_{EQ}$ may influence the propagation of both resolved and parameterized waves, it is of interest to determine which of these two forcing types contributes most strongly, and in what regions, to the extratropical response. To determine this, we diagnose the mean meridional circulation (MMC) that is forced by the eddies in the two runs.

3.3.1 Transformed Eulerian Mean equations

In this section we review the Transformed Eulerian Mean (TEM) formulation of the primitive equations, which provides a suitable theoretical framework for analyzing the forcing of the zonal-mean flow by eddies [Andrews et al., 1987]. Eddy forcing in the TEM system is represented as $\nabla \cdot F$, the divergence of the EP flux vector $F$, which is here determined from the resolved winds and temperature in CMAM or in ERA-40 (hereinafter $\nabla \cdot F$ will refer to the forcing due to resolved waves). The total zonal forcing is $\bar{F} = (\rho_0 a \cos \phi)^{-1} \nabla \cdot F + \bar{X}$, where $\bar{X}$ is the forcing by parameterized gravity waves. On seasonal timescales in the extratropics, $\bar{F}$ is mostly balanced by the Coriolis torque due to poleward-moving air. Thus the MMC in the TEM system, which is referred to as the residual circulation, is determined by $\bar{F}$. The radiative heating or cooling associated with $\bar{T}$, in turn, balances the vertical component of the MMC (again, on seasonal timescales), and $\bar{u}$ is in thermal-wind balance with $\bar{T}$. Hence by diagnosing the MMC response to changes in both $\nabla \cdot F$ and $\bar{X}$, we may determine the contribution of each of these forcings to $\Delta \bar{u}$ and $\Delta \bar{T}$.
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We now use the TEM equations to lay out the above argument more explicitly. The TEM equations are [Andrews et al., 1987]

\[
\frac{\partial \bar{u}}{\partial t} + \bar{v}^* \left( \frac{1}{a \cos \phi} \frac{\partial (\bar{u} \cos \phi)}{\partial \phi} - f \right) + \bar{w}^* \frac{\partial \bar{u}}{\partial z} = \frac{1}{\rho_0 a \cos \phi} \nabla \cdot \mathbf{F} + \bar{X} \tag{3.1}
\]

\[
\bar{u} \left( f + \frac{\bar{u}}{a} \tan \phi \right) + \frac{1}{a} \frac{\partial \Phi}{\partial \phi} = G \tag{3.2}
\]

\[
\frac{\partial \Phi}{\partial z} = \frac{R}{H} \bar{\theta} e^{-\kappa z/H} \tag{3.3}
\]

\[
\frac{1}{a \cos \phi} \frac{\partial (\bar{v}^* \cos \phi)}{\partial \phi} + \frac{1}{\rho_0} \frac{\partial (\rho_0 \bar{w}^*)}{\partial z} = 0 \tag{3.4}
\]

\[
\frac{\partial \bar{\theta}}{\partial t} + \bar{v}^* \frac{1}{a} \frac{\partial \bar{\theta}}{\partial \phi} + \bar{w}^* \frac{\partial \bar{\theta}}{\partial z} = \bar{Q} \tag{3.5}
\]

where \( G \) represents eddy fluxes that are negligible in comparison to the gradient wind balance terms in Eq. 3.2, and an eddy heat flux term in the thermodynamic equation, Eq. 3.5, has been neglected on the assumption that it is small compared to \( \bar{Q} \). Symbols have their standard meanings, following Sec. 3.5 of Andrews et al. [1987]. The divergence of the EP flux \( \mathbf{F} \) is given by

\[
\nabla \cdot \mathbf{F} = \frac{1}{a \cos \phi} \frac{\partial (F^{(\phi)} \cos \phi)}{\partial \phi} + \frac{\partial F^{(z)}}{\partial z} \tag{3.6}
\]

where

\[
F^{(\phi)} = \rho_0 a \cos \phi (\bar{u} \bar{v}^* \bar{\theta}/\bar{\theta}_z - \bar{v}^* \bar{w}^*), \tag{3.7}
\]

\[
F^{(z)} = \rho_0 a \cos \phi \left( f - (a \cos \phi)^{-1}(\bar{u} \cos \phi \bar{v}^*/\bar{\theta}_z - \bar{v}^* \bar{w}^*) \right). \tag{3.8}
\]

The residual circulation \((\bar{v}^*, \bar{w}^*)\) is defined as

\[
\bar{v}^* = \bar{v} - \rho_0^{-1}(\rho_0 \bar{v}^* \bar{\theta}/\bar{\theta}_z) \tag{3.9}
\]

\[
\bar{w}^* = \bar{w} + (a \cos \phi)^{-1}(\cos \phi \bar{v}^* \bar{\theta}/\bar{\theta}_z). \tag{3.10}
\]

The transformed variable \( \bar{w}^* \) approximates the mass flux across isentropic surfaces, and hence must balance the net diabatic heating \( \bar{Q} \). The spatial structure of \( \bar{w}^* \), in
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The diabatic mass flux is enabled by \( \dot{Q} \) and is forced mechanically by the torque on the zonal-mean flow due to eddy motions. Just as cross-isentropic motion may only occur when \( \dot{Q} \neq 0 \), motion across isopleths of zonal-mean zonal angular momentum per unit mass \( \dot{m} \), where \( \dot{m} = a \cos \phi (\dot{u} + a \Omega \cos \phi) \), may only occur when air parcels experience a net torque \( \rho_0 a \cos \phi \dot{F} \). This is made apparent when Eq. 3.1 is rewritten in the form \( \rho_0 D\dot{m}/Dt = \nabla \cdot \mathbf{F} + \rho_0 a \cos \phi \ddot{X} = \rho_0 a \cos \phi \dot{F} \). While radiation in the middle atmosphere can be reasonably approximated as a relaxational damping, the total zonal force results from intrinsic atmospheric dynamical variability due to waves and turbulence on a variety of spatial scales. Thus the zonal force may be usefully regarded as driving the MMC, with \( \dot{Q} \) facilitating the circulation by allowing air parcels to cross isentropic surfaces.

The relationship between the net torque and the MMC may be formulated as a diagnostic relation derived from Eqs. 3.1-3.5. The result is that on seasonal timescales, so that \( \partial \dot{m}/\partial t \approx 0 \), the circulation at an altitude \( z \) is controlled by the total amount of torque above \( z \). The diagnostic expression for the TEM streamfunction \( \Psi^* \) is \cite{Haynes et al., 1991}

\[
\Psi^* = \int_z^\infty \left( \frac{\rho_0 a^2 \dot{F} \cos^2 \phi}{\dot{m}_\phi} \right)_{\phi=\phi(z')} dz'.
\] (3.11)

which in terms of the vertical velocity \( \bar{w}^* \) becomes

\[
\bar{w}^* = \frac{1}{\rho_0 \cos \phi} \frac{\partial}{\partial \phi} \left\{ \int_z^\infty \left( \frac{\rho_0 a \dot{F} \cos^2 \phi}{\dot{m}_\phi} \right)_{\phi=\phi(z')} dz' \right\}.
\] (3.12)

Eq. 3.12 expresses the “downward control” principle that the MMC \( (\bar{v}^*, \bar{w}^*) \) at an altitude \( z \) is controlled by the distribution of total mechanical forcing \( \dot{F} \) at altitudes above \( z \). The limit in Eq. 3.12 extends to infinity (i.e., the “top” of the atmosphere), but most of the contribution to the integral at any given \( z \) arises from \( \dot{F} \) within 1-2 density scale heights of \( z \). The integral is evaluated along \( \dot{m} \)-isopleths, the contours \( \phi = \phi(z') \), which are approximately vertical in the extratropics. Isopleths of \( \dot{m} \) are less vertical in the tropics, where “sideways control” \cite{Dunkerton, 1991} would be a more
appropriate term; the principle, which is a consistency condition on the \( \bar{m} \) budget, is the same in either case. In practice it becomes difficult to apply Eq. 3.12 in the tropics because the sideways tilt of \( \bar{m} \)-isopleths may result in \( \bar{m}_\phi \approx 0 \). Additionally, the time required for \( (\bar{v}^*, \bar{w}^*) \) to equilibrate with a given \( \bar{F} \), such that Eq. 3.12 is satisfied, is proportional to \( 1/f \) and thus the steady state assumption \( \partial \bar{m}/\partial t \approx 0 \) will be a poor one in the tropics. Without making this assumption, Eqs. 3.11–3.12 will still hold provided that \( \bar{F} \) is replaced with \( \bar{F} - \partial \bar{u}/\partial t \). The problem with \( \bar{m}_\phi \) will remain, however, and accurate calculation of the transient contribution would also be required, which would require that the correlation between \( \bar{m}_\phi \) and \( \partial \bar{u}/\partial t \) be taken into account. Moreover, if \( \partial \bar{u}/\partial t \) is included then the causal utility of the diagnostic is lost, since \( \partial \bar{u}/\partial t \) cannot be partitioned into contributions from different wave forcings. As a straightforward diagnostic of the seasonal “pumping” of the MMC by the zonal torque due to eddies, Eq. 3.12 is useful mainly in the extratropics.

3.3.2 \( \bar{w}^* \) and \( \nabla \cdot F \) composites

Fig. 3.5 shows the seasonal average differences between the QBO and control runs for \( \nabla \cdot F \) and \( \bar{w}^* \). \( \nabla \cdot F \) has been scaled by \( (\rho_0 a \cos \phi)^{-1} \) to show it as a wind tendency (i.e., force per unit mass) in m/s/day. If \( \Delta (\nabla \cdot F) \) is the dominant contribution to \( \Delta \bar{F} \), then \( \Delta \bar{w}^* \) in the extratropics should be consistent with \( \Delta (\nabla \cdot F) \) in the sense implied by Eq. 3.12.

In the extratropical stratosphere, Fig. 3.5 shows the spatial structure of \( \Delta \bar{w}^* \) to be qualitatively consistent with the \( \Delta \bar{T} \) patterns shown in Fig. 3.2 – that is, regions of anomalous cooling correspond to regions of anomalous upward residual motion (\( \Delta \bar{w}^* > 0 \)) and vice versa. For \( \Delta (\nabla \cdot F) \), Fig. 3.5 shows that during the winter seasons (DJF and JJA), \( \Delta \bar{w}^* \) is broadly consistent with the downward control principle, in that anomalously positive (negative) \( \nabla \cdot F \) is found above and slightly equatorward of regions of anomalous upwelling (downwelling).

However, \( \Delta \bar{X} \) (due to parameterized gravity waves) may also contribute to \( \Delta \bar{w}^* \), particularly in the mesosphere. The fact that \( \Delta \bar{w}^* \neq 0 \) during the equinox seasons (MAM, SON), when the differences in \( \nabla \cdot F \) are slight, is suggestive of this possi-
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Figure 3.5: Seasonal QBO-control differences in 149 year-composite vertical residual velocity \( \bar{w}^* \) (left) and total EP flux divergence \((\rho_0 a \cos \phi)^{-1} \nabla \cdot \mathbf{F}\) (right). Contour intervals for \( \bar{w}^* \) are 0.1 mm/s in the range \([-0.5, 0.5]\), 1 mm/s for \(1 \leq |\bar{w}^*| \leq 5\), and 5 mm/s for \(|\bar{w}^*| \geq 10\). Contour intervals for \((\rho_0 a \cos \phi)^{-1} \nabla \cdot \mathbf{F}\) are 0.5 m/s/day in the range \([-1.5, 1.5]\) and 2 m/s/day for values \(|(\rho_0 a \cos \phi)^{-1} \nabla \cdot \mathbf{F}| \geq 2\). Shading and colours as in Fig. 3.2.
bility (although this behaviour could also indicate transience, as the steady state approximation is poorer for these seasons than during the solstitial seasons). A more quantitative consideration of the momentum budget, based on Eq. 3.12, is desirable and is considered next, in Sec. 3.3.3.

For completeness it should also be noted that Fig. 3.5 shows that substantial changes in $\bar{w}^*$ and $\nabla \cdot \mathbf{F}$ occur at high altitudes (above 40 km) in the tropics. Due to the large effect of the parameterized nonorographic gravity waves in this region (Sec. 1.5.1), such changes are not surprising. Detailed investigation of the structure of these changes will not be pursued in the present work.

3.3.3 Momentum budget diagnostics

Using Eq. 3.12, the effect of a given zonal forcing $\bar{F}$ on the MMC may be diagnosed. Physically, the response to $\bar{F}$ is balanced partially by a MMC and partially by a zonal wind tendency, as indicated by Eq. 3.1; the exact partitioning between $(\bar{u}^*, \bar{w}^*)$ and $\partial \bar{u}/\partial t$ depends on both the latitudinal and vertical structure of the forcing as well as the ratio of the forcing timescale to the radiative timescale [Haynes et al., 1991]. In general, the response to forcings that are of large enough latitudinal and vertical scales and that are “slow” in comparison with the radiative timescale will be closest to the downward control limit (i.e., to Eq. 3.12). Radiative timescales in the stratosphere vary with altitude, tending to be short ($\approx 2$ days) in the upper stratosphere and long ($\approx 2$ weeks or more) in the lower stratosphere. Hence the response to the large-scale seasonal forcings shown in Fig. 3.5 should be well characterized by downward control over the whole depth of the extratropical stratosphere.

We use Eq. 3.12 to diagnose the $\bar{w}^*$ response to each of four forcing cases in turn: parameterized nonorographic gravity waves, parameterized orographic gravity waves, resolved waves (already shown in Fig. 3.5), and the sum ($\bar{F}$) of all three forcings. The $\bar{w}^*$ response to each forcing is determined by replacing $\bar{F}$ in Eq. 3.12 with that forcing. However, a technical detail must be noted at this point. Due to the large amounts of data involved, parameterized gravity wave tendencies were not saved for the whole 150 years of both CMAM runs. Instead, 20 years of each experiment were re-run
and sampled at a higher rate (every six hours, vs. 24-hour sampling for the 150-year data) in order to determine the behaviour of the parameterized gravity wave forcing. This was judged to provide an adequate estimation of the climatological values of the gravity wave forcing, given the uncertainties inherent in the diagnostic approach utilizing Eq. 3.12. Since our goal is only to statistically attribute the extratropical \( \bar{w}^* \) changes to the different types of zonal forcing, this measure of the parameterized gravity wave contributions is sufficient.

Fig. 3.6 shows the zonal forcing \( F_{NGW} \) due to parameterized nonorographic gravity waves (hereinafter NGW) in the 20-year reruns for (a) the QBO run, (b) the QBO-control difference in \( F_{NGW} \), and (c) the difference between the \( \bar{w}^* \) fields diagnosed independently from \( F_{NGW} \) in each run using Eq. 3.12. Since \( \partial \bar{m} / \partial \phi \) approaches zero at low latitudes, the domain shown in Fig. 3.6(c) is restricted to \( \phi > 35^\circ N \), over which Eq. 3.12 is valid. For these latitudes, and in the stratosphere, comparison of Fig. 3.6(c) and Fig. 3.5 (the DJF panel) shows that \( F_{NGW} \) contributes negligibly to \( \Delta \bar{w}^* \). Hence changes in NGW forcing are ruled out as the direct cause of the extratropical QBO-control differences shown in Figs. 3.2 and 3.5. Of course, the NGWs contribute indirectly to these changes, since they induce a QBO in the model.
3. Comparison of the CMAM QBO and control runs

3.1 Comparison of the CMAM QBO and control runs

Figure 3.7: Zonal forcing $F_{OGW}$ and $\bar{w}^*$ response due to parameterized orographic gravity waves (OGW) for DJF, for the 20-year reruns. a) $F_{OGW}$ for the QBO run; b) QBO-control $F_{OGW}$ difference; c) QBO-control difference in the $\bar{w}^*$ response diagnosed from $F_{OGW}$ in each run. Contours and colours as in Fig. 3.6.

(Sec. 1.5.1). In the tropics, where a greater portion of the response to a given $\bar{F}$ is felt as a wind tendency rather than as a MMC [Haynes, 1998], NGWs cause large changes in $\bar{u}$ that affect the propagation of both resolved waves and parameterized orographic gravity waves (hereinafter OGW). The large tropical $\Delta \bar{u} > 0$ feature that is seen in all seasons in Fig. 3.2 is consistent with the nature of the tropical response. Although the exact spatial structure of the tropical $\Delta \bar{u}$ does vary with season (indicating feedbacks due to other wave forcings), $\Delta \bar{u} > 0$ above $z \approx 40$ km in the tropics is a consistent feature of all four seasons. In contrast, the extratropical response to a given $\bar{F}$ is felt mainly as a MMC, to which $F_{NGW}$ has been shown to make a negligible direct contribution.

Fig. 3.7 shows the QBO-control differences in parameterized orographic gravity wave drag (hereinafter OGW). The spatial structure of $F_{OGW}$ (Fig. 3.7(a)) results from the latitudinal distribution of orography and the effect of $\bar{u}$ on the saturation of the waves. The $F_{OGW} < 0$ region at $35^\circ$N-$40^\circ$N, 20 km results from large-amplitude orography (principally the Himalayas) as well as $\partial \bar{u}/\partial z < 0$ above the tropospheric jet. Unlike the NGW case, the parameters of the OGW scheme are identical in both CMAM runs, but changes in $\bar{u}$ will affect the distribution of $F_{OGW}$. With $\bar{u} > 0$, increased $\bar{u}$ raises the waves’ saturation threshold, leading to less OGW absorption,
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Figure 3.8: Zonal forcing and $\bar{w}^*$ response due to the EP flux divergence, for DJF, for the actual QBO-control differences (i.e., not the 20-year reruns). a) Zonal forcing $(\rho_0 a \cos \phi)^{-1} \nabla \cdot \mathbf{F}$ in the QBO run; b) QBO-control $(\rho_0 a \cos \phi)^{-1} \nabla \cdot \mathbf{F}$ difference; c) QBO-control difference in the $\bar{w}^*$ response induced by $(\rho_0 a \cos \phi)^{-1} \nabla \cdot \mathbf{F}$ in each run. Contours and colours as in Fig. 3.6.

but since the overall amount of OGW momentum flux is set by wave sources at the ground, decreased absorption\(^5\) at a lower altitude must be accompanied by greater absorption at a higher altitude (and vice versa). Comparison of $\Delta \bar{u}$ in Fig. 3.2 (for DJF) with $\Delta \bar{F}_{OGW}$ in Fig. 3.7(b) indicates that this occurs.

Comparison of Figs. 3.7(b) and 3.5 (for DJF) shows that the $\bar{F}_{OGW}$ difference is comparable in magnitude, but of opposite sign, to the $(\rho_0 a \cos \phi)^{-1} \nabla \cdot \mathbf{F}$ difference, although the resolved waves still dominate $\bar{F}$ below 50 km. Above 50 km, $\bar{F}_{OGW}$ and $(\rho_0 a \cos \phi)^{-1} \nabla \cdot \mathbf{F}$ are of roughly equal importance in the 35°N-50°N region. The $\bar{w}^*$ difference diagnosed from $\bar{F}_{OGW}$ is shown in Fig. 3.7(c). Its magnitude is comparable to that of the actual DJF $\bar{w}^*$ difference (Fig. 3.5) above 40 km and equatorward of 50°N. It is of opposite sign to the actual difference in the 40°N-60°N region – consistent with $\bar{F}_{OGW}$ and $(\rho_0 a \cos \phi)^{-1} \nabla \cdot \mathbf{F}$ having opposite sign, as already noted – indicating that the OGWs provide partial compensation in this region to $\bar{w}^*$ changes induced by the resolved waves.

Resolved waves make the largest contribution to the momentum budget of the extratropical stratosphere. Fig. 3.8 shows the forcing and diagnosed DJF $\bar{w}^*$ differences

\(^5\)i.e., positive anomalous $\bar{F}_{OGW}$, since $\bar{F}_{OGW} < 0$ in the stratosphere, as Fig. 3.7(a) shows.
for the actual QBO and control runs (i.e., all 149 years of data), while Fig. 3.9 shows the same thing for the 20-year reruns. Results from the full runs and the reruns are seen to be similar\textsuperscript{6}. Panels (a) and (b) of both figures show that \((\rho_0 \, a \cos \phi)^{-1} \nabla \cdot \mathbf{F}\) is the dominant contributor to \(\bar{F}\) in the stratosphere, although (as noted above) \(\bar{F}_{OGW}\) also becomes relatively important above 40 km.

The \(\bar{w}^*\) differences diagnosed from \((\rho_0 \, a \cos \phi)^{-1} \nabla \cdot \mathbf{F}\) are shown in Figs. 3.8(c) and 3.9(c). The diagnosed \(\bar{w}^*\) differences (Figs. 3.8(c) and 3.9(c)) and the actual \(\bar{w}^*\) difference (Fig. 3.5) have similar spatial distribution and magnitude below 50 km. Disagreement occurs above 50 km, indicating the importance of \(\bar{F}_{OGW}\) at these altitudes. If not for the OGWs, the resolved waves would – all else being equal – force a larger midlatitude cooling (i.e., a more positive \(\bar{w}^*\) difference) in the stratosphere than is actually seen. Hence the OGW feedback acts to limit the magnitude of the midlatitude \(\bar{T}\) response to the QBO, but the qualitative character of the response in the stratosphere is nevertheless determined by the resolved waves.

Is the OGW feedback responsible for the fact that the CMAM QBO-control \(\Delta \bar{T}\) (Fig. 3.2) and the ERA-40 W-E \(\Delta \bar{T}\) (Fig. 3.3) do not look very similar? Fig. 3.7 shows that the strongest OGW effects are centered on midlatitudes and above 40 km, indicating that the direct effect of OGWs cannot be responsible for the fact that

\[\text{Figure 3.9: As Fig. 3.8, but for the 20-year rerun composite differences.}\]

\[\text{\footnotesize \textsuperscript{6}This indicates that the 20-year reruns, from which the NGW and OGW forcings were obtained, are representative of climatological conditions in the full 149 years of the original runs.}\]
3. Comparison of the CMAM QBO and control runs

CMAM fails to simulate the polar cooling that characterizes the ERA-40 response to $\Delta \bar{u}_{EQ}$ (Fig. 3.3). The fact that the OGWs exert a strong feedback on the response of $\bar{w}^*$ to $\Delta (\nabla \cdot \bar{F})$ in CMAM is due to the fact that the most significant changes in resolved wave dissipation in CMAM occur at higher altitudes and further equatorward than in ERA-40. In other words, significant values of $\Delta (\nabla \cdot \bar{F})$ occur in regions where OGW forcing is climatologically strong, resulting in a strong OGW feedback. This feedback, wherein a $\bar{u}$ change at lower altitudes is accompanied by an opposite $\bar{u}$ anomaly at higher altitudes, is a physically robust effect that is observed to accompany observed stratospheric sudden warmings [Holton, 1983]. It should occur in the real atmosphere’s response to $\Delta \bar{u}_{EQ}$ – as $\Delta \bar{T}$ in Fig. 3.3 indicates – but it is likely that the CMAM response to $\Delta \bar{u}_{EQ}$ involves an OGW feedback that is stronger than the observed feedback, due to the $\Delta \bar{u}$ response in CMAM occurring closer to the region of strongest OGW forcing.

The combined effect of all three forcings ($\bar{F}_{NGW}$, $\bar{F}_{OGW}$, and $\nabla \cdot \bar{F}$) is shown in Fig. 3.10 for the 20-year reruns, with panel (b) showing the QBO-control difference in total zonal forcing $\bar{F}$. Comparison of the actual $\bar{w}^*$ response (panel (a)) with the diagnosed $\bar{w}^*$ response (panel (c)) shows that the major aspects of the response are well captured. As described above, the OGW feedback is seen to reverse the resolved wave response above 50 km and equatorward of 50 °N, leading to a vertical quadrupole
in $\Delta \bar{w}^*$ (Fig 3.5), and therefore in $\Delta \bar{T}$ also (Fig. 3.2).

Hence forcing by resolved waves is the major contributor to stratospheric $\Delta \bar{w}^*$ and $\Delta \bar{T}$ in CMAM. We now consider, in Sec. 3.4, the nature of QBO-control $\nabla \cdot \mathbf{F}$ differences in more detail.

### 3.4 Eliassen-Palm flux changes

For the DJF QBO-control (CMAM) and W-E (ERA-40) cases, the EP flux components $F^{(\phi)}$, $F^{(z)}$ and the divergence $\nabla \cdot \mathbf{F}$ are shown in Figs. 3.11, 3.12, and 3.15, and the separate contributions of $F^{(\phi)}$ and $F^{(z)}$ to $\nabla \cdot \mathbf{F}$ are shown in Figs. 3.13 and 3.14, respectively. The EP fluxes are computed from the total resolved eddy fluxes, containing all zonal wavenumbers and phase speeds. A description and comparison of the QBO-control and W-E differences is given first, followed by discussion, and then further diagnostics. The main question of interest is the means by which low-latitude changes in $\bar{u}$ are able to influence the seasonal development of $\bar{u}$ at high latitudes.

#### 3.4.1 Description of DJF EP flux changes

The climatological sense of wave propagation throughout most of the stratosphere is equatorward, as indicated by $F^{(\phi)} < 0$ in Figs. 3.11(a,b,d,e), although poleward propagation occurs in the polar lower stratosphere. The differences shown in Figs. 3.11(c,f) indicate substantially increased equatorward propagation ($\Delta F^{(\phi)} < 0$) in a latitudinally broad region that tilts equatorward with increasing altitude. The peak $\Delta F^{(\phi)} < 0$ signal in this region is centred at roughly 50°N, 30 km, and the magnitude of this difference is $\approx 25\%$, in both CMAM and ERA-40. Equatorward propagation is reduced ($\Delta F^{(\phi)} > 0$) in two regions that flank the central $\Delta F^{(\phi)} < 0$ region. Thus an equatorward-tilted tripole pattern in $\Delta F^{(\phi)}$ characterizes both datasets, but the shape of this feature differs slightly between them: the subtropical $\Delta F^{(\phi)} > 0$ cell at $z \approx 20$ km has a stronger equatorward tilt in CMAM, but its spatial extent is larger in ERA-40. Also, the altitude at which the tripole pattern is centred is lower in CMAM than in ERA-40.
Figure 3.11: DJF $F^{(\phi)}$ for CMAM (top row) and ERA-40 (bottom row); see text for details. Contours are spaced at intervals of $2 \times 10^n$ kg/s$^2$ where $n = (0, 1, 2, ...)$ and values of $10^n$ are highlighted (e.g. 2, 4, 6, 8, 10, 20, 40, ...). Colours and shading as in Fig. 3.2. The $\bar{u} = 0$ line for the two categories in each dataset is shown superimposed (green = CMAM-QBO or QBO-W, magenta = CMAM-control or QBO-E).

The meridional width of the stratospheric waveguide for stationary planetary waves is expected to affect the waves’ equatorward propagation [Matsuno, 1970], and as discussed in Chap. 1, it has been hypothesized that tropical westerlies during the QBO-W phase create an expanded meridional waveguide that allows increased equatorward propagation [Holton and Tan, 1980]. The large $\Delta F^{(\phi)} < 0$ signal centered at roughly $50^\circ$N, 30 km in Figs. 3.11(c,f) seems consistent with that expectation, but two points are worth remarking on. First, $\Delta F^{(\phi)}$ is not always negative, and the subtropical $\Delta F^{(\phi)} > 0$ cell at $z \approx 20$ km occurs poleward of the tropical westerlies in the lower stratosphere, where we might have instead expected enhanced equator-
ward propagation due to an expanded waveguide. Second, the $\Delta F^{(\phi)} < 0$ region tilts equatorward in the mid-stratosphere, so that enhanced equatorward propagation at these altitudes occurs just where the $\bar{u} = 0$ line is further poleward. In the tropical stratosphere, however, the behaviour of $\Delta F^{(\phi)}$ in both datasets is mostly consistent with the expectation that regions of enhanced equatorward propagation ($\Delta F^{(\phi)} < 0$) should occur where the $\bar{u} = 0$ line shifts southward, and vice versa. (The $z \approx 25$-30 km region in CMAM may be an exception.)

The position of the $\bar{u} = 0$ contour determines the meridional extent of the stratospheric region that is accessible to propagating stationary planetary waves, and thus confines wave activity to the winter hemisphere. Horizontally propagating waves effect a meridional redistribution of momentum, as expressed concisely by the quasi-
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The climatological sense of vertical wave propagation is everywhere upward, indicated by $F(\phi) > 0$ in Figs. 3.12(a,b,d,e). Unlike the $\Delta F(\phi)$ patterns, the CMAM and ERA-40 $\Delta F(z)$ patterns are strikingly different. In ERA-40, Fig 3.12(f) shows that QBO-W is associated with reduced upward propagation, $\Delta F(z) < 0$, in midlatitudes, which is a statistically significant feature that extends from the tropopause upward to $z \approx 25$ km. In CMAM, conversely, Fig 3.12(c) shows enhanced upward propagation, $\Delta F(z) > 0$, in roughly the same region, although slightly further poleward. Similarly to the oppositely-signed ERA-40 feature, the statistically significant CMAM $\Delta F(z)$ extends all the way down to the tropopause, $z \approx 10$ km. Unlike the ERA-40 feature, however, it extends further upward. Both the CMAM and ERA-40 midlatitude features change sign at $z \approx 30$ km, but the CMAM $\Delta F(z)$ remains statistically significant for $z > 30$ km while the ERA-40 $\Delta F(z)$ does not.

The fact that changes in both meridional and vertical propagation occur as a response to a tropical wind perturbation indicates that $\Delta F(z)$ and $\Delta F(\phi)$ should not be considered in isolation of each other. Since planetary waves propagate both upward and equatorward, changes in $F(z)$ at lower altitudes may affect both $F(z)$ and $F(\phi)$ in regions “downstream” of the $F(z)$ change – that is, further upward and equatorward. Comparison of panels (c) and (f) of Figs. 3.11 and 3.12 shows that this occurs. The CMAM midlatitude $\Delta F(\phi) < 0$ feature (centred at roughly 50°N, 30 km; Fig. 3.11(c)) has an interior extremum while the corresponding $\Delta F(\phi)$ in ERA-40 is more barotropic (Fig. 3.11(f)). This difference may be attributed to the different $\Delta F(z)$ behaviour in the two cases: increased upward propagation in CMAM (Fig. 3.12(c)) has contributed to increased equatorward propagation at higher altitudes. However, comparison of the magnitudes of $\Delta F(\phi)$ and $\Delta F(z)$ relative to the climatological strengths of $F(\phi)$ and $F(z)$ - which are 25% and 10%, respectively – as well as consideration of the shape of
the $\Delta F^{(\phi)} < 0$ feature (which only overlaps partially with the $\Delta F^{(z)}$ feature) indicate that the enhanced equatorward propagation is not solely a result of enhanced upward propagation. Similarly, the reduced equatorward propagation ($\Delta F^{(\phi)} > 0$) feature at 30°N-40°N in the lower stratosphere in both datasets is partially due to the regions of reduced upward propagation, $\Delta F^{(z)} < 0$, just below them. At polar latitudes in the mid-upper stratosphere (in both datasets) reduced equatorward propagation ($\Delta F^{(\phi)} > 0$) is consistent with reduced upward flux ($\Delta F^{(z)} < 0$) into this region from lower altitudes, while the reduced upward flux itself is consistent with the increased equatorward propagation ($\Delta F^{(\phi)} < 0$) occurring at lower altitudes.

Thus the extratropical response of resolved wave activity to $\Delta \bar{u}_{EQ}$ involves changes in both the meridional and vertical propagation of waves, in both CMAM and ERA-40. While plots of the fluxes ($F^{(\phi)}, F^{(z)}$) and their differences ($\Delta F^{(\phi)}, \Delta F^{(z)}$) are helpful for understanding these changes, their potential dynamical impact is better assessed by plotting the individual contributions of ($F^{(\phi)}, F^{(z)}$) to $\nabla \cdot F$. These are shown in Figs. 3.13 and 3.14, respectively. Comparison of these plots with their sum $\nabla \cdot F$, shown in Fig. 3.15, indicates that the relative contributions of ($F^{(\phi)}, F^{(z)}$) to $\nabla \cdot F$ and of ($\Delta F^{(\phi)}, \Delta F^{(z)}$) to $\Delta (\nabla \cdot F)$ vary with location.

The $F^{(\phi)}$ contribution to $\nabla \cdot F$ indicates the meridional convergence and divergence of $\bar{m}$. In a horizontal layer this may be regarded solely as the rearrangement of $\bar{m}$ within the layer, provided that the vertical contribution to $\nabla \cdot F$ does not change. Since the $F^{(z)}$ contribution does change, as shown in Fig. 3.14(c,f), we cannot in general regard $\bar{m}$ within a layer as being fixed, although there are regions where this is approximately true. One such region is the subtropical mid-stratosphere, at $\phi \approx 20°N-40°N$ and $z \approx 30-40$ km in ERA-40, $z \approx 25-35$ km in CMAM. Comparison of the difference panels (c) and (f) in Figs. 3.13, 3.14 and 3.15 shows that the $\Delta (\nabla \cdot F) < 0$ feature in this region is dominated by the $\Delta F^{(\phi)}$ contribution (in both datasets). This agrees with the anticipated effect of the poleward-shifted $\bar{u} = 0$ line at these altitudes (shown superimposed in the plots), which may be expected to enhance Rossby wave breaking (RWB) in a latitudinally broad “surf zone” in a manner that is well captured by layerwise (two-dimensional) dynamics [Juckes and McIntyre, 1987],
3. Comparison of the CMAM QBO and control runs

Figure 3.13: DJF \((\rho_0 \cos \phi)^{-1} (a \cos \phi)^{-1} \partial F^{(\phi)} / \partial \phi\) for CMAM (top row) and ERA-40 (bottom row); see text for details. Contours are as in Fig. 3.6, except that on the difference plots, contours with spacing 0.1 m/s/day have been added in the range \([-0.5, 0.5]\) m/s/day so as to make lower stratospheric changes visible. Colours, shading and superimposed \(\bar{u} = 0\) lines as in Fig. 3.11.

without requiring any vertical redistribution of \(\bar{m}\) between layers. The enhanced convergence of \(F^{(\phi)}\) is greater in ERA-40. This is qualitatively consistent with the much larger modulation of the \(\bar{u} = 0\) position in this layer; the CMAM \(\bar{u} = 0\) shift is modest in comparison\(^7\).

Moving poleward in the same horizontal layer (the mid-stratosphere) into the \(\phi \approx 40^\circ\text{N}-70^\circ\text{N}\) region, Fig. 3.13(a,b,d,e) shows that in both datasets the \(F^{(\phi)}\) contribution to \(\nabla \cdot \mathbf{F}\) becomes divergent. Comparison of the difference plots shows that in ERA-40, layerwise dynamics again characterizes the overall \(\Delta(\nabla \cdot \mathbf{F})\) pattern in

\(^7\)Examination of \(\Delta(\nabla \cdot \mathbf{F})\) for the W-E difference in the CMAM QBO run confirms that a stronger \(\bar{u} = 0\) shift does yield a stronger subtropical \(\Delta(\nabla \cdot \mathbf{F})\) signal (not shown).
3. Comparison of the CMAM QBO and control runs

Figure 3.14: DJF \((\rho_0 a \cos \phi)^{-1} \partial F(z)/\partial z\) for CMAM (top row) and ERA-40 (bottom row); see text for details. Contours, colours, shading and superimposed \(\bar{u} = 0\) lines as in Fig. 3.13.

This region. Although \(\Delta F(z)\) is convergent (Fig. 3.14(f)), the divergence of \(\Delta F(\phi)\) (Fig. 3.13(f)) dominates the \(\Delta(\nabla \cdot F)\) difference pattern (Fig. 3.15(f)). In CMAM this is not the case: \(\Delta F(z)\) is convergent (Fig. 3.14(c)) and slightly stronger than the divergent \(\Delta F(\phi)\) contribution (Fig. 3.13(c)); hence the net result is that \(\Delta(\nabla \cdot F)\) is weakly convergent (Fig. 3.15(c)), rather than strongly divergent as in ERA-40 (Fig. 3.15(f)). This difference in \(\Delta(\nabla \cdot F)\) is the reason for the qualitatively different high-latitude mean-flow responses to tropical wind changes in the two datasets. Thus it appears that while the \(F(\phi)\) response to a poleward-shifted \(\bar{u} = 0\) line is fairly robust – as comparison of panels (c) and (f) of Fig. 3.13 strongly suggests – the vertical redistribution of momentum is a more subtle issue, and one that determines the character of the extratropical response to \(\Delta \bar{u}_{EQ}\).
3. Comparison of the CMAM QBO and control runs

Figure 3.15: DJF \((\rho_0 a \cos \phi)^{-1} \nabla \cdot \mathbf{F}\) for CMAM (top row) and ERA-40 (bottom row); see text for details. Contours, colours, shading and superimposed \(\bar{u} = 0\) lines as in Fig. 3.13.

Similar considerations prevail in the lower stratosphere, at \(z \approx 15\text{-}25\) km, in both datasets. In the subtropics, a meridional dipole in the \(\Delta F^{(\phi)}\) contribution to \(\Delta (\nabla \cdot \mathbf{F})\), with the node of the dipole at roughly \(30^\circ\)N, occurs slightly equatorward of its oppositely-signed counterpart in the mid-stratosphere, a feature that is again consistent with the \(\bar{u} = 0\) shift, in that a southwards shift of the line coincides with anomalously divergent meridional EP flux on the northern flank of the QBO winds. Unlike in the mid-stratosphere, \(\Delta F^{(\phi)}\) and \(\Delta F^{(z)}\) are of comparable strength throughout the lower stratosphere. The fact that both components are of equal importance in the \(20^\circ\text{-}30^\circ\)N region may perhaps in hindsight have been anticipated by noting that the \(\bar{u} = 0\) line is less vertical than in the mid-stratosphere. Poleward of this region, divergent \(\Delta F^{(z)}\) overwhelms convergent \(\Delta F^{(\phi)}\) in midlatitudes, resulting
in overall $\Delta(\nabla \cdot \mathbf{F}) > 0$. Changes at polar latitudes also involve compensation between the two components, now of opposite sign to the midlatitude changes, with the result of a negligibly small $\Delta(\nabla \cdot \mathbf{F})$ that is not statistically significant (in both datasets). A final notable feature is that the total change $\Delta(\nabla \cdot \mathbf{F})$ in the lower stratosphere is stronger in ERA-40 than in CMAM. A possible reason for this is that the $\bar{u} = 0$ shift in CMAM ($\approx 10^\circ$) is roughly half of that in ERA-40 ($\approx 20^\circ$).

The greater vertical domain of CMAM also allows consideration of $\Delta(\nabla \cdot \mathbf{F})$ at altitudes not well represented in the ERA-40 data, and a significant $\bar{u} = 0$ shift occurs at these altitudes ($z > 40$ km). The $\Delta F^{(\phi)}$ contribution to $\Delta(\nabla \cdot \mathbf{F})$ is seen to be dominant at these altitudes, except in the polar upper stratosphere where its convergent contribution is offset by divergent $\Delta F^{(z)}$. In the $\phi \approx 0^\circ$-$20^\circ$N region, Fig. 3.13(c) shows that the southwards-shifted $\bar{u} = 0$ line is followed closely by convergent $\Delta F^{(\phi)}$, consistent with the same type of behaviour at lower altitudes: enhanced Rossby wave breaking occurs in the vicinity of the $\bar{u} = 0$ line, so that shifts in the position of $\bar{u} = 0$ cause a latitudinal shift of the surf zone. This interpretation is supported by the fact that $\Delta F^{(z)}$ is a secondary contribution to $\Delta(\nabla \cdot \mathbf{F})$ in this region. Moving poleward from the $0^\circ$-$20^\circ$N region, $\Delta F^{(\phi)}$ is divergent, recreating the banded structure already seen at lower altitudes (Fig. 3.13(c)). In contrast to the situation at lower altitudes, Fig. 3.14(c) shows divergent $\Delta F^{(z)}$ occurring on the poleward flank of the divergent $\Delta F^{(\phi)}$ feature, reinforcing it and creating a strongly divergent total $\Delta(\nabla \cdot \mathbf{F})$. It is this region of $\Delta(\nabla \cdot \mathbf{F}) > 0$ in CMAM that is responsible for the QBO-control $\Delta \bar{u} > 0$ feature that is seen in midlatitudes and extends down to the lower stratosphere (Fig. 3.2). Poleward of this feature, for $\phi > 50^\circ$N and above the stratopause, we have $\Delta(\nabla \cdot \mathbf{F}) < 0$, and it is notable that this enhanced convergence of wave activity results from $\Delta F^{(\phi)}$, even though $\Delta F^{(z)}$ is divergent in this region due to a reduction in upward propagation that results from the increased equatorward propagation occurring at midlatitudes in the mid-stratosphere (Fig. 3.11). (This reduction also reinforces the already-described divergent $\Delta F^{(\phi)}$ that occurs further south.) Thus there is an increased net convergence of wave activity into this region in spite of the fact that the total amount of wave activity propagating upward
from below has been reduced; evidently, the reduction in equatorward propagation (Fig. 3.11(c)) is more influential, and wave activity appears to be trapped at high latitudes.

### 3.4.2 Interpretation of DJF EP flux changes

Changes in subtropical $\nabla \cdot \mathbf{F}$ have been shown to correlate well with the movement of the $\bar{u} = 0$ line, which is in accord with the dominant role of stationary waves in the stratosphere (for which $\bar{u} = 0$ is a critical surface). As waves propagate towards a critical surface, their intrinsic phase speed $c - \bar{u}$ approaches zero, material contours become increasingly deformed, and the waves may eventually break – meaning that deformations of material surfaces may become so great that the meridional PV gradient $P_y$, to which the waves owe their existence, will be destroyed. This process involves wave amplitudes that, in observations, are seen to span a wide latitude range [McIntyre and Palmer, 1983]. Moreover, waves are expected to break before they reach their critical surfaces [Randel and Held, 1991]. These two facts strongly suggest that the latitudinal width of the $\Delta(\nabla \cdot \mathbf{F}) < 0$ feature occurring in the subtropical mid-stratosphere is due to the shifted $\bar{u} = 0$ line causing a latitudinal shift of the stratospheric surf zone. In the lower stratosphere this modulation appears to be weaker but of the same sense (i.e., of opposite sign to the mid-stratospheric change, since $\bar{u} = 0$ is shifted in the opposite direction). The weakness of the lower level response may be due to a less vigorous surf zone, an interpretation that is suggested by the climatological frequency and intensity of Rossby wave breaking (RWB) at these altitudes [Hitchman and Huesmann, 2007]. In CMAM, a similar effect also occurs at higher altitudes ($z > 40$ km), resulting in large $\nabla \cdot \mathbf{F}$ changes that are associated with a very vigorous surf zone extending into the mesosphere on the equatorward side of the mesospheric jet.

The $\Delta F(\phi)$ contribution to $\Delta(\nabla \cdot \mathbf{F})$ consists of a banded structure that tilts equatorward with increasing altitude, and Figs. 3.13(c,f) show that it is very similar in both datasets. If a latitudinal $\bar{u} = 0$ shift induces anomalous EP flux convergence – i.e., $\Delta(\nabla \cdot \mathbf{F}) < 0$ – in a latitudinally-broad region just poleward of the shifted $\bar{u} = 0$
position, then why should this structure consist of more than one band? A simple argument appeals to the conservation of $\bar{m}$ within a horizontal layer. If the role of a poleward-shifted $\bar{u} = 0$ line is to induce enhanced RWB near the equatorward edge of the surf zone, then $\Delta(\nabla \cdot F) < 0$ must be induced in that region. Conservation of $\nabla \cdot F$ (recalling that time-mean $\nabla \cdot F$ represents sources and sinks of $\bar{m}$) then implies that compensating $\Delta(\nabla \cdot F) > 0$ occurs elsewhere. Since latitudes south of the $\bar{u} = 0$ line are inaccessible to Rossby waves, and since by assumption there is no vertical flux of wave activity out of the layer (i.e., the total $\bar{m}$ of the layer is unchanged), the $\Delta(\nabla \cdot F) > 0$ anomaly must occur at higher latitudes. The latitudinal width of the surf zone suggests that the resulting dipole in $\Delta(\nabla \cdot F)$ should be sufficiently broad to link together the region of the tropical $\bar{u} = 0$ shift with the polar vortex [McIntyre and Palmer, 1983; O’Sullivan and Salby, 1990; Polvani et al., 1995]. Moreover, empirical RWB diagnostics performed by Baldwin and Holton [1988] indicate that the signature of RWB, which they defined as reversals of $P_y$ (as in Huesmann and Hitchman [2007]), moves poleward with time over the winter. Baldwin and Holton [1988]’s result thus lends plausibility to the notion that tropical-extratropical coupling in the mid-stratosphere occurs as a result of the nonlinear dynamics of the surf zone.

The above argument explains why the effect of a $\bar{u} = 0$ shift should be felt at latitudes outside the subtropics, under the assumption that $\bar{m}$ within a layer does not change very much. However, it has already been shown that the role of $\Delta F^{(z)}$ cannot always be neglected, and the description in Sec. 3.4.2 of the $\Delta(\nabla \cdot F)$ response indicates those regions where it is important. We may therefore expect the above argument regarding horizontal redistribution of $\bar{m}$ to apply in those regions where $\Delta F^{(d)}$ dominates $\Delta(\nabla \cdot F)$; these include the subtropical mid-stratosphere in both datasets, extending to the higher-latitude mid-stratosphere in ERA-40, and to the upper stratosphere in CMAM. In these regions, the magnitude of the equatorward enhancement of wave breaking is not compensated by vertical redistributions of $\bar{m}$, and adjustment of the $\bar{m}$-balance in a layer extends to higher latitudes. (As noted above, it cannot extend to lower latitudes because the $\bar{u} = 0$ line constitutes a barrier to wave propagation).
Sec. 3.4.2 showed that $\Delta F^{(z)}$ contributions to $\Delta(\nabla \cdot F)$ increase in importance at higher latitudes and lower altitudes. Figs. 3.13(c,f) indicate that if $\Delta F^{(\phi)}$ were the sole contributor to $\Delta(\nabla \cdot F)$, then $\Delta(\nabla \cdot F) > 0$ would occur at $\phi \approx 50^\circ$N-$60^\circ$N in the mid- and lower stratosphere; the $\bar{u} = 0$ shift would thus be associated with a robust strengthening of the polar vortex in midwinter (DJF). That this is not the case, is due to $\Delta F^{(z)}$. In CMAM, the compensation from $\Delta F^{(z)}$ is large enough to change the sign of $\Delta(\nabla \cdot F)$ in the mid-stratosphere at $50^\circ$N, and there is effectively no $\Delta(\nabla \cdot F)$ signal at polar latitudes. It is for this reason, as noted in the previous section, that $\Delta \bar{u} \approx 0$ in the polar lower stratosphere in CMAM (Fig. 3.2) at these altitudes. In ERA-40 the effect is weaker, but still plays a role: comparison of Figs. 3.13(c,f) and Figs. 3.15(c,f) shows that the statistical significance of the divergent $\Delta F^{(\phi)}$ extends to higher latitudes than the statistical significance of $\Delta(\nabla \cdot F)$, and there is no $\Delta(\nabla \cdot F)$ signal in the polar lower stratosphere (in both datasets). Elucidating the mechanism for the changing relative importance of vertical and latitudinal redistributions of $\nabla \cdot F$ is therefore the key question that addresses how the effect of a tropical $\bar{u} = 0$ change is communicated to higher latitudes.

The partitioning of $\nabla \cdot F$ redistribution between vertical and horizontal components at high latitudes may be caused by the polar vortex itself acting to modulate the flux of wave activity that enters the stratosphere. Studies with mechanistic models have shown that the polar vortex may modulate $F^{(z)}$ along its axis, and that this behaviour may be considered an internal mode of vortex variability as it does not require the existence of independently variable tropospheric wave sources [Scott and Polvani, 2004, 2006]. A stratosphere-only mechanism for changes in $\Delta F^{(z)}$ is desirable here, since Figs. 3.12(c,f) and 3.14(c,f) show that statistically and dynamically significant changes in upward propagation do not extend down to altitudes below the tropopause.

In summary, a hypothesis for the mechanism by which low-latitude $\bar{u} = 0$ changes are communicated to the high-latitude polar vortex is as follows. Changes in the location of the $\bar{u} = 0$ line cause a latitudinal shift of the surf zone. Empirical diagnostics suggest that the RWB pattern may propagate poleward on a timescale of roughly
1-2 months [Baldwin and Holton, 1988]. This may potentially induce a positive \( \bar{m} \) anomaly at polar latitudes – i.e., a strengthening of the polar vortex. However, the polar anomaly that actually results will also depend on the wave-vortex feedback that is characteristic of the internal variability of the polar vortex [Scott and Polvani, 2004, 2006]. If the vortex geometry is such as to induce an overcompensating feedback, in that vertical propagation of wave activity into the stratosphere is strongly enhanced at high latitudes, then the vertical redistribution of \( \nabla \cdot \mathbf{F} \) may trump the horizontal redistribution at those latitudes. If the vortex geometry does not induce such a large feedback, then horizontal redistribution of \( \nabla \cdot \mathbf{F} \) will prevail. We hypothesize that the former situation occurs in CMAM while the latter one prevails in ERA-40. The notion that the vortex geometry is different in the two datasets is supported by Fig. 3.1, which shows that the axis of the CMAM DJF stratospheric jet is shifted upward and equatorward with respect to that of the ERA-40 jet.

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8Our motivation for suggesting a variable timescale for the poleward propagation of RWB anomalies is that large variability is seen in the rates of poleward progression of \( P_y \) reversals in Baldwin and Holton [1988]’s Fig. 4.
Chapter 4

The Holton-Tan effect in CMAM and ERA-40

4.1 Introduction

The effect of the QBO on the extratropical mean state of CMAM was examined in Chap. 3 by comparing the QBO and control runs. Here in Chap. 4 we examine the effect of the QBO on extratropical interannual variability in the model.

The NH winter polar vortex has large interannual variability regardless of whether the model includes a QBO. Without a QBO, the interannual variability of the vortex mostly reflects the stratospheric response to tropospherically-forced upward-propagating planetary waves, which are themselves highly variable from year to year. When a QBO is included in the model, it acts to modulate the behaviour of these waves, resulting in a Holton-Tan effect (HTE). In this chapter we assess the realism of the HTE in the CMAM-QBO run by comparing it to the observed (ERA-40) HTE.

In observations, correlations between the QBO and the vortex are optimized when a lower QBO altitude, 40 hPa or 50 hPa, is used to define QBO phase, but it is not known if these altitudes are causal for the HTE. The downward propagation of the QBO and its temporal periodicity, and the fact that a QBO signal in observations is evident over the whole depth of the tropical stratosphere, makes it difficult (or impossible) to infer causality from observed correlations alone. It is therefore an open
question as to what altitudes are most important for HTE causality (which includes the possibility that the whole depth of the tropical stratosphere is important).

Since interannual variability in the tropical stratosphere is unrealistic in most GCMs, a better understanding of HTE causality would be helpful for the purpose of assessing the impact of this common model deficiency on the simulation of extratropical variability. While tropical stratospheric interannual variability in the CMAM-QBO run remains unrealistic in some respects (as noted in Chap. 2), it is improved in comparison to that of QBO-less GCMs. It is therefore of interest to ask, first, what type of HTE is obtained in the model, and second, if any insight into HTE causality may be obtained by comparison of the HTE in CMAM and in ERA-40.

4.2 Defining QBO phase

Diagnosing the HTE requires an objective definition of QBO phase. The method most commonly used in the literature has been to pick a single altitude and define QBO phase as W or E based on the sign of the equatorial zonal-mean zonal wind, $\bar{u}_{EQ}$, at that altitude [Holton and Tan, 1980; Naito and Hirota, 1997; Lu et al., 2008]. A lower stratospheric altitude such as 40 hPa or 50 hPa has typically been chosen because this appears to maximize the observed QBO-vortex correlation for NH winter [Dunkerton and Baldwin, 1991; Baldwin and Dunkerton, 1998]. A further motivation for choosing a lower stratospheric altitude is that there exists a long (1953-present) record of radiosonde wind observations at near-equatorial stations, with measurements extending up to 10 hPa [Naujokat, 1986]. Although a QBO signal appears to be evident as high as the stratopause [Pascoe et al., 2005], the only record of equatorial winds above 10 hPa that extends for a period of time comparable to the radiosonde observations is that of the reanalysis datasets. The NCEP/NCAR reanalysis suffers from an unrealistically small QBO amplitude as well as an abrupt discontinuity in 1979, apparently due to the introduction of satellite data [Huesmann and Hitchman, 2001, 2003]. The ERA-40 representation of the QBO appears to be more realistic [Baldwin and Gray, 2005], although a shift of $\approx 8$ m/s in annual mean $\bar{u}_{EQ}$ at altitudes above 10 hPa, cen-
tered roughly on 1979, has been noted\textsuperscript{1} [Punge and Giorgetta, 2007]. Independently of the reanalyses, rocketsonde observations at two near-equatorial stations (Kwajalein at 8.7\textdegree N and Ascension at 7.6\textdegree S) provide data extending up to \(\approx 60\) km. However, the data have a number of gaps and cover only the period 1962-1991, with frequent observations (more than 30 soundings per year at both stations combined) only in the period 1964-1985 [Dunkerton and Delisi, 1997]. In summary, either the equatorial radiosonde record or the ERA-40 reanalysis provide the best long-term records of equatorial winds suitable for studying atmospheric variability on decadal timescales (e.g., the HTE), with the lower stratospheric region (at or below 10 hPa) appearing to be the most reliable due to the long equatorial radiosonde record (note that ERA-40 assimilates the radiosonde measurements, and is virtually identical to them in the lower stratosphere [Baldwin and Gray, 2005]). This provides some justification for picking a single lower stratospheric altitude to define QBO phase.

On physical grounds, it is not expected that planetary waves and the polar vortex are sensitive only to tropical winds at a single altitude. Planetary waves have large vertical wavelengths, of the order of the depth of the stratosphere, and may be expected to respond to equatorial wind anomalies, \(\Delta \bar{u}_{EQ}(z)\), that occur over a range of altitudes. Exactly which altitudes are most important for HTE causality is unknown\textsuperscript{2}, making it difficult to choose the optimal definition of QBO phase that should be used to diagnose the HTE. Defining QBO phase by a single altitude \(z_0\) means implicitly that \(\bar{u}_{EQ}(z_0)\) serves as a proxy for an overall vertical structure \(\bar{u}_{EQ}(z)\). Due to the intrinsic variability of QBO cycles (as discussed in Chap. 2), there is not a perfect correspondence between \(\bar{u}_{EQ}(z_0)\) and \(\bar{u}_{EQ}(z)\) for \(z \neq z_0\). Additionally, because \(\text{sgn}(\bar{u}_{EQ}(z_0))\) only distinguishes two possible QBO states, the vertical structure

\textsuperscript{1}The shift constitutes a difference in the climatological annual mean \(\bar{u}_{EQ}\) values between the 1960-1979 and 1980-1999 periods. The early period is biased by up to \(-8\) m/s with respect to the late period at altitudes above 10 hPa, but the peak-to-peak QBO amplitude is relatively similar in the two periods.

\textsuperscript{2}This question has been the subject of numerous mechanistic modelling studies, most of which employ idealized forms of \(\Delta \bar{u}_{EQ}(z)\), as reviewed in Sec. 1.3.1. Note that the symbol \(\Delta \bar{u}_{EQ}\) was first defined while discussing Bridger [1984] at the beginning of Sec. 1.3.1, where it represented anomalous \(\bar{u}_{EQ}\) that in mechanistic studies was imposed (and \(\bar{u}_{EQ}\) was defined, above, as equatorial \(\bar{u}\)). Here it is taken to represent the deviation from climatological \(\bar{u}_{EQ}\), i.e. \(\Delta \bar{u}_{EQ} = \bar{u}_{EQ} - \bar{u}_{climEQ}\).
4. The Holton-Tan effect in CMAM and ERA-40

4. The Holton-Tan effect in CMAM and ERA-40

Figure 4.1: HTE diagnosed as W-E $\Delta \bar{u}$ response, using 50 hPa equatorial $\bar{u}$ to define QBO phase, in ERA-40 (top row) and CMAM (bottom row). Nov-Feb progression. Contour interval is 2 m/s, with 95% differences (by the $t$-test) shaded. Red (blue) = positive (negative).

$\bar{u}_{EQ}(z)$ can vary considerably within each of these two classifications. For example, $\text{sgn}(\bar{u}_{EQ}(z_0))$ is the same at both the beginning and end of a W or E phase at the level $z_0$, but $\bar{u}_{EQ}(z)$ is not: the beginning of a W phase at $z_0$ implies $\partial \bar{u}_{EQ}/\partial z > 0$ for altitudes immediately above $z_0$, while the end of a W phase at $z_0$ implies $\partial \bar{u}_{EQ}/\partial z < 0$ at those same altitudes. Hence a single-altitude definition of QBO phase may be inadequate for proper characterization of the HTE.

We illustrate how one particular choice of single-level QBO phase definition leads to a misleading diagnosis of the HTE – or more specifically, a misleading comparison between CMAM and ERA-40. Fig. 4.1 shows the $\Delta \bar{u}$ response to W-E $\Delta \bar{u}_{EQ}$ for ERA-40 (top row) and the CMAM QBO run (bottom row), where QBO-W and E phases have been defined by the sign of $\bar{u}_{EQ}$ at 50 hPa. The HTE in ERA-40 consists of a strengthened polar vortex that persists throughout the winter, with statistical significance being highest in Nov and declining as winter progresses. In comparison, the CMAM HTE appears to be unrealistic. However, Fig. 4.1 shows that the vertical
structure of $\Delta \bar{u}_{EQ}$ differs between ERA-40 and CMAM, due to the shorter vertical wavelength of the CMAM QBO. Hence this particular proxy for $\Delta \bar{u}_{EQ}$ – i.e., 50 hPa QBO phase – seems inappropriate as a basis for comparing the model with measurements. A more appropriate proxy would be one that matches more closely the vertical structures of $\Delta \bar{u}_{EQ}$ in ERA-40 and CMAM.

The potential inadequacy of a single-altitude QBO phase definition is a point well appreciated in the literature, as attested by the number of mechanistic modelling studies that explore the effects of different types of $\Delta \bar{u}_{EQ}(z)$ on the polar vortex (Sec. 1.3.1). It has been suggested that planetary waves and SSWs are likely to be sensitive to a deep single-layer $\Delta \bar{u}_{EQ}(z)$ [McIntyre, 1982; Dunkerton et al., 1988], and while this has been shown to be true for the extreme case of a single-layer $\Delta \bar{u}_{EQ}(z)$ spanning the depth of the stratosphere [Gray et al., 2003], a sensitivity to single-layer perturbations of smaller vertical scale, $\approx 10$-20 km, has also been demonstrated [Holton and Austin, 1991; O’Sullivan and Young, 1992; O’Sullivan and Dunkerton, 1994; Gray, 2003; Gray et al., 2004]. The response of the vortex to realistic forms of $\Delta \bar{u}_{EQ}(z)$ (such as result from a downward-propagating QBO), rather than the simplified cases used in mechanistic studies, remains unclear.

In this chapter we diagnose the HTE by using two alternative definitions of QBO phase, both of which effectively distinguish between QBO states $\bar{u}_{EQ}(z)$ at a resolution finer than 0.5 QBO cycles (hereinafter “cycles” for brevity). The utility of this approach is contingent on the amount of data available, since we desire that statistically significant results be obtained. Using the 44 years of the ERA-40 analysis, subdivision of the QBO state at a resolution of 0.25 cycles will yield W and E composite groups with $\approx 10$ members each (in contrast, a single-level definition of QBO phase would yield groups with $\approx 20$ members each), which we deem to be an acceptable number. For comparison, note that diagnoses of the HTE in relation to the solar cycle have been made using composite groups with as few as 6 members [Naito and Hirota, 1997]. Using the 150 years of the CMAM-QBO run, somewhat larger groups may be obtained, with $\approx 15$-35 members each. We first define QBO phase in terms of the timing of phase transitions, with respect to the annual cycle, at a selected alti-
tude (Sec. 4.3). QBO phase is then defined by means of a phase angle representation utilizing the two leading EOFs of $\bar{u}_{EQ}(z)$, which provide a compact representation of $\bar{u}_{EQ}(z)$ that captures most (upward of 90%) of the variability [Wallace et al., 1993; Baldwin and Dunkerton, 1998] (Sec. 4.4). Similar results are obtained with both methods. Zonal cross sections showing the $(\phi, z)$ structure of the HTE signals may be found in Sec. 4.5.

4.3 QBO phase defined by the timing of phase transitions with respect to the annual cycle

Although the descent rate of QBO phases varies somewhat (Sec. 2.3), the regularity of the QBO implies that the timing of QBO phase transitions with respect to the annual cycle can serve as a proxy for $\bar{u}_{EQ}(z)$. The lag of a phase transition with respect to NH winter then determines, to first order, the form of $\bar{u}_{EQ}(z)$ that occurs during NH winter. By partitioning the data into subsamples defined by a specified range of phase transition lags (with respect to NH winter), it is possible to isolate the effect of different forms of $\bar{u}_{EQ}(z)$ on the polar vortex.

More explicitly, this partitioning is performed as follows. A subsample of years is defined by choosing (1) a vertical level at which QBO phase transitions occur, and (2) a range of lags with respect to NH winter. Then, whenever a QBO phase transition occurs at the chosen vertical level during the period defined by the chosen range of lags, its corresponding NH winter is included in the subsample. For example: one possible subsample may be defined by choosing the 30 hPa (25 km) level and the period 4-9 months prior to January (i.e., Apr-Sep of the year leading up to NH winter). Then, out of all the NH winters in the whole data record, only those winters for which a 30 hPa transition occurred sometime during the prior Apr-Sep period are included in the subsample. Those winters for which a 30 hPa transition did not occur during the prior Apr-Sep are excluded from the subsample. For example, the tropical radiosonde data indicate (Fig. 2.8) that in April 1975, a QBO-W phase initiated at 30
hPa. Hence the 1975/76 NH winter is included in the subsample. In January 1974, a QBO-E phase initiated at 30 hPa (and no subsequent QBO phase transition occurred until April 1975). Hence the 1974/75 NH winter is excluded from the subsample.

Having defined a pool of years as the subsample, those years are then partitioned into W and E categories. If a W (E) phase initiated at the chosen level (e.g. 30 hPa) during the chosen period (e.g., Apr-Sep leading up to NH winter), then the NH winter is assigned to the W (E) category. Differences between the composites of these two groups — i.e., W and E categories within the subsample — may then be used to measure the magnitude of QBO influence on the extratropical winter circulation.

To fix terminology, we hereinafter refer to the subsample as the “phase bin”, with the notation \( \tau(p) \), where \( p \) is the pressure in hPa, representing the lag (in months), relative to January, of a QBO phase initiation at the level \( p \). We use the convention that months prior to (i.e., leading) NH winter correspond to \( \tau(p) > 0 \). Hence the above example — the phase bin for which QBO phase initiations occur during the period 4-9 months prior to January — is denoted as \( \tau(30) \in [4, 9] \).

### 4.3.1 ERA-40 results

To illustrate the different forms of \( \Delta \bar{u}_{EQ}(z) \) that are associated with different phase bin choices, Fig. 4.2 shows the W-E composite differences \( \Delta \bar{u}(z) \) at \( \phi = 15^\circ \text{N} \) for the two phase bins \( \tau(30) \in [9, 14] \) (Nov-Apr) and \( \tau(30) \in [4, 9] \) (Apr-Sep), for ERA-40. (The choice of these particular two bins will be justified below). The equatorial composite difference \( \Delta \bar{u}_{EQ}(z) \) (i.e., \( \Delta \bar{u}(z) \) at \( \phi = 0 \)) looks qualitatively similar but has roughly twice the amplitude\(^3\); 15°N is chosen because it is closer to latitudes at which the strongest modulation of stratospheric EP-flux convergence by the location of the \( \bar{u} = 0 \) line occurs (e.g., Fig. 3.15), while still displaying a clear QBO signal. Fig. 4.2 indicates that these two phase bins differ by roughly a quarter cycle in QBO phase.

By systematically varying \( \tau(p) \), we determine the form of \( \Delta \bar{u}_{EQ}(z) \) associated

\(^3\)Recall that the QBO amplitude has approximately a Gaussian shape, centered on the equator, with a half-width of \( \phi \approx 12^\circ \); see also Fig. 2.7(a,b).
4. The Holton-Tan effect in CMAM and ERA-40

Figure 4.2: (a) ERA-40 W-E composite difference $\Delta \bar{u}(z)$ at $15^\circ$N for the $\tau(30) \in [9,14]$ (Nov-Apr) phase bin. The dashed line marks the January of the NH winter that is classified as W or E within the bin. (b) As in (a), but for the $\tau(30) \in [4,9]$ (Apr-Sep) phase bin. Contour interval is 2 m/s, with red (blue) indicating positive (negative) values, the zero line in black, and 95% significant differences (by the $t$-test) are shaded.

with significant interannual variability in polar vortex strength. We define the phase bins $\tau(p) \in [\tau_1, \tau_2]$, with $\tau_2 = \tau_1 + \Delta \tau$, and vary $\tau_1$ in 1-month increments while the bin size, $\Delta \tau$, is kept constant. $\Delta \tau$ is specified to be large enough that meaningful statistics can be obtained, but small enough that differences in QBO phase may be distinguished at higher resolution than 0.5 cycles, so as to improve upon the single-level definition of QBO phase. For each phase bin, composites of the W and E groups are made and the W-E composite difference $\Delta \bar{u}(\phi, z)$ is calculated.

Choosing $\Delta \tau = 6$ months, using the $p = 30$ hPa (25 km) level to define QBO phase transitions, and with the $60^\circ$N, 50 hPa (21 km) gridpoint representing the vortex strength, Fig. 4.3(a) shows the variation with changing phase bin (i.e., changing $\tau_1$).

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4 As in Chap. 2, the equatorial radiosonde winds (at $p = 30$ hPa) are used to define the QBO time series, which is first deseasonalized and then smoothed with a 5-month running mean before the time of phase transitions (defined simply as zero crossings) is determined. The results shown in Fig. 4.3 are essentially the same if the radiosonde winds are replaced with ERA-40 $\bar{u}$ and/or if the running mean is not used.

5 It will be shown in Sec. 4.5 that the $60^\circ$N, 50 hPa gridpoint is a suitable proxy for polar vortex
of the $t$-statistic corresponding to $\Delta \bar{u}(\phi, z)$, for different times during NH winter. The early winter (Nov-Dec average) signal peaks for the $\tau(30) \in [9, 14]$ (Nov-Apr) phase bin, while the late winter (Feb-Mar average) signal peaks for the $\tau(30) \in [4, 9]$ (Apr-Sep) phase bin. Note that $\Delta \bar{u}_{EQ}(z)$ corresponding to these two bins was shown in Fig. 4.2. The size of the W and E groups within each bin is shown in Fig. 4.3(c). The $\tau(30) \in [9, 14]$ bin contains 11 W and 9 E winters, while the $\tau(30) \in [4, 9]$ bin contains 8 W and 10 E winters. Provided the data satisfy the assumptions underlying the $t$-test, the thresholds $t > 2.1$ (2.9) indicate roughly the 95% (99%) significance levels [von Storch and Zwiers, 1999]. Hence the statistical significance of the Nov signal exceeds 99% while that of the Feb signal exceeds 95%.

Several features are notable in Fig. 4.3(a). The early and late winter signals peak at different phase bins, suggesting that different forms of $\Delta \bar{u}_{EQ}(z)$ may be associated with the HTE in early and late winter. The Feb peak is separated from the Nov (or Dec) peak by 4-5 bins (i.e., by a change in $\tau$ of 4-5 months), which is shown in Fig. 4.2 to correspond to a difference in $\Delta \bar{u}_{EQ}(z)$ of roughly a quarter cycle in QBO phase. The Mar peak, however, is separated from the Nov peak by only 1-2 bins, indicating a more modest contrast in $\Delta \bar{u}_{EQ}(z)$ between these two cases. Also notable in Fig. 4.3(a) is that as the phase bin shifts later in the year, the Feb signal only strengthens as the Nov signal weakens. This behaviour is consistent with the Nov and Feb signals being induced by differing forms of $\Delta \bar{u}_{EQ}(z)$, but may also indicate that the early winter HTE somehow interferes with the late winter HTE, since evidently the Feb vortex is more sensitive to QBO phase during those winters when the Nov sensitivity is reduced. The Jan signal in Fig. 4.3(a) has a timing and sign similar to that of the Feb-Mar signal, but its significance is much lower. It is possible that the Jan signal, which is fairly broad in $\tau$, partially represents the fading echo of the early winter (Nov, Dec) signal due to the intraseasonal memory (autocorrelation) of the vortex.

We choose the form of $t$ that assumes the two samples, i.e. the W and E groups within any given phase bin, to have equal variances [von Storch and Zwiers, 1999]. We assume that all years are independent realizations, so that the number of years in the two samples gives the degrees of freedom.
Figure 4.3: (a) $t$-statistic corresponding to ERA-40 $\Delta \bar{u}(\phi, z)$ at 60°N, 50 hPa as a function of 6-month ($\Delta \tau = 6$) phase bin. Early winter (Nov-Dec average) and late winter (Feb-Mar average) results are shown as bold dashed lines while the six months of NH winter are shown separately. The month indicated on the abscissa corresponds to the earliest month, $\tau_1 + \Delta \tau$, of each phase bin. The vertical dashed line marks the January of the NH winter that is classified as W or E within the bin (as in Fig. 4.2). (b) Statistical test using $n = 5000$ random white noise values. This plot shows the probability of a given $t$-vs.-bin curve, i.e. one of the curves shown in (a), achieving at least the number of exceedances of the given $t$-values under the null hypothesis that there exists no correlation between QBO phase and polar vortex variability; see text for details. (c) Sizes of the W and E groups within each phase bin shown in (a). For each bin, the first (second) dot indicates the size of the W (E) group.
Finally, no Oct signal is apparent. Although there is an extratropical response during Oct, Sec. 4.5 will show that it does not extend as far poleward as 60°N.

We now consider in more detail the statistical significance of the results shown in Fig. 4.3(a). Note first from Fig. 4.3(c) that the size of W and E groups in different phase bins is not uniform and that some phase bins contain W or E groups with as few as 5 or 6 members. However, these are not the bins that correspond to the strongest signals in Fig. 4.3(a). Deciding what bin size is large enough for meaningful statistics is of course subjective, but the fact that stronger signals occur for larger bin sizes increases confidence in the results.

Another question concerns the likelihood of observing \( t \)-curves similar to those shown in Fig. 4.3(a) under the null hypothesis \( (H_0) \) that polar vortex interannual variability is independent of the QBO. To address this, we remove any causal link between the QBO and the polar vortex by replacing the 60°N, 50 hPa time series with a white noise time series, and repeat the analysis using \( n = 5000 \) white noise processes, counting the number of times that large magnitudes of \( t \) occur. It is found that the likelihood of obtaining extreme values of \( t \) for at least some of the phase bins is higher than the putative significance level of the \( t \)-value. This may be understood as follows: if it were the case that all the phase bins contained independent data, then on average 1 in 20 of the \( t \)-values obtained from a white noise time series would exceed the 95% significance level. However, the data in adjacent phase bins is not completely independent since adjacent bins share some months; this is a consequence of the fact that our analysis partitions the available data into subsamples\(^7\).

We therefore assess the statistical significance of the results shown in Fig. 4.3(a) by asking: what is the likelihood under \( H_0 \) of obtaining multiple exceedances of specified \( t \)-thresholds? Fig. 4.3(b) shows this likelihood for \( n = 5000 \) white noise processes. The Feb-Mar peak in Fig. 4.3(a) contains five exceedances of \( |t| = 2.4 \); according

\(^7\)The situation here is analogous to that which is addressed by field significance testing [von Storch and Zwiers, 1999]. Fig. 4.3(a) displays multiple \( t \)-values, and hence represents the outcome of multiple tests (with each test involving a single \( t \)-value). To properly assess the likelihood that the behaviour shown in Fig. 4.3(a) would occur under \( H_0 \), we must consider the distribution of multiple \( t \)-values under \( H_0 \).
to Fig. 4.3(b) this has only a 4% chance of occurring under \( H_0 \) (i.e. the Feb-Mar peak has a significance of 96%, by this test). The Feb peak, in contrast, contains two exceedances of \(|t| = 2.6\), which has a 12% likelihood under \( H_0 \). The Nov-Dec peak has only a 3% likelihood under \( H_0 \). If the 60°N, 10 hPa (32 km) gridpoint is used instead, the significance of the Nov signal increases while that of the late winter signal remains about the same.

Replacing the white noise processes with AR(1) processes\(^8\) that mimic the intraseasonal autocorrelation of the polar vortex has no appreciable effect on the significance of the results shown in Fig. 4.3(a). However, using AR(1) processes that mimic the interannual autocorrelation of the extratropics does reduce the significance. By performing this test we are assuming that there is a mechanism, independent of the QBO, that induces vortex anomalies to switch signs in consecutive years. In this case the degrees of freedom will be less than the number of years in the two samples (W and E groups). Interannual lag 1-year autocorrelations, \( \rho \), for the NH vortex in ERA-40 are at most \( \rho \approx -0.4 \) in the vicinity of 10 hPa (not shown). Specifying an AR(1) process with \( \rho = -0.4 \), the significances of the Nov-Dec and Feb-Mar peaks are reduced to 92% and 85%, respectively. Of course, if the HTE is real then the observed extratropical \( \rho \) will be largely caused by the QBO itself, in which case this test is overly conservative.

### 4.3.2 CMAM results

The same analysis as in Sec. 4.3.1 for ERA-40 is now performed for the CMAM-QBO run. As for ERA-40, we choose \( \Delta \tau = 6 \) months, use the \( p = 30 \) hPa (25 km) level to define QBO phase transitions, and choose the 61°N, 48 hPa (21 km) gridpoint to represent the vortex strength (this being the CMAM gridpoint that is closest to the 60°N, 50 hPa gridpoint used for ERA-40).

Fig. 4.4(a) shows the CMAM result. The HTE signals for different NH winter months are seen to peak at different phase bins, which is similar to the ERA-40 result.

\(^8\)“AR(1)” denotes a first-order autoregressive process [von Storch and Zwiers, 1999].
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Figure 4.4: (a,b,c) As in Fig. 4.3, but for the CMAM-QBO run. In (a), the $t$-statistic corresponds to $\Delta \bar{u}(\phi,z)$ at the 61°N, 48 hPa gridpoint. Also note that the Feb-Mar average used in Fig. 4.3(a) has been replaced with the Jan-Feb average. In (c), the bin sizes are larger than those in Fig. 4.3(c), due to the larger number of years (150) in the CMAM-QBO run. In (b) the $t$-thresholds and probabilities differ from those in Fig. 4.3(b), due to the larger CMAM bin sizes.
although the bins (i.e., $\tau$-values) corresponding to the peaks are different in CMAM and ERA-40 (cf. Fig. 4.3(a)). The Nov signal peaks for the $\tau(30) \in [3, 8]$ (May-Oct) bin, while Jan and Feb signals occur for later bins: $\tau(30) \in [1, 6]$ (Jul-Dec) and $\tau(30) \in [-4, 1]$ (Dec-May), respectively. The Nov and Feb peaks are separated by 7 bins, while Nov and Jan are separated by only 2 bins. A strong Mar peak also occurs for the $\tau(30) \in [6, 11]$ (Feb-Jul) bin, 3 bins earlier than the Nov peak. A small Oct signal occurs near the Mar peak. The Dec signal is weak, and peaks at a bin intermediate between the Nov and Feb peaks; unlike in ERA-40, the Nov and Dec signals in CMAM do not peak at the same bin, and hence in CMAM the combined Nov-Dec signal follows the Nov signal less closely than in ERA-40 (cf. Figs. 4.3(a) and 4.4(a)). The wide separation of the CMAM Feb and Mar peaks leads to a combined Feb-Mar signal that does not resemble the ERA-40 Feb-Mar peak at all, and therefore is not shown; instead, the Jan-Feb average is used to obtain a more robust late winter HTE signal than is achieved by the Jan or Feb peaks alone. The Jan-Feb average smears together the two distinct Jan and Feb peaks (as was similarly the case with the Feb and Mar peaks in ERA-40).

Owing to the large number of years (150) in the CMAM-QBO run, larger phase bins are possible than for ERA-40 (cf. Figs. 4.3(c) and 4.4(c)). For example, Fig. 4.4(c) shows that the $\tau(30) \in [3, 8]$ (May-Oct) bin, at which the Nov signal peaks, contains 23 W and 37 E winters, while the $\tau(30) \in [-4, 1]$ (Dec-May) bin, at which the Feb signal peaks, contains 30 W and 17 E winters. Hence the peak $t$-values shown in Fig. 4.4(a) have high statistical significance. By the conventional $t$-test, the Nov, Jan and Mar peaks all have 99% significance or higher, while for the other months we have (in descending order): Feb 98.4%, Oct 96.4%, and Dec 93.7%. Using the “multiple exceedances” test of Sec. 4.3.1, the significances are reduced somewhat (as was the case for ERA-40). The probabilities for multiple exceedances of various $t$-thresholds for the CMAM phase bins are shown in Fig. 4.4(b) (as with ERA-40, we replace the vortex time series with $n = 5000$ white noise processes). With this test we find the following significances (again in descending order): Mar 96%, Nov 94%, JF (Jan-Feb average) 93%, Feb 91%, Jan 87%, Oct 77%, Dec 55%. Naturally the significances
are further lowered if we assume the vortex to possess an interannual memory that is independent of the QBO, although there is no particular reason to assume that the CMAM run contains a mechanism that would lead to such a memory.

Some further similarities and differences between the CMAM and ERA-40 results are worth highlighting. The fact that different months’ signals peak at different phase bins suggests that distinct forms of $\Delta \bar{u}_{EQ}(z)$ may be associated with the HTE in CMAM at various times during NH winter. Fig. 4.4(a) shows that, as in ERA-40, a Feb peak occurs at a somewhat later bin (i.e., smaller $\tau$) than the Nov peak, and appears to strengthen only when the Nov peak weakens. Unlike ERA-40, CMAM has a strong Jan signal. Nevertheless, the Jan peak falls between the Nov and Feb peaks, consistent with a tendency – evident from the locations of the Nov and Feb peaks for both CMAM and ERA-40 – for peak HTE signals to occur for progressively later bins (smaller $\tau$) at later times during winter. The Mar peak, however, is an exception to this rule – in both CMAM and ERA-40 (cf. Figs. 4.3(a) and 4.4(a)). Similarly to ERA-40, the Mar peak in CMAM is displaced towards earlier bins (larger $\tau$) relative to the Feb peak, but unlike ERA-40, the Mar peak in CMAM occurs at larger $\tau$ than the Nov peak. Finally, a weak Oct peak, near the Mar peak, is seen in CMAM but not in ERA-40; notwithstanding this weak feature, the lack of a strong Oct HTE at polar latitudes in ERA-40 is echoed in CMAM.

An important similarity between ERA-40 and CMAM becomes evident when the $\Delta \bar{u}_{EQ}(z)$ patterns corresponding to the Nov and Feb peaks in Fig. 4.4(a) are examined. Fig. 4.5 shows the CMAM W-E composite differences $\Delta \bar{u}(z)$ at $\phi = 13^\circ N$, which for brevity’s sake we shall refer to as $\Delta \bar{u}_{EQ}$\textsuperscript{10}. Recalling that the period of the CMAM QBO is 25% longer than the observed QBO period of $\approx 28$ months,

\textsuperscript{9}The SSTs in CMAM have no interannual variability, and external forcings that would be present in ERA-40 – such as GHG changes, volcanoes, or ENSO – are absent in CMAM (Sec. 1.5.1). Thus the only source of interannual memory besides the QBO would be other modes of internal atmospheric variability with timescales longer than a year. One possibility is the low-latitude flywheel mechanism [Scott and Haynes, 1998]. However, lag-1 year autocorrelations in the CMAM control run do not show any significant signal (not shown), suggesting that the low-latitude flywheel is unimportant in CMAM. Hence any interannual memory in the CMAM-QBO run is most likely due to the QBO.

\textsuperscript{10}Similarly to ERA-40, the equatorial difference $\Delta \bar{u}_{EQ}(z)$ is qualitatively similar to $\Delta \bar{u}(z)$ at $\phi = 13^\circ N$, but of larger amplitude.
Figure 4.5: As in Fig. 4.2, but for the CMAM-QBO run. (a) W-E composite difference $\Delta \bar{u}(z)$ at 13°N for the $\tau(30) \in [3, 8]$ (May-Oct) phase bin. (b) As in (a), but for the $\tau(30) \in [-4, 1]$ (Dec-May) phase bin. Contours, colours and shading as in Fig. 4.2.
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The 7-month separation of these two bins is comparable to the 5-month separation seen for the observed QBO, and likewise corresponds to roughly a quarter cycle in QBO phase. Although the CMAM and ERA-40 HTE signals peak at different bins (i.e., different $\tau$), the longer QBO period in CMAM implies that the same $\tau$ will not correspond to the same $\Delta \bar{u}_\text{EQ}(z)$ as for ERA-40. Comparison of Figs. 4.2 and 4.5 shows that the CMAM $\Delta \bar{u}_\text{EQ}(z)$ patterns corresponding to the Nov and Feb HTE signals are strikingly similar to the ERA-40 patterns over the altitude range 25-50 km. This suggests that the polar vortex response to $\Delta \bar{u}_\text{EQ}(z)$ in CMAM is somewhat realistic in both its sign and seasonal timing. Furthermore, the discrepancy between the CMAM and ERA-40 $\Delta \bar{u}_\text{EQ}(z)$ patterns in the lowermost stratosphere, $z < 25$ km, for the late winter signals (cf. Figs. 4.2(b) and 4.5(b)) may indicate that these altitudes are not causal for the late winter HTE.

Given the good comparison between the CMAM and ERA-40 $\Delta \bar{u}_\text{EQ}(z)$ patterns associated with early and late winter HTE signals, it is worth scrutinizing this correspondence in close detail. (The reader who is already convinced of the strong similarity in the 25-50 km layer between the $\Delta \bar{u}_\text{EQ}$ patterns shown in Figs. 4.2 and 4.5 may avoid these details by skipping the next two paragraphs.) Comparing Figs. 4.2(a) and 4.5(a), corresponding to the early winter (Nov) HTE signals, $\Delta \bar{u}_\text{EQ}(z)$ during Oct and Nov in the two datasets is most similar in the mid-stratospheric E layer that lies at $z \approx 30$-40 km. The range of altitudes spanned by the 95%-significant portion of this E difference in ERA-40 (CMAM) is 31.2-44.1 (31.8-41.8) km in Oct and 29.3-40.4 (31.2-39.2) km. The peak magnitudes of the E difference are -20.4 (-26.8) m/s in Oct and -19.9 (-27.5) m/s in Nov. (Note that the actual equatorial differences are larger: -38 (-45) m/s during Nov, about a factor of two greater than the 15°N (13°N) differences.) The altitude of the peak difference is 37.1 (36.0) km in Oct and 34.7 (35.1) km in Nov. Thus the mid-stratospheric E layers of the $\Delta \bar{u}_\text{EQ}(z)$ pattern in both datasets are, during early winter (Oct and Nov), centered at roughly the same altitude and occupy roughly the same depth; the CMAM E layer is shallower by $\approx 3$ km and stronger by $\approx 7$ m/s. Note that the shallower layer in CMAM is consistent with the reduced vertical scale of the CMAM QBO (Fig. 2.7). The early winter lower
stratospheric W layers of the $\Delta \bar{u}_{EQ}(z)$ patterns for ERA-40 and CMAM are also similar, albeit less so than the mid-stratospheric E layers. The greatest W difference in Oct and Nov occurs 2-3 km lower in ERA-40 than in CMAM, and the vertical extent of the W layer in ERA-40 (CMAM) is 16.9-28.8 (19.0-31.1) km in Oct and 16.8-27.3 (19.1-29.6) km in Nov. The peak W difference is 14.0 (14.1) m/s in Oct and 12.3 (15.4) m/s in Nov. Thus the W layer depths and magnitudes in the two datasets also agree fairly well, with the main difference being the upward displacement of the CMAM layer by $\approx 2$ km. Finally, the upper W layers ($z > 42$ km) during Oct and Nov in the two datasets do not agree well.

Considering the $\Delta \bar{u}_{EQ}(z)$ patterns corresponding to the late winter (Feb) HTE signals, shown in Figs. 4.2(b) and 4.5(b), the 95%-significant portion of the mid-stratospheric E layer in ERA-40 (CMAM) spans the altitude range 35.3-41.0 (34.4-43.1) km in Jan and 35.2-42.7 (34.5-43.8) km in Feb. The peak magnitudes are -12.6 (-19.2) m/s at $z = 40.7$ (38.2) km in Jan and -14.1 (-17.0) m/s at $z = 40.7$ (38.2) km in Feb. The mid-stratospheric E layer in CMAM is thus deeper (by 2-3 km) and stronger (by 3-5 m/s) than in ERA-40, and is also displaced slightly downward. The lower W layer in ERA-40 (CMAM) spans the altitude range 18.9-31.3 (24.6-33.0) km in Jan and 18.7-30.4 (24.4-33.3) km in Feb. Its peak magnitudes are 18.5 (24.0) m/s at 26.0 (29.4) km in Jan and 16.3 (23.7) m/s at 26.0 (29.4) km in Feb. Thus the lower stratospheric W layer in CMAM is shallower (by 3-4 km) and stronger (by 6-7 m/s) than in ERA-40, and is displaced slightly upward. The CMAM lower W layer does not extend below $z \approx 25$ km – a feature not observed in ERA-40. Overall, the $\Delta \bar{u}_{EQ}(z)$ patterns corresponding to the late winter (Feb) HTE signals are similar in CMAM and ERA-40 during Jan and Feb in the altitude range 25-50 km, with the closest correspondence occurring for the mid-stratospheric E layer. For $z < 25$ km, $\Delta \bar{u}_{EQ}(z)$ in CMAM and ERA-40 is of opposite sign.

To summarize Secs. 4.3.1 and 4.3.2, the timing of HTE signals during NH winter in CMAM and ERA-40 share some common features. It is especially notable that equatorial wind changes $\Delta \bar{u}_{EQ}(z)$ that are of similar form over the range $z \approx 25-40$ km are associated with early and late winter HTE signals in the two datasets. Fur-
ther discussion of this similarity is deferred until Sec. 4.6. In Sec. 4.4 we use an alternate method to diagnose the dependence of the HTE on $\Delta \bar{u}_{EQ}(z)$, to confirm the robustness of our results.

4.4 QBO phase defined by EOF phase angle

4.4.1 EOF analysis of tropical winds

Sec. 4.3 suggests that it is useful to resolve changes in the vertical structure of tropical winds, $\Delta \bar{u}_{EQ}$, at a precision finer than 0.5 cycles. We may make precise the relation between cycles of QBO phase and $\Delta \bar{u}_{EQ}$ by adopting an EOF representation of the QBO [Wallace et al., 1993; Randel and Wu, 1996; Baldwin and Dunkerton, 1998; Logan et al., 2003]. $\bar{u}_{EQ}$ is expressed as

$$\bar{u}_{EQ}(z,t) = \sum_{i=1}^{n} F_i(z) P_i(t), \ z \in z^{EOF}$$

with $F_i(z)$ the EOFs and $P_i(t)$ the principal component time series. The $F_i(z)$ are found as the eigenvectors of the $n \times n$ covariance matrix $C$ of the time series $\bar{u}_{EQ}(z,t)$, where $n$ is the number of vertical levels [von Storch and Zwiers, 1999]. $\bar{u}_{EQ}(z,t)$ is deseasonalized and then smoothed with a 5-month running mean prior to calculation of $C$. The range of altitudes $z^{EOF}$ over which the functions $F_i(z)$ are defined is arbitrary; for any reasonable choice of $z^{EOF}$ that captures the QBO, the two leading EOF modes $i = (1, 2)$ capture roughly 95% of the variance of the smoothed time series $\bar{u}_{EQ}(z,t)$. Since $F_i$ and $F_j$ are orthogonal for $i \neq j$, the $P_i(t)$ represent projections of $\bar{u}_{EQ}(z,t)$ onto the spatial patterns $F_i(z)$. Following Wallace et al. [1993], we may treat the time series $P_1(t), P_2(t)$ as Cartesian planar coordinates $(x, y)$ and then convert to polar coordinates

$$\psi(t) = \frac{1}{2\pi} \tan^{-1}\left( \frac{y(t)}{x(t)} \right), \ r(t) = \sqrt{x^2(t) + y^2(t)}$$

where $(x, y)$ are identified either as $(P_1, P_2)$ or $(P_2, P_1)$ so as to make the phase angle
\( \psi(t) \) increase with time when QBO phase propagates downward. Apart from QBO stallings, most of the variance of \( \bar{u}_{EQ}(z, t) \) is expressed by the steady increase of \( \psi \) with time, while the phase radius \( r(t) \) varies much more erratically and captures less of the variance. \( \psi \) has been defined to take values in the range \([-0.5, 0.5]\), so that a change over time of \( \Delta \psi = 1 \) indicates that one QBO cycle has elapsed.

### 4.4.2 Contrast between opposing QBO phases

Years may be binned into composite groups defined by a range of \( \psi \) values, which we denote as \( \psi^G \). For example, \( \psi^G \in [-0.5, -0.3] \) indicates that the range \( \psi^G \) includes values of \( \psi \) for which \(-0.5 \leq \psi \leq -0.3\). Two contrasting cases of \( \psi^G \), analogous to the W and E groups of Sec. 4.3, are chosen, and the t-statistic is computed from the difference of means of the two groups. By varying \( \psi^G \) smoothly through the whole range of \( \psi \) (i.e. from -0.5 to 0.5), plots similar to Figs. 4.3(a) and 4.4(a) of Sec. 4.3 are constructed. A series of \( \psi \)-bins, analogous to the phase bins of Sec. 4.3, is created by specifying a series of ranges \( \psi^G \in [\psi_1^G, \psi_2^G] \) (where \( \psi_2^G > \psi_1^G \)). We set \( \Delta \psi^G \equiv \psi_2^G - \psi_1^G \) to a fixed value and increment \( \psi_1^G \) in steps of size \( \delta \psi_1^G \) through the range \([-0.5, 0.5]\).

For example, with \( \Delta \psi^G = 0.2 \) and \( \psi_1^G = -0.3 \), the group \( \psi^G \in [-0.3, -0.1] \) is defined. For each such group, an opposing group \( \tilde{\psi}^G \) is also defined by letting \( \psi_1^G \rightarrow \psi_1^G + 0.5 \), with \( \Delta \psi^G \) unchanged. (For the example \( \psi^G \in [-0.3, -0.1] \), the opposing group would be \( \tilde{\psi}^G \in [0.2, 0.4] \).) A schematic illustration of the definition of these \( \psi \)-bins is shown in Fig. 4.6. The two opposing groups are analogous to the W and E groups of Sec. 4.3: for any given \( \psi^G \), W and E QBO phases occur over altitude ranges \( z^W(\psi^G) \) and \( z^E(\tilde{\psi}^G) \), and since \( \psi^G \) and \( \tilde{\psi}^G \) are separated by 0.5 cycles we have \( z^W(\psi^G) \approx z^E(\tilde{\psi}^G) \) and \( z^E(\psi^G) \approx z^W(\tilde{\psi}^G) \).

We first consider the observed HTE. Using the seven vertical levels (10, 15, 20, 30, 40, 50, and 70 hPa) of the Singapore wind data as the \( \bar{u}_{EQ}(z, t) \) from which \( \psi(t) \) is defined, and taking \( \Delta \psi^G = 0.25 \) and \( \delta \psi_1^G = 0.0125 \), the t-statistic corresponding to the difference of composite means for ERA-40 \( \bar{u} \) at the 60°N, 50 hPa gridpoint is

\[ z^W(\psi^G) \approx z^E(\tilde{\psi}^G) \]
Figure 4.6: Schematic illustration of $\psi$-bin definitions. The QBO phase angle $\psi$ is defined to take values in the range $[-0.5, 0.5]$, and increases in the counterclockwise direction (as indicated). Two examples of specific $\psi$-bins are shown: (a) $\Delta \psi^G = 0.25$, and the two opposing groups are $\psi^G \in [-0.25, 0]$ (i.e., $\psi^G_1 = -0.25$) and $\psi^G \in [0.25, 0.5]$; (b) $\Delta \psi^G = 0.15$, and the two opposing groups are $\psi^G \in [-0.15, 0]$ (i.e., $\psi^G_1 = -0.15$) and $\psi^G \in [0.35, 0.5]$. To create a series of $\psi$-bins, $\Delta \psi^G$ (the angular width of a yellow region) is kept constant while $\psi^G_1$ (the lower boundary of the $\psi^G$ group) is incremented in uniform steps of size $\delta \psi^G_1$ – i.e., the yellow-shaded regions are rotated in the counterclockwise direction, with each new rotation through an angle $\delta \psi^G_1$ defining a new $\psi$-bin. See text for further details.

shown in Fig. 4.7(a). The value $\Delta \psi^G = 0.25$ has been chosen so that the group sizes, shown in Fig. 4.7(b), are similar to those used in Fig. 4.3. The abscissa is labeled by $\psi^G_1$, the lower boundary of one of the groups. Because opposing groups are defined by a separation of 0.5 cycles, a progression of $\psi^G_1$ by 0.5 simply changes the sign of $t$; hence only a range of 0.5 cycles (i.e., $-0.5 < \psi < 0$) is displayed in Fig. 4.7. The value of $\psi$ during Nov is used to define membership in the groups\textsuperscript{12}. (The effect of varying the month used to define $\psi$ will be noted in Sec. 4.6.1.)

Fig. 4.7(a) shows distinct Nov and Feb peaks, separated by roughly 0.25 cycles, as was seen in Fig. 4.3(a). The Feb peak in Fig. 4.7(a) occurs to the left of the Nov peak, indicating that the Feb HTE signal is strongest for values of $\psi$ that are less advanced (i.e., that occur earlier in a QBO cycle) than the values that give the strongest Nov HTE signal. This is consistent with Fig. 4.3(a), in which the Feb peak

\textsuperscript{12}For additional clarity regarding the meaning of $\psi$, the reader may wish to glance ahead to Fig. 4.18(a), which is a scatterplot displaying the relationship between $\psi$ and the equatorial 50 hPa QBO phase (by means of which the HTE has often been defined in the literature, as noted in Sec. 1.2).
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Figure 4.7: (a) t-statistic corresponding to ERA-40 $\Delta \bar{u}(\phi, z)$ at 60°N, 50 hPa for $\psi$-bins defined by Singapore winds. The abscissa is labeled by $\psi^G_1$, the lower boundary of one of the groups. See text for details. (b) Sizes of the opposing groups for each $\psi$-bin shown in (a). Red (blue) indicates the size of the first (second, i.e. opposing) group.

occurred to the right of the Nov peak, because the values taken by $\psi$ during a given NH winter will be less advanced when a QBO phase transition occurs closer to (i.e., at a shorter lag $\tau$ relative to) that winter. The Nov peak in Fig. 4.7(a) is stronger and broader\(^\text{13}\) than the Feb peak, consistent with the relative strengths of the Nov and Feb peaks seen in Fig. 4.3(a). Similar values of $t$ for the Nov and Feb peaks are achieved using both methods (cf. Figs. 4.3(a) and 4.7(a)), and the averaged Feb-Mar signal in Fig. 4.7(a) achieves similar strength to that of the Feb signal alone. Another

\(^{13}\)Note that a portion of the $t > 0$ Nov peak appears as negative values of $t$ on the left side of the plot. If the plot covered the whole range $\psi \in [-0.5, 0.5]$, these values would appear as positive and connected to the right side of the $t > 0$ Nov peak.
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Figure 4.8: Vertical profiles of ERA-40 $\bar{u}(z)$ at 15°N during Nov (a,b) and Feb (c,d) for the $\psi$-bins $\psi^G_1 = -0.05$ (a,c) and $\psi^G_1 = -0.25$ (b,d), which correspond roughly to the Nov and Feb peaks, respectively, in Fig. 4.7(a). Red (blue) indicates the first (second, i.e. opposing) group. The composite mean $\bar{u}(z)$ profiles for the two groups are shown as bold dashed lines (also in red and blue), and the bold dashed black line indicates the difference of the composite means.

feature shared by Figs. 4.3(a) and 4.7(a) is a Dec HTE signal that is coincident with, but weaker than, the Nov signal. Two differences between Figs. 4.3(a) and 4.7(a) are also apparent. The Mar signal in Fig. 4.7(a) is weak, only just barely approaching $t = 2$ for a single $\psi$-bin. Fig. 4.7(a) also shows a weak Jan signal, with two peaks of $t \approx 2$, which did not occur in Fig. 4.3(a). Its shape mimics that of both the Nov and Feb curves, suggesting that it represents a weak echo of the early winter (Nov) HTE and a weak premonition of the late winter (Feb) HTE.

To illustrate the correspondence between $\psi$ and $\bar{u}_{EQ}(z)$, Fig. 4.8 shows ERA-
40 $\bar{u}(z)$ at $15^\circ$N during Nov and Feb for the $\psi$-bins defined by $\psi^G_1 = -0.05$ and $\psi^G_1 = -0.25$, which correspond roughly to the locations of the Nov and Feb peaks (respectively) in Fig. 4.7(a). The $\bar{u}(z)$ profiles in the left column (panels (a,c)) of Fig. 4.8 belong to later stages of their respective QBO cycles than the profiles in the right column (panels (b,d)), indicating that $\psi$ isolates distinct stages of QBO phase progression. The scatter of $\bar{u}(z)$ profiles in Fig. 4.8 is due to the finite group widths ($\Delta \psi^G$). With $\Delta \psi^G = 0.25$, each group contains roughly one-fourth of the available years in ERA-40, and the number of years in each pair of groups ($\psi^G$ and $\tilde{\psi}^G$) for the $\psi$-bins defined by $\psi^G_1 = -0.05$ and $\psi^G_1 = -0.25$ are, respectively, 14 vs. 11 and 11 vs. 9. (These sizes are indicated in Fig. 4.7(b).) The scatter of $\bar{u}(z)$ profiles in Fig. 4.8 would be reduced if a smaller value of $\Delta \psi^G$ were chosen, but too small a $\Delta \psi^G$ will reduce the group sizes to the point where statistically meaningful results cannot be obtained. For the purpose of validating the EOF representation, plots such as those shown in Fig. 4.8 have been constructed for smaller values of $\Delta \psi^G$, for many different values of $\psi^G_1$, at the equator and at $15^\circ$N (not shown); they confirm that the scatter of $\bar{u}(z)$ profiles reduces with decreasing $\Delta \psi^G$. Thus the single variable $\psi$ provides an adequate representation of QBO phase.

Fig. 4.8 indicates the $\bar{u}_{EQ}(z)$ profiles that are associated with the ERA-40 HTE in early and late winter. The differences $\Delta \bar{u}_{EQ}$ (black dashed lines) in Fig. 4.8(a,c) correspond well to the differences shown in Fig. 4.2(a) for Nov and Feb; likewise for the comparison of $\Delta \bar{u}_{EQ}$ between Fig. 4.8(b,d) and Fig. 4.2(b). Fig. 4.8 is complementary to Fig. 4.2 in that it indicates the degree to which the composite mean profiles represent the behaviour of $\bar{u}_{EQ}(z)$ in individual years.

The composite means and individual profiles of $\bar{u}(z)$ at $15^\circ$N shown in Fig. 4.8 provide information about the actual location of the $\bar{u} = 0$ line. (Differences alone, of course, cannot provide this.) Examination of latitudinal profiles $\bar{u}(\phi)$ at various stratospheric altitudes for the months of Oct-Mar (not shown) indicates that in Nov,

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14Recall also that the EOFs were defined on the basis of deseasonalized, 5-month running-mean equatorial (Singapore) winds, whereas Fig. 4.8 shows monthly-average ERA-40 $\bar{u}$ at $15^\circ$N that has not been deseasonalized or otherwise filtered.
\( \bar{u}(\phi = 15^\circ \text{N}) < 0 \) implies that the \( \bar{u} = 0 \) line is located in the NH, where it constitutes the northernmost edge of a deep tropical E layer that extends across the equator and is contiguous with the E winds of the summer hemisphere. We might anticipate that \( \bar{u}(\phi = 15^\circ \text{N}) > 0 \) in Nov likewise implies the existence of a deep tropical W layer that is contiguous with the W winds of the winter hemisphere; however, this is only true at altitudes above 10 hPa. At altitudes below 10 hPa, \( \bar{u}(\phi = 15^\circ \text{N}) > 0 \) in Nov indicates the presence of a weak W or E layer (of peak strength between roughly -5 and 5 m/s) that separates the equatorial QBO-W phase from the extratropical winter W winds. This fact is a consequence of the seasonal cycle. In Oct, and below 10 hPa, there is always a \( \bar{u} = 0 \) line in the NH, regardless of the QBO phase at the equator. This is due to the existence of an E layer in the subtropics below 10 hPa, which is a holdover from the NH summer E winds. From Oct to Nov this subtropical E layer weakens. For QBO-W at the equator below 10 hPa, the subtropical E layer has vanished by Dec. Thus the hemisphere of the \( \bar{u} = 0 \) line below 10 hPa is unaffected by the QBO in Oct and is not uniquely determined by it in Nov; only from Dec onwards does QBO-W at the equator imply a robust shift of \( \bar{u} = 0 \) into the SH at altitudes below 10 hPa.

If shifts in the hemisphere of the \( \bar{u} = 0 \) line are responsible for the HTE, the above facts suggest that the causality of the Nov HTE arises from the QBO at altitudes above 10 hPa. It should be cautioned, however, that the shallow subtropical W or E layer associated with QBO-W in Nov below 10 hPa may influence the propagation of planetary waves, since it provides only a weak barrier between the winter extratropics and the equatorial W waveguide [Chen, 1996]. From Dec onwards, this ambiguity does not exist and the QBO robustly modulates the hemisphere of the \( \bar{u} = 0 \) line, at least up to \( z \approx 40 \text{ km (3 hPa)} \), above which the QBO manifests itself as a modulation of the strength of the SAO-E phase. (The QBO also modulates the strength of the SAO-W phase in Oct and Nov.)

Bearing the above comments in mind, Fig. 4.8 indicates whether or not the QBO robustly modulates the hemisphere of the \( \bar{u} = 0 \) line for each of the \( \psi \)-bins shown. Fig. 4.8(a) shows that the Nov HTE is associated with a robust, deep layer of tropical
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E winds in the \( z \approx 30-35 \) km layer. In contrast, Fig. 4.8(b) shows that for the \( \psi_1^G = -0.25 \) bin, for which only a very weak Nov HTE signal occurs, the distinction between the two opposing groups (\( \psi^G \) and \( \tilde{\psi}^G \), i.e. red and blue groups) in this same layer is ambiguous. Although deep tropical E layers at 30-35 km may occur in the \( \psi_1^G = -0.25 \) bin, they are not the norm (as they are in the \( \psi_1^G = -0.05 \) bin). In the 25-30 km layer, the situation is reversed: in the \( \psi_1^G = -0.05 \) bin (strong Nov HTE) the distinction between opposing groups is ambiguous, while it is fairly well defined in the \( \psi_1^G = -0.25 \) bin (weak Nov HTE). Below \( z \approx 25 \) km, the distinction between opposing groups in both bins is similar. These facts suggest that the 30-35 km layer is causally most significant for the Nov HTE.

For the late winter HTE, Fig. 4.8(d) shows that a strong Feb HTE is found when the two opposing groups are most distinct in the lower stratosphere, for \( z \approx 20-30 \) km. The opposing groups are less distinct from each other in the 30-40 km layer. Fig. 4.8(c) shows that for the \( \psi_1^G = -0.05 \) bin, for which the Feb HTE is very weak (or perhaps absent), the two groups are poorly separated in the middle of the 20-30 km layer, and well separated in the mid-stratosphere. This suggests that the late winter HTE may be most affected by the QBO in the lower stratosphere, where the hemisphere of the \( \bar{u} = 0 \) line is robustly modulated by QBO phase from Dec-onwards. However, Fig. 4.8(d) shows that some separation between the opposing groups also occurs in the layer \( z \approx 35-40 \) km, and the hemisphere of the \( \bar{u} = 0 \) line is also modulated by the QBO at these altitudes. Thus while the distinctions between \( \bar{u}_{EQ}(z) \) profiles shown in Fig. 4.8 might be taken to suggest that late winter HTE causality originates mostly from the QBO at lower stratospheric altitudes, a firm conclusion in this respect is not possible.

The \( \psi \)-bin method is now applied to the CMAM-QBO run. Seven vertical levels in the range \( z \approx 20-32 \) km are used to define \( \psi^{15} \). \( \bar{u} \) averaged over \( 2^\circ S-2^\circ N \) (the two

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\(^{15}\)The high vertical resolution of CMAM – roughly 0.5 km up to the 10 hPa level (Sec. 1.5.1) – allows a large number of vertical levels to be used for the EOF analysis, but this excess of levels is unnecessary. \( \psi \) was found to be virtually identical when seven levels were used as when all vertical levels were used. Hence we use a vertical resolution similar to that of the Singapore winds to define \( \psi \) for CMAM.
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Figure 4.9: As in Fig. 4.7, but for the CMAM-QBO run. (a) $t$-statistic corresponding to $\Delta \bar{u}(\phi, z)$ at 61°N, 48 hPa. (b) Sizes of the opposing groups for each $\psi$-bin shown in (a).

As was seen in Fig. 4.4(a), Fig. 4.9(a) shows distinct Nov, Jan, Feb and Mar HTE peaks. The peaks occur in the same order, with respect to advancing QBO phase, in both Fig. 4.4(a) and 4.9(a). The separation of peaks, however, is altered slightly: the

model gridpoints closest to the equator) is used as the equatorial wind time series $\bar{u}_{EQ}$. All other aspects of the EOF analysis are unchanged from the case of the Singapore winds, including the specification of the $\psi$-bins, with one exception: for CMAM we set $\Delta \psi^G = 0.15$, since this value gives group sizes similar to those used in Fig. 4.4. (The larger number of years in the CMAM-QBO run, as compared to ERA-40, allows finer resolution in $\psi$ while still achieving statistically reliable results.) The variation of the $t$-statistic with $\psi$-bin is shown in Fig. 4.9(a), while Fig. 4.9(b) shows the group sizes (cf. Fig. 4.4(c)).
Feb peak now appears to be separated from the Nov peak by roughly 0.1 cycles, in contrast to the separation of 0.25 cycles that occurred between the strongest Nov and Feb \( t \)-values in Fig. 4.4(a). The Mar peak in Fig. 4.9(a) is also separated from the Nov peak by 0.1 cycles, but in the opposite direction of \( \psi \). Unlike ERA-40, the Mar peak is still present in the \( \psi \)-bin analysis for CMAM. The Jan peak in Fig. 4.9(a) occurs roughly midway between the Nov and Feb peaks, slightly closer to the Feb peak.

Comparison of Figs. 4.9(a) and 4.7(a) shows that the HTE signals in ERA-40 and CMAM are alike in some respects and differ in others. A separation of Nov and Feb peaks, with the Nov peak being stronger than the Feb peak, is common to both datasets. The magnitude of separation differs, being roughly 0.25 cycles in ERA-40 and 0.1 cycles in CMAM. The general tendency for the early winter HTE to weaken for values of \( \psi \) at which the late winter HTE strengthens, and vice versa, is a common feature. The occurrence of a robust HTE in Jan is not: the Jan peak in ERA-40 is much weaker than in CMAM. Likewise, ERA-40 does not exhibit a strong Mar HTE signal while CMAM does.

Another difference between the datasets, not visible in Figs. 4.9(a) and 4.7(a), is that the peak \( t \)-values of the CMAM signals tend to weaken as \( \Delta \psi^G \) increases. E.g. for \( \Delta \psi^G = 0.20 \), the CMAM peaks become broader and flatter (although peak values of \( t \) still lie between 2.5 and 3.0 and hence still indicate high statistical significance). This suggests that the QBO-vortex coupling in CMAM is generally weaker than in ERA-40, in the sense that the HTE signal is only evident for a narrower range of \( \psi \) values. Hence a long model run is required to detect it. Still another difference is found when plots equivalent to Figs. 4.9(a) and 4.7(a), but for \( t \) at the 10 hPa, 60°N gridpoint, are examined (not shown). In both datasets the strength of the Feb HTE signal is similar at 10 hPa and 50 hPa, but in ERA-40 the Nov HTE achieves significantly higher \( t \)-values at 10 hPa than at 50 hPa, while the Nov HTE in CMAM is of roughly the same strength at 10 hPa and 50 hPa.

The \( \bar{u}_{EQ}(z) \) profiles associated with early and late winter HTE signals in CMAM are shown in Fig. 4.10. We choose \( \psi^G_1 = -0.40 \) and \( \psi^G_1 = -0.50 \) to represent the
Figure 4.10: Vertical profiles of CMAM-QBO run $\bar{u}(z)$ at 13°N during Nov (a,b) and Feb (c,d) for the $\psi$-bins $\psi^G = -0.40$ (a,c) and $\psi^G = -0.50$ (b,d), which correspond roughly to the Nov and Feb peaks, respectively, in Fig. 4.9(a). Colours, etc, as in Fig. 4.8.
early and late winter HTE signals, respectively, since these $\psi_1^G$-values correspond to the peak $t$-values for the Nov and Feb signals (Fig. 4.9(a)). Examination of latitudinal profiles $\bar{u}(\phi)$ at various stratospheric altitudes for CMAM (not shown) indicates that the comments made above regarding QBO-modulation of the hemisphere of the $\bar{u} = 0$ line below 10 hPa apply equally well to CMAM as to ERA-40. (This is reassuring, since one would expect a GCM to have a good representation of the seasonal cycle.) Thus Fig. 4.10(a) indicates that a strengthened Nov polar vortex in CMAM, as in ERA-40, is associated with a deep E layer in roughly the $z \approx 30-35$ region. The peak E strength is slightly higher up in CMAM, at $z \approx 34-35$ km in Fig. 4.10(a), rather than at $z \approx 32-33$ km as in Fig. 4.8(a). The contrast between opposing groups in the two $\psi$-bins – comparing panels (a) and (b) of Fig. 4.10 – is, however, less marked in CMAM than in ERA-40. In particular, the $\psi_1^G = -0.50$ bin, which obtains a weak (or nonexistent) Nov HTE signal, also possesses an E layer in the $z \approx 30-35$ region. The layer is, however, weaker (roughly 50% of the amplitude) and less robust (i.e., there is more scatter) than the corresponding E layer for the $\psi_1^G = -0.40$ bin. The reduced contrast between bins is due, of course, to the reduced separation of the Nov and Feb peaks (which is only 0.1 cycles). Hence while the composite mean difference in Fig. 4.10(a) compares well to Fig. 4.5(a), Figs. 4.10(b) and 4.5(b) compare less well: the altitudes of the nodes (of the composite mean difference) are 2-3 km lower in Fig. 4.10(b), consistent with a 0.1-cycle shift (note that a CMAM QBO phase must traverse a vertical distance of roughly 15 km over 0.5 cycles).

The weaker contrast between bins in CMAM indicates that a strong HTE signal only emerges from the background noise when a fairly narrow $\Delta \psi^G$ is chosen. This may indicate, as noted above, that the influence of tropical winds on the polar vortex in the model is unrealistically weak. Nevertheless, the similarity between Figs. 4.10(a) and 4.8(a) is striking. Since the similarities are greatest in the mid-stratosphere – note the greater contrast between ERA-40 and CMAM in the 20-30 km layer – this comparison may be interpreted as further evidence that causality for the Nov HTE originates in the mid-stratosphere.

Fig. 4.10(d), for the $\psi_1^G = -0.50$ bin, shows the form of $\bar{u}_{EQ}(z)$ associated with the
strongest Feb HTE signal in CMAM. Similar to ERA-40 (Fig. 4.8(d)) it shows that a deep W phase in the lower stratosphere is associated with a colder polar vortex in Feb. Note that for \( z < 25 \text{ km} \), the W bias in the CMAM QBO is apparent in Fig. 4.10(d), and hence the two opposing groups do not differ very much in this region. Also similar to ERA-40, Fig. 4.10(d) shows that a mid-stratospheric contrast between the opposing groups exists, but is less clear than in the lower stratosphere. In contrast, Fig. 4.10(c) shows that for the \( \psi_{\text{G}1} = -0.40 \) bin, for which the Feb HTE signal in Fig. 4.9 is very weak, QBO phases in the lower stratosphere are shallower than in the \( \psi_{\text{G}1} = -0.50 \) case. The mid-stratospheric QBO phases in the \( \psi_{\text{G}1} = -0.40 \) bin are deeper and the opposing groups better separated than in the \( \psi_{\text{G}1} = -0.50 \) bin. Thus the \( \bar{u}_{\text{EQ}}(z) \) structures associated with the Feb HTE in CMAM compare well with their ERA-40 counterparts, but the distinction between bins – i.e., between cases in which a Feb HTE is or is not obtained – is less clear than in ERA-40, owing to the smaller separation between Nov and Feb peaks (0.1 cycles, rather than 0.2 cycles).

The smaller separation of Nov and Feb peaks in CMAM suggests that the reason why the Nov and Feb peaks are separated at all is because the HTE mechanism propagates downward with the QBO. That is, we may hypothesize that an influence of tropical winds on the polar vortex is always due to the same form of \( \bar{u}_{\text{EQ}}(z) \) (i.e. the same values of \( \psi \)), and that the timing of the HTE depends on when during NH winter this optimal form of \( \bar{u}_{\text{EQ}}(z) \) occurs. If this is the case, then – recalling that the \( \psi \)-bins used to create Fig. 4.9 were defined by the value of \( \psi \) in Nov – a separation of Nov and Feb peaks will appear in a plot such as Fig. 4.9(a), and their separation will be equal to the typical progression of \( \psi \) that occurs between Nov and Feb. Given an average QBO period of 35 months, the average rate of phase progression is \( d\psi/dt = 1/35 = 0.029 \) cycles/month, or 0.087 cycles in three months – reasonably close to the observed separation of 0.1 cycles. However, examination of time series of \( d\psi/dt \) (not shown) indicates that smaller values of \( d\psi/dt \) are more typical during NH winter. Additionally, a reduction of \( \Delta\psi_{\text{G}} \) to 0.1 cycles (not shown) still yields distinct Nov and Feb peaks, but the separation between the strongest
t-values increases slightly, to 0.125 cycles\textsuperscript{16}. Finally, while panels (a) and (d) of Fig. 4.10 are somewhat similar, there are also differences. The mid-stratospheric E layer in Fig. 4.10(d) exhibits much more scatter than in Fig. 4.10(a), while the depth of the lower stratospheric QBO phases in Fig. 4.10(d) is greater than in Fig. 4.10(a). In summary, these facts are not wholly inconsistent with the hypothesis that downward propagation of the HTE mechanism with QBO phase is responsible for the separation of the Nov and Feb HTE peaks, but they also suggest that this may not be the whole story. Moreover, this hypothesis appears unlikely to explain the separation of Nov and Feb HTE peaks in ERA-40, where the separation (of 0.2 cycles) is roughly double the average rate of QBO phase progression over a three-month period ($d\psi/dt = 1/28 = 0.036$ cycles/month, or 0.11 cycles over three months). However, it is possible that the larger group sizes (e.g. $\Delta\psi_G = 0.25$ in Fig. 4.7), which are required for ERA-40 because of the shorter record, result in a smearing out of the two peaks that obscures their true separation. Given the length of the ERA-40 record, it is not possible to resolve this issue in a statistically reliable manner.

### 4.4.3 Contrast between QBO phase and all other years

In Sec. 4.4.2 the HTE was diagnosed as the difference of the composite means of two groups, $\psi_G$ and $\tilde{\psi}_G$, separated by 0.5 cycles. In this section we keep $\psi_G$ the same, but redefine the second group $\tilde{\psi}_G$ as containing all other years in the data that are not included in $\psi_G$. Similar results are found if $\tilde{\psi}_G$ is defined as the climatology (i.e., all years in the data)\textsuperscript{17}. However, in this case the distribution of the $t$-statistic (corresponding to the $\psi_G - \tilde{\psi}_G$ composite mean difference) will not have the standard $t$-distribution\textsuperscript{18}. While there is nothing wrong with $t$ having a different distribution,

\textsuperscript{16}For $\Delta\psi_G = 0.1$, the groups contain 10-20 members each; this is less than for the $\Delta\psi_G = 0.15$ case shown in Fig. 4.9, but is still a statistically reasonable number.

\textsuperscript{17}In particular, the results are extremely similar for CMAM, since $\Delta\psi_G = 0.15$ implies that only 10-20\% of the available years belong to any given group $\psi_G$ – as shown in Fig. 4.12(b). For ERA-40, $\Delta\psi_G = 0.25$ implies that 20-30\% of the available years belong to any given $\psi_G$, as shown by Fig. 4.11(b).

\textsuperscript{18}This fact has been verified by direct Monte Carlo simulation of the distributions of $t$ for normally distributed white noise (not shown). It occurs because the two samples share some degrees of freedom – since the $\psi_G$ years are also included in the $\tilde{\psi}_G$ group (the climatology) – and are therefore not
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Figure 4.11: (a) $t$-statistic corresponding to ERA-40 $\Delta \bar{u}(\phi, z)$ at 60°N, 50 hPa for $\psi$-bins defined by Singapore winds. Unlike Fig. 4.7, here the mean of a single group is subtracted from the mean of all remaining years. As in Fig. 4.7, $\Delta \psi^G = 0.25$, and the abscissa is labeled by $\psi^G_1$, the lower boundary of the single group. (b) Size of the single group for each $\psi$-bin shown in (a).

It is more convenient for it to have the standard $t$-distribution.

The results for ERA-40 are shown in Fig. 4.11. It is seen that the largest values of $|t|$ correspond to negative vortex anomalies. For the Feb signal, this may be due to the fact that SSWs represent the skewed tail of the distribution of polar vortex $\bar{u}$. The negative Nov and Feb HTE signals occur at distinct values of $\psi$, separated by roughly 0.2 cycles, echoing the $\psi$-separation of Nov and Feb peaks seen in Fig. 4.7. The positive anomalies, which exhibit weaker $|t|$ values, occur for a wider range of $\psi$ values, and no separation between different months’ HTE signals is apparent.
Figure 4.12: As in Fig. 4.11, but for the CMAM-QBO run. (a) $t$-statistic corresponding to $\Delta \bar{u}(\phi, z)$ at $61^\circ N, 48$ hPa. (b) Size of the single group for each $\psi$-bin shown in (a). As in Fig. 4.9, $\Delta \psi^G = 0.15$.

Nevertheless, Fig. 4.11 verifies that choosing two opposing groups to have a separation of $\Delta \psi^G = 0.5$ will yield a strong HTE signal, thus validating the choice of $\Delta \psi^G$ used in Sec. 4.4.2.

The corresponding CMAM results are shown in Fig. 4.12. In contrast to ERA-40, the largest $|t|$ values occur for positive vortex anomalies. This may be due to the CMAM polar vortex being unrealistically weak, with SSWs occurring more often than the observed frequency (which is 0.62 events per year [Charlton and Polvani, 2007]). Negative CMAM vortex anomalies correspond to weaker $|t|$ values. Distinct early and late winter peaks occur for both positive and negative $t$. The separation between positive and negative peaks is roughly 0.5 cycles, verifying that as with ERA-40, the
4.5 Seasonal progression of the Holton-Tan effect

In Secs. 4.3 and 4.4 it was shown that distinct early and late winter HTE signals occur for different alignments of QBO phase with respect to the annual cycle. The method employed to demonstrate this fact, in which the $t$-statistic was plotted as a function of phase bin or $\psi$-bin, allowed for efficient determination of the optimal HTE signal. It was, however, limited to a single gridpoint, and thus was unable to indicate the meridional and vertical structure of the extratropical response. Reliance on $t$ also obscures the actual amplitude of $\Delta \bar{u}(\phi, z)$, since $t$ is a normalized measure of variability. In this section, zonal cross sections of $\Delta \bar{u}(\phi, z)$ are shown for selected subsamples that optimize the early and late winter HTE signals.

Two $\psi$-bins from Sec. 4.4.2 for opposing groups $\psi^G$ and $\tilde{\psi}^G$, analogous to the W and E composite groups defined in Sec. 4.3, are chosen here as representative of the early and late winter extratropical responses to QBO phase. (We might equally well have chosen two phase bins from Sec. 4.3; the results are not sensitive to this choice.) Fig. 4.13(a) displays the Oct-Mar progression of ERA-40 $\Delta \bar{u}(\phi, z)$ for the $\psi$-bin defined by $\psi^G_1 = -0.05$, which gives a strong early winter HTE signal (Fig. 4.7(a)); corresponding values of the $t$-statistic are shown in Fig. 4.13(b). The extratropical response is seen to extend poleward into the lower stratosphere by Nov, extending also through the depth of the troposphere. The same pattern of tropospheric anomaly is retained throughout the Oct-Mar period, but high statistical significance is achieved only in Nov (Fig. 4.13(b)). A strong polar stratospheric signal persists as late as Jan, moving poleward and downward over the Dec-Feb period; although a positive polar anomaly is retained in Feb and Mar, its significance is low (Fig. 4.13(b)). A highly significant subtropical stratospheric anomaly is also seen, weakening in Feb-Mar as does the polar signal, indicating that the QBO-induced anomalies project onto the annular mode of extratropical variability [Baldwin and Dunkerton, 1999]. It is also notable that a downward progression of the signal is evident over the Dec-Jan period, but
Figure 4.13: Oct-Mar progression of the ERA-40 HTE signal for a $\psi$-bin that gives a strong early winter HTE (the bin denoted by $\psi^G_1 = -0.05$ in Fig. 4.7). (a) $\Delta \bar{u}(\phi, z)$, contour interval: 2 m/s. (b) $t$-statistic corresponding to (a), contour interval: 0.5 (up to maximum $|t|$ of 10). Red (blue) = positive (negative).
Figure 4.14: Oct-Mar progression of the CMAM HTE signal for a $\psi$-bin that gives a strong early winter HTE (the bin denoted by $\psi_1^G = -0.40$ in Fig. 4.9). (a) $\Delta \bar{u}$. (b) $t$-statistic corresponding to (a). Contours and colours as in Fig. 4.13.
not the Nov-Dec period. This may be consistent with faster downward propagation of annular mode anomalies in Nov than in later months of NH winter [Baldwin and Dunkerton, 1999], in which case the highly significant tropospheric HTE signal in Nov (Fig. 4.13(b)) represents the projection of the QBO onto the mode of stratosphere-troposphere coupling that seems to be well captured by the annular mode description of variability.

The corresponding CMAM early winter HTE signal is shown in Fig. 4.14. As in ERA-40, a vertically deep polar $\Delta \bar{u}(\phi, z)$ is evident in Nov, although it lacks the strong connection to the troposphere seen in ERA-40. Also similarly to ERA-40, subtropical anomalies of high significance (Fig. 4.14(b)) accompany the early winter polar anomalies, and weaken in late winter concurrently with the weakening of the polar anomalies. This latitudinal dipole structure of $\Delta \bar{u}(\phi, z)$, as well as its vertical depth, suggests that the modelled HTE also projects onto an annular mode of extratropical variability. Unlike in ERA-40, a downward movement of the polar anomaly is evident in the Nov-Dec period. Finally, although the general pattern of development is reminiscent of the ERA-40 signal, the peak value of the polar Nov HTE signal in CMAM is roughly 2 m/s weaker than the ERA-40 Nov signal. The contrast between the CMAM and ERA-40 polar HTE signals in Dec is more extreme: although of similar magnitude in the lower stratosphere, the ERA-40 Dec signal strengthens with increasing altitude while the CMAM Dec signal decays.

For a $\psi$-bin that optimizes the late winter HTE, Fig. 4.15 shows the ERA-40 response. The subtropical anomaly in Oct is similar to that in the early winter case (Fig. 4.13), but a poleward extension is lacking. In Nov, similarly, a highly significant subtropical upper stratospheric anomaly (Fig. 4.15(b)) is not accompanied by a strong poleward extension (cf. Figs. 4.13 and 4.15). Starting in Dec, a polar anomaly is seen to propagate poleward and downward over the Dec-Mar period. The signal has high statistical significance in the upper stratosphere in Dec and in the lower stratosphere in Feb (and hence is consistent with the Feb signal shown in Fig. 4.7 for the $60^\circ$N, 50 hPa gridpoint). The Jan anomaly has lower significance, but is otherwise consistent with the Dec-Mar pattern of development. Note also that the subtropical anomalies
Figure 4.15: As in Fig. 4.13, but for a \( \psi \)-bin that gives a strong late winter HTE (the bin denoted by \( \psi_1^G = -0.25 \) in Fig. 4.7). (a) \( \Delta \bar{u}(\phi, z) \). (b) \( t \)-statistic corresponding to (a). Contours and colours as in Fig. 4.13.
Figure 4.16: As in Fig. 4.13, but for a ψ-bin that gives a strong late winter HTE (the bin denoted by $\psi_1^G = -0.50$ in Fig. 4.9). (a) $\Delta \bar{u}$. (b) $t$-statistic corresponding to (a). Contours and colours as in Fig. 4.13.
during Dec-Mar are weaker and have lower statistical significance than in the early winter case, with weaker tropospheric extensions (cf. panels (b) of Figs. 4.13 and 4.15).

The development of the late winter CMAM HTE signal is shown in Fig. 4.16. As in ERA-40, subtropical stratospheric anomalies during Oct and Nov bear some similarity in the early and late winter cases (cf. Figs. 4.14 and 4.16), but poleward extensions of these anomalies do not occur in Fig. 4.16. Polar anomalies, accompanied concurrently by opposite-signed subtropical anomalies, develop in Jan and Feb; both the polar and subtropical anomalies have high significance (Fig. 4.16(b)). The peak value of the Jan anomaly is roughly 2 m/s weaker than the ERA-40 peak Jan anomaly, and the peak Feb anomaly is roughly 6 m/s weaker than the ERA-40 peak Feb anomaly. As in ERA-40, only a weak connection to the troposphere is evident in the late winter HTE case.

In summary, these results show that the diagnostics employed in Secs. 4.3 and 4.4, while relying on a single gridpoint, are nevertheless indicative of QBO influence over the entire extratropical stratosphere. For ERA-40 in Nov, QBO influence also extends into the troposphere with high statistical significance. Similar patterns of development in the stratosphere are seen to occur in both ERA-40 and CMAM, albeit with some discrepancies in the exact timing and amplitude of the responses.

4.6 Discussion

4.6.1 Causality of the Holton-Tan effect in early and late winter

A large number of mechanistic modelling studies have addressed the HTE by imposing various forms of idealized single-layer $\Delta \bar{u}_{EQ}$, as discussed in Chap. 1 and at the beginning of this chapter. Since $\Delta \bar{u}_{EQ}$ is an imposed, rather than spontaneous, feature of the models employed in such studies, the causal attribution of changes in polar vortex strength and/or variability due to the imposed $\Delta \bar{u}_{EQ}$ is straightforward. This approach has been used to examine the sensitivity of the polar vortex to imposed
\( \Delta \bar{u}_{EQ} \) at different altitudes [O'Sullivan and Young, 1992; Gray, 2003; Gray et al., 2004], in an attempt to discern whether HTE causality originates from the QBO at lower or upper stratospheric altitudes.

When the observed HTE is considered, or when the QBO in a model (either spontaneous or imposed) is allowed to propagate downward, rigorous attribution of HTE causality to a specific single layer of tropical winds is not possible. For the observed HTE, inferences about causality must of course be based only on observed correlations. With a GCM study it is possible to attribute climatological changes in the model to the occurrence of a QBO, whether the QBO is spontaneous (as in Chap. 3) or prescribed (e.g. Hamilton [1998]). But this does not address the question of which tropical layer is most important for HTE causality. We may nevertheless take a diagnostic approach and attempt to infer the likely origin of HTE causality from statistical analyses of both observations and CMAM. The comparison between ERA-40 and CMAM in this chapter represents an attempt to do that, and we now draw some conclusions from the results.

It is natural to first question the very assumption that a single layer of tropical winds is the causal agent of the HTE. Why not instead the whole depth of the tropical stratosphere? It is in fact likely that a GCM requires realistic tropical winds over the whole stratospheric depth in order to obtain a realistic HTE [Gray, 2003; Pascoe et al., 2006]. It is difficult, however, to diagnostically assess the causal role of W or E perturbations \( \Delta \bar{u}_{EQ} \) occurring simultaneously at different altitudes. The most widely accepted theory of the HTE stipulates that modulation of the hemisphere of the \( \bar{u} = 0 \) line by the QBO affects the propagation of stationary planetary waves by moving their critical lines away from or towards the polar vortex [Holton and Tan, 1980], as discussed in Chap. 3. This mechanism is only unambiguous when a single-layer \( \Delta \bar{u}_{EQ} \) is assumed, since oppositely-signed \( \bar{u}_{EQ} \) perturbations at different altitudes

\[ \text{\footnotesize\textsuperscript{19}} \text{The more general issue for models is whether they can simulate realistic interannual variability of the polar vortex, and this is likely to depend on other factors besides the QBO-vortex interaction. Nevertheless, the strength of the observed HTE suggests it is probably true that a realistic polar vortex simulation requires a realistic HTE. The present discussion will focus only on the issue of obtaining a realistic HTE.} \]
may equally well be supposed to affect the vortex. An intriguing possibility is that opposite-signed QBO phases at different altitudes do in fact exert opposite influences on the vortex and that the observed HTE represents the outcome of this tug-of-war – but this notion is highly speculative. We therefore follow the philosophy of mechanistic studies by regarding an understanding of the effects of a single-layer $\Delta \bar{u}_{EQ}$ as a necessary prerequisite to understanding the more complicated (and realistic) multiple-layers scenario.

Causality in early winter

We first consider the HTE in Nov. It was noted in Sec. 4.4.2 that QBO-modulated changes in the hemisphere of the $\bar{u} = 0$ line do not appear below 10 hPa in Oct. In Nov, and below 10 hPa, it was noted that the relationship between QBO phase and the hemisphere of the $\bar{u} = 0$ line is ambiguous. Above $z \approx 45$ km in Oct and $z \approx 40$ km in Nov, the QBO signal appears as a modulation of the SAO-W phase, and hence the $\bar{u} = 0$ line is in the SH at these altitudes during both QBO phases. These facts suggest that $\Delta \bar{u}_{EQ}$ in roughly the 30-40 km layer is probably the cause of the Nov HTE. This notion is supported by comparison of the $\bar{u}_{EQ}(z)$ structures associated with the Nov HTE in both ERA-40 and CMAM, since the strongest similarities between $\bar{u}_{EQ}(z)$ in the two datasets, in Nov, were found in the 30-40 km layer (Sec. 4.3) or the 30-35 km layer (Sec. 4.4.2). In particular, it is notable that a Nov HTE occurs in CMAM despite the fact that the QBO in the model is grossly unrealistic for $z < 25$ km (due to the W bias at these altitudes, discussed in Chap. 2). This fact suggests that the QBO in the lowermost stratosphere does not affect the Nov HTE. However, a note of caution is warranted, as Sec. 4.5 showed that the amplitude of the Nov HTE in CMAM is approximately half that of the Nov HTE in ERA-40. It is possible that the weak HTE amplitude in CMAM is related to deficiencies in the simulated QBO (although of course it could also be due to other model errors\textsuperscript{20}). Finally, it was

\textsuperscript{20}In Chap. 3 it was shown that the climatological strength of equatorward planetary wave propagation is weaker in CMAM than in ERA-40. This may result from unrealistic aspects of the extratropical stratospheric planetary waveguide that are unrelated to the QBO. For example, the tropospheric jets in CMAM are unrealistically strong (at their peak values) by $\approx 5$ m/s, a dis-
noted in Chap. 3 that the largest changes in $\nabla \cdot \mathbf{F}$ at subtropical latitudes, in terms of acceleration of the zonal flow, arose in the 30-40 km region. Changes in $\nabla \cdot \mathbf{F}$ at lower altitudes become more prominent from Dec onwards. In summary, all of these lines of evidence point to the causal efficacy of the 30-40 km $\Delta \bar{u}_{\text{EQ}}$ being the simplest explanation for the Nov HTE.

It is worth commenting briefly on the possibility that the above hypothesis is wrong and that it is in fact the lower QBO altitudes – roughly, the 20-30 km layer – that are causal for the Nov HTE. For this to be the case, planetary wave propagation and dissipation must exhibit a sensitivity to the two contrasting cases of subtropical $\bar{u}$ below 10 hPa that were described in Sec. 4.4.2, namely: QBO-E, during which the $\bar{u} = 0$ line represents the northernmost boundary of a deep tropical E layer that stretches across the equator, and QBO-W, during which a weak layer of W or E $\bar{u}$ separates the equatorial QBO-W phase from the winter westerlies (and for which the $\bar{u} = 0$ line may or may not be in the NH). Using a barotropic model, Chen [1996] showed that planetary wave breaking was sensitive to QBO phase only when $\bar{u}$ at 30°N was positive and sufficiently large (as noted in Sec. 1.3.1). If this conclusion applies in a three-dimensional atmosphere, it suggests that the 20-30 km layer cannot be causal for the Nov HTE. On the other hand, it is possible that planetary waves are able to tunnel through a weak subtropical barrier to reach the W winds that exist at the equator during QBO-W [Harnik, 2002], and in this case the Nov HTE may be affected by QBO phase in the 20-30 km layer. Examination of CMAM latitudinal profiles $\bar{u}(\phi)$ for QBO-W and QBO-E phases in the 20-30 km layer during Nov (not shown) indicates that a deep tropical E layer is associated with stronger W winds at subtropical latitudes poleward of the $\bar{u} = 0$ line than is the case for a weak W or E subtropical layer; this suggests that QBO phase in the 20-30 km layer does in fact affect planetary wave propagation and dissipation in the subtropics at these altitudes. However, it is also seen that this effect occurs in phase bins that do give a Nov HTE discrepancy that is of large magnitude relative to the interannual variability of the jet strength. The behaviour of the tropospheric jets is determined by the synoptic-scale waves that, in turn, may well affect the $\bar{u}$ distribution in the lowermost extratropical stratosphere [Thorncroft et al., 1993].
as well as in those that do not. Since the effect of QBO phase on subtropical $\bar{u}$ in the (roughly) 20-30 km layer is similar in both cases – i.e., whether or not a Nov HTE occurs – this suggests that the influence of 20-30 km QBO phase on the Nov HTE is unimportant.

The above inferences may be compared to results from mechanistic modelling studies. It is notable that most such studies have imposed perpetual January radiative conditions, which may limit the applicability of their findings in regard to the Nov HTE. A robust finding of mechanistic studies, noted in Sec. 1.3.1, is that the timing of SSWs is sensitive to QBO phase for an intermediate range of planetary wave forcing amplitudes [O’Sullivan and Salby, 1990; Holton and Austin, 1991; O’Sullivan and Young, 1992], but SSWs are extremely rare in Nov. Nevertheless, we note that Gray et al. [2004] found that the most pronounced control of $\Delta \bar{u}_{EQ}$ over polar vortex behaviour was obtained when an easterly $\bar{u}_{EQ}$ perturbation was imposed in the 30-40 km layer (see their Fig. 4). The response of the vortex was a robust initial cooling (reminiscent of the Nov HTE), followed by the robust occurrence of warmings. Imposed easterly $\bar{u}_{EQ}$ perturbations at other altitudes also produced warmings, but with more variable timing. (The case of a 25-35 km E layer also produced some regularity of warmings, but to a lesser degree than the 30-40 km case.) The duration of the initial cooling in their 30-40 km case was only 20-30 days, but they noted that the rapid adjustment of the model to imposed January radiative conditions (starting from initial conditions typical of August) may have shortened the duration of the “early winter” portion of their simulations.

Contrary to the above argument, Pascoe et al. [2006] concluded that the QBO below 10 hPa (32 km) mainly influenced polar variability in early winter, while the QBO at higher altitudes became important for late winter polar variability. Using a dynamics-only GCM with a realistic seasonal cycle but no spontaneously occurring QBO, they used a relaxation method (following Hamilton [1998]) to impose two different types of downward-propagating QBO with a 27-month period: one occupying roughly the 20-30 km range, and the other extending over the 20-60 km range. A realistic SAO was also imposed, so that the 20-60 km QBO case yielded a QBO mod-
ulation of the SAO (consistent with the nature of the true QBO signal at \( z > 40 \) km). Compared to their control simulation, in which neither a realistic SAO nor QBO was imposed, interannual variability of North Pole temperature at 50 hPa during Nov was increased when both the SAO and the 20-30 km QBO were imposed. Late winter variability was affected by imposition of the SAO, but unchanged by the addition of the 20-30 km QBO. When the SAO and the 20-60 km QBO were imposed, and compared to the SAO + 20-30 km QBO case, Nov variability was similar but Jan variability was enhanced. On this basis they concluded that the 20-30 km QBO mainly affects early winter variability.

We suggest the following mechanism to potentially reconcile the Pascoe et al. [2006] result with the hypothesis that the 30-40 km QBO is the cause of the Nov HTE. ERA-40 composite W-E differences shown in both Figs. 4.1 and 4.13(a) indicate that \( \Delta \bar{u}_{EQ} < 0 \) in the 30-40 km layer is associated with a subtropical \( \Delta \bar{u} < 0 \) anomaly that extends downward from the 30-40 km region through the 20-30 km region, with a magnitude of \( \approx -2 \) m/s, which is large relative to the interannual variability in this region, particularly in Nov. (This fact is indicated by the high statistical significance of the \( t \)-statistic for the Nov difference, shown in Fig. 4.13(b).) Early winter subtropical \( \nabla \cdot \mathbf{F} \) is stronger, in terms of acceleration, in the 30-40 km layer than in the 20-30 km layer (not shown). Offline calculations of the \( \bar{u} \) response to an imposed zonal torque [Plumb, 1982; Garcia, 1987] indicate that this \( \nabla \cdot \mathbf{F} \) anomaly is sufficiently strong to cause the downward-extending subtropical \( \Delta \bar{u} < 0 \) anomaly, while the \( \nabla \cdot \mathbf{F} \) anomaly in the 20-30 km layer is not (not shown). This suggests that the subtropical \( \Delta \bar{u} < 0 \) anomaly in the 20-30 km layer is caused by \( \Delta \bar{u}_{EQ} < 0 \) in the 30-40 km layer. The effect of this anomaly will be to inhibit the ability of the QBO-W phase in the 20-30 km layer to move the \( \bar{u} = 0 \) line across the equator in Nov. Hence we speculate that the QBO in the 30-40 km layer may inhibit the early winter influence of the QBO in the 20-30 km layer by delaying the occurrence of robust QBO-modulated \( \bar{u} = 0 \) line shifts in the NH subtropical 20-30 km layer until after Nov. Thus the effect of the 20-30 km layer QBO on Nov variability in Pascoe et al. [2006] may perhaps be an artefact of their model lacking a QBO in the 30-40 km
region. This hypothesis is speculative, but it is consistent with the available facts.

Causality in late winter

From Dec onwards, QBO phase robustly modulates the hemisphere of the $u = 0$ line over the whole altitude range 20-40 km. Sec. 4.3 indicated that different forms of $\Delta u_{EQ}$, separated by roughly a quarter of a QBO cycle, are associated with the Nov and Feb HTE signals. Sec. 4.4.2 upheld this conclusion for ERA-40, while for CMAM it indicated that the $\psi$-separation between Nov and Feb HTE signals may be only 0.1 cycles. Additionally, Secs. 4.3 and 4.4.2 both showed the existence of Jan and Mar HTE signals in CMAM, both of them separated from the $\psi$ that gives the Nov HTE. A robust feature, shared by ERA-40 and CMAM, was seen to be the weakness or absence of a Nov HTE signal at values of $\psi$ for which late winter (Jan-Mar) HTE signals are found. The existence of ranges of $\psi$ for which the Nov HTE is weak or absent is itself a novel finding, one that to the best of the author’s knowledge has not yet been reported in the literature.

Thus the results seem to indicate that the $\Delta u_{EQ}$ that causes the late winter HTE differs from that which causes the Nov HTE, but the true magnitude of this difference (in terms of $\psi$) is not entirely clear. Nor is there an obvious indication as to which single layer (if any) might be causally responsible for the late winter HTE, given the occurrence of QBO-modulated shifts of the $u = 0$ line over most of the stratosphere. A straightforward possibility is that the deep W or E layer that occurs in the 20-30 km region for $\psi$-bins that give a Feb HTE (Fig. 4.8(d)) is required to affect the timing of SSWs, in conformity with the findings of mechanistic studies that imposed $\Delta u_{EQ} < 0$ over roughly a 15 km depth in the lower stratosphere [Holton and Austin, 1991; O’Sullivan and Young, 1992] and with the hypothesis of McIntyre [1982]. However, it is unclear why an equally-deep shift of the $u = 0$ line in the 25-35 km layer (Fig. 4.8(c)) should not also lead to a Feb HTE. Additionally, comparison of panels (c) and (d) of Fig. 4.10 indicates that for CMAM, the $\Delta u_{EQ}$ structures that yield Nov and Feb HTE signals are more similar to each other than is the case in ERA-40, further begging the question as to why a Feb HTE is only observed in
one of these cases. A number of studies have argued that $\Delta \bar{u}_{EQ}$ over the whole stratospheric depth – that is, above 10 hPa as well as below it – is important for the HTE in mid or late winter [Gray et al., 2001a; Gray, 2003; Gray et al., 2004]. This strongly argues against making an assumption that a deep QBO phase in the 20-30 km layer is the main cause of the late winter HTE. Such may be the case, but no singularly compelling reason to believe that it is has emerged from the present results or in the literature.

We therefore discuss some other hypotheses. One of these is the idea already mentioned that the HTE mechanism is most effective for a particular range of $\psi$, which we will denote symbolically as $\psi^{HTE}$, and that the $\psi$-separation of Nov and Feb HTE peaks is due to the progression of QBO phase into the state $\psi \in \psi^{HTE}$ in Nov for the Nov HTE signal, and in Feb for the Feb HTE signal. The results of Secs. 4.3 and 4.4.2 are only partially consistent with this hypothesis, as discussed in Sec. 4.4.2. However, a slight modification of the idea may be fruitful: if it is assumed that $\psi^{HTE}$ is a function of the seasonal cycle, perhaps because of the seasonal variation of planetary wave amplitude as well as radiative forcing strength, then the $\psi$-separation of Nov and Feb HTE signals would be equal to the difference in $\psi^{HTE}$ between early and late winter, plus the typical change in $\psi$ between Nov and Feb. If $\psi^{HTE}$ decreases (i.e. shifts to “earlier” values in a QBO cycle) from Nov to Feb, this could explain why the $\psi$-separation of Nov and Feb HTE signals appears to be larger than the typical change in $\psi$ between Nov and Feb (as was noted in Sec. 4.4.2). A change in $\psi^{HTE}$ between early and late winter may be consistent with the apparent sensitivity of the polar vortex to $\Delta \bar{u}_{EQ}$ over the whole stratospheric depth that has been explored in various studies by L. J. Gray and collaborators. Additionally, a robust conclusion shared by almost all mechanistic modelling studies (as noted in Sec. 1.3.1) is that the vortex only responds to QBO phase for an intermediate range of planetary wave forcing amplitudes. That is, there exist two regimes – that of weak vortex / strong waves, and that of strong vortex / weak waves – in which the HTE does not function. Mechanistic studies have generally assumed perpetual January radiative conditions, as described in Sec. 1.3.1 (with O’Sullivan and Dunkerton [1994] being a notable exception), and
the strength of planetary wave forcing that is required to overcome the “radiative spring” and prevent the vortex from assuming a cold and undisturbed state must be determined by the strength of radiative forcing. It is therefore not unreasonable to assume that the vortex strength that lies in the “intermediate” forcing range may have a seasonal dependence. Given this assumption, there is no particular reason to assume that \( \psi_{HTE} \) remains unchanged throughout the winter. However, beyond attempting to infer the change in \( \psi_{HTE} \) between Nov and Feb from the diagnostic results in Secs. 4.3 and 4.4.2, we do not have any other basis for estimating \( \psi_{HTE} \) beyond direct numerical simulation (potentially a subject for future work)\(^{21}\).

Given the speculative nature of the above hypothesis regarding seasonal variation in \( \psi_{HTE} \), we consider now a potentially simpler hypothesis, for which motivation is provided by several findings from this chapter. Comparison of panels (c) and (d) of Fig. 4.8, as noted above, raises the question as to why a Feb HTE in ERA-40 does not occur when there is a strong contrast between opposing QBO phases in the 25-35 km layer. Comparison of panels (c) and (d) of Fig. 4.10 showed that there is not a large difference between the \( \psi \)-bins that do and do not give a Feb HTE in CMAM. For the Jan HTE in CMAM, Fig. 4.9(a) indicates that the \( \psi \)-separation of Jan and Nov HTE signals is even smaller (0.05-0.075 cycles) than that between Feb and Nov (0.1 cycles), and although there is a weak Jan HTE at the same \( \psi \) that gives the Nov HTE \( (t \approx 1.7) \), it does not represent the peak Jan HTE strength \( (t \approx 3) \). It is also notable that the Nov HTE signal is weak at values of \( \psi \) for which strong late winter HTE signals are found (in both ERA-40 and CMAM). Finally, one further piece of evidence, not yet discussed in this chapter, is of interest. In the course of attempting to optimize the HTE signals of Sec. 4.4.2, it was found that the strength of the Jan and Feb peaks in CMAM decreased as the month in which \( \psi \) was defined advanced from Nov to Feb. This is surprising, since one might intuitively expect the QBO-vortex correlation to be maximal between concurrent QBO and vortex states;

\(^{21}\)An analytical theory describing the dependence of (nonlinear) planetary wave forcing on the seasonally-varying stratospheric basic state would presumably provide such an estimate. Unfortunately, such a theory has not been discovered.
the existence of a 2-3 month lag is perplexing.

We attempt to explain the above results using insights gleaned from mechanistic modelling studies of the HTE (Sec. 1.3.1). As stated above, a robust result of many such studies has been the finding that the vortex is sensitive to QBO phase for only a limited, or “intermediate”, range of planetary wave forcing amplitudes. Since the vortex strength to a large degree determines the planetary wave forcing [Scott and Polvani, 2004]\(^{22}\), we may suppose that the HTE mechanism functions only for a limited range of vortex strengths; when the vortex is too cold or too disturbed, its evolution is insensitive to QBO phase. Given this assumption, we now imagine two thought experiments. In each experiment an ensemble of winter stratospheric states is initialized with a range of different vortex strengths as initial conditions. In one experiment, “A”, the set of initial conditions encompasses the range from extreme cold to extreme warm vortex states, including the intermediate states. In the other experiment, “B”, the extreme cold and extreme warm states are excluded from the initial conditions. In both experiments the QBO is in a state \(\psi\) for which the HTE mechanism will function, provided that the initial vortex strength does not preclude this from happening. Over a timescale for which some memory of the initial conditions is retained, how will ensembles A and B evolve? It is plain that under the given assumption (which is based on a robust conclusion from mechanistic studies) ensemble B will evince a strong HTE signal, while ensemble A will not.

This straightforward idea may be applied to our results as follows. In early winter (Oct and Nov), interannual variability of the NH winter stratosphere is weak. Hence the range of initial conditions is narrow, and we may expect a robust HTE if \(\psi\) is correctly aligned (i.e. \(\psi \in \psi^{HTE}\) in Nov). The Nov HTE will lead to the vortex evolving towards strong or weak states in which it is less sensitive to QBO phase (in a manner reminiscent of the bifurcating behaviour found by O’Sullivan and Dunkerton [1994]). Hence the two opposing groups (W vs. E, or \(\psi^G\) vs. \(\tilde{\psi}^G\)) will each contain

\(^{22}\)Of course one might also say that the planetary waves determine the vortex strength. The essential point is that these two quantities – vortex strength and planetary wave forcing – are mutually dependent, since the wave-vortex coupling is nonlinear.
differently biased ranges of vortex states. Persistence of the vortex state will lead to a decaying HTE signal, weaker than the Nov HTE signal, in Dec (as seen in both ERA-40 and CMAM) and possibly persisting for longer if the vortex persistence is greater, as is suggested by Fig. 4.7(a) for ERA-40 (note the steady reduction in $t$ from Nov-Feb at the $\psi^G_{1} = -0.05$ bin). Hence we suppose that the range of "initial conditions" available in the two groups is unsuitable for independent QBO influence to take effect after Nov, and that the weaker HTE signals for Dec onwards that occur at the same $\psi$ as the Nov HTE signal are due simply to the Nov HTE and to vortex persistence.

Consider also what happens in years when $\psi \not\in \psi^{HTE}$ in Nov, i.e. when no Nov HTE signal occurs. In these cases, the two opposing groups (W vs. E, or $\psi^G$ vs. $\tilde{\psi}^G$) will contain similar ranges of initial conditions, without being biased strongly towards one or the other end of the spectrum of possible vortex states (recalling that interannual variability in Nov is in general small, and is smaller for a collection of years in which a Nov HTE does not occur). If $\psi$ in early winter is in a state that would (given sufficient time) lead to a HTE, then the available range of initial conditions in the two opposing groups now permits this. This may then explain why the Nov HTE is separated (in $\psi$) from the other HTE signals (Feb in both ERA-40 and CMAM, as well as Jan and Mar in CMAM). Development of these signals occurs when $\psi$ is in a state that only weakly excites the HTE mechanism, and hence a HTE signal takes longer to develop. The time required for development of the HTE signal increases (we assume) as $\psi$ moves further from the value $\psi^{HTE}$ that excites the Nov HTE\textsuperscript{23}. An additional prerequisite is that the range of vortex initial states in early winter must be such that "intermediate" values are present. This is most likely to happen in those years when a Nov HTE does not occur.

The above hypothesis has several merits beyond its inherent simplicity. One is that it does not require us to assume that the late winter HTE may only occur for a narrow range of $\psi$ values. Thus it explains why a separation of Nov and Feb

\textsuperscript{23}This is a perfectly reasonable assumption, because if it were not the case then a Nov HTE would occur also at these values of $\psi$. 

peaks of only 0.1 cycles (in CMAM), corresponding to very little contrast in $\Delta \bar{u}_{EQ}$ between different $\psi$-bins, may be associated with the presence or absence of a Feb HTE signal. Rather than assume that a downward propagation of QBO phase by only 2-3 km is very important to the propagation and dissipation of planetary waves – again, recalling the contrast between panels (c) and (d) of Fig. 4.10 – we assume rather that a Feb HTE would occur for both cases (c) and (d) if not for the occurrence of a Nov HTE in case (c). The hypothesis is therefore robust in the sense that it does not depend on any precise assumptions about the exact nature of $\Delta \bar{u}_{EQ}$ that excites a late winter HTE. Another merit is that it appears to explain the fact, noted above, that the Feb and Jan HTE signals are stronger when $\psi$ is defined by its Oct-Dec values rather than by its Jan or Feb values, because the late winter HTE is supposed to be due to a more prolonged development of the vortex from its early winter state (i.e. QBO influence on planetary wave behaviour is weaker, and hence the HTE is “slower”, in the sense that a polar response to the QBO state takes longer to develop). And, of course, it explains why late winter HTE signals are only found at $\psi$ values for which the Nov HTE is weak – without making an explicit assumption about whether those peaks should occur at values of $\psi$ that are earlier or later (in a QBO cycle) than the $\psi$ that gives the Nov HTE. Finally, it is probably consistent with the fact that the persistence of the NH polar vortex in CMAM tends to be low in comparison with ERA-40 (not shown). A more rapid erasure of the “memory” of early winter conditions would be consistent with a strong Jan HTE signal developing in CMAM but not in ERA-40.

### 4.6.2 Internal stratospheric variability at decadal timescales

The previous section, Sec. 4.6.1, discussed hypothetical mechanisms that might explain why different QBO states appear to optimally excite the early and late winter HTE. But regardless of its cause, we may take the $\psi$-separation of the early and late winter HTE signals as a given, and consider its possible implications for the occurrence of internal stratospheric variability on decadal timescales. If decadal clustering of QBO phase transitions – or equivalently, of $\psi$ – occurs, this could potentially lead to
apparent decadal variations in the strength of the HTE. How likely is this possibility, and is there evidence that it has occurred in the real atmosphere?

In Chap. 2, Fig. 2.8 showed the time-varying seasonality of 30 hPa QBO phase transitions for the tropical radiosonde winds\textsuperscript{24}. Fig. 2.8 indicates that 30 hPa transitions are more common in the second half of the calendar year during the second half of the observed record. Fig. 4.3(a) showed that the \( \tau(30) \in [4, 9] \) (Apr-Sep) phase bin gave a strong Feb HTE signal, but that further reduction of \( \tau(30) \) caused the Feb HTE signal to weaken – e.g., for the \( \tau(30) \in [3, 8] \) (May-Oct) or \( \tau(30) \in [2, 7] \) (Jun-Nov) phase bins. This suggests that the late winter HTE is not optimally excited when 30 hPa transitions occur during the second half of the calendar year. Combining these two results suggests that the late winter influence of the QBO on the polar vortex should be weaker in the second half of the observed record than in the first half.

In a recent paper, Lu \textit{et al.} [2008; L08] noted that correlations between the 50 hPa QBO and the polar vortex appeared to be weaker during the 1977-1997 period than during the pre-1977 and post-1997 periods. The pre-1977 period corresponds to that originally examined by Holton and Tan [1980], in which both early and late winter QBO-vortex correlations were found to be strong. In contrast, L08 found the early winter HTE to be slightly weakened, and the late winter (Feb-Mar) HTE substantially weakened, during the 1977-1997 period. Our diagnosis of an apparent connection between HTE strength and 30 hPa QBO phase transition seasonality may provide an explanation for this behaviour, as follows. Fig. 2.8 indicates that during the 1977-1997 period, 8 out of 17 QBO phase transitions at 30 hPa occurred in the May-Aug period, a high number in comparison to the rest of the record (in which only 6 out of 27 transitions occurred during the May-Aug period). The remaining 9 transitions occurred during the Oct-Mar period, which Fig 4.3 shows is associated with a HTE in early, but not late, winter. Fig 4.3 also shows that a decrease in the late winter (Feb-Mar, or Feb) signal occurs when April (9 months prior) is eliminated from

\textsuperscript{24}Fig. 2.8 showed the seasonality of QBO phase transitions at six vertical levels (10, 15, 20, 30, 40, and 50 hPa). We focus attention here on the 30 hPa level because the phase bins of Sec. 4.3 were defined by the 30 hPa QBO phase transitions.
the phase bin, which suggests that the late winter HTE is absent for phase bins that are dominated by the May-Aug period. Hence we might expect the 1977-1997 period to show a strong HTE signal in early winter, but not in late winter; this is consistent with the results of L08. Our analysis therefore suggests that the seasonal clustering of QBO phase transitions on a decadal timescale (Fig. 2.8) may be responsible for there being a weak or nonexistent late winter HTE during the 1977-1997 period.

A complementary view of the behaviour shown in Fig. 2.8 is provided by examining the interannual variability of $\psi$. Fig. 4.17(a) shows the Oct-Mar values of $\psi$ for the tropical radiosonde winds, and Fig. 4.17(b) shows $\psi$ averaged over the whole Oct-Mar period. Both plots indicate that the distribution of $\psi$ is somewhat clustered in the early (roughly, pre-1975) and late (roughly, post-1995) parts of the record, while $\psi$ drifts more uniformly in the middle part of the record. Panels (c)-(e) of Fig. 4.17 show the same data as panel (b), but the years classified as W or E in three different phase bins are highlighted: (c) $\tau(30) \in [7, 12]$ (Jan-Jun), (d) $\tau(30) \in [4, 9]$ (Apr-Sep), (e) $\tau(30) \in [1, 6]$ (Jul-Dec). Referring to Fig. 4.3(a) shows that these phase bins correspond respectively to strong, strong, and absent\textsuperscript{25} HTE signals in late winter (either Feb or the Feb-Mar average). The membership of the W and E groups (i.e., the red and blue dots) in panel (c) of Fig. 4.17 is clearly biased towards the earlier part of the record, while the membership of the W and E groups in panel (e) is clearly biased towards the later part of the record (which includes the 1977-1997 period emphasized by L08). The group members in panel (d) are an intermediate case, in that no clear bias towards the first or second halves of the record is evident. The sequence of panels (c)-(e) suggests that the later part of the record contains fewer years in which QBO phase is optimal for the late winter HTE.

Panels (c)-(e) of Fig. 4.17 show that the highlighted (red and blue) years occupy approximately horizontal bands. This indicates that 30 hPa QBO phase transitions serve as a good proxy for QBO phase defined by the EOF phase angle $\psi$. Although neither phase transitions nor $\psi$ are perfect proxies for $\bar{u}_{EQ}(z)$, the reasonably close cor-

\textsuperscript{25}Or perhaps reversed, since $t \approx -1$ for this case. The low value of $t$ indicates that the reversal has very low statistical significance.
Figure 4.17: Interannual variability of the QBO phase angle $\psi$, defined by tropical radiosonde winds. (a) Values of $\psi$ for Oct-Mar. The values occurring in a given NH winter are displayed at the same location on the abscissa, i.e. they are stacked vertically. The year labeled on the abscissa indicates the year in which a given NH winter begins (i.e. 1960 indicates the 1960/61 winter). (b) As in (a), but the average Oct-Mar value of $\psi$ is shown. (c) As in (b), the Oct-Mar average of $\psi$, but W (E) years from the $\tau(30) \in [7, 12]$ (Jan-Jun) phase bin are highlighted in red (blue). (d) As in (c), but years from the $\tau(30) \in [4, 9]$ (Apr-Sep) phase bin are highlighted. (e) As in (c), but years from the $\tau(30) \in [1, 6]$ (Jul-Dec) phase bin are highlighted.
respondence between them underlies the similarity of the results obtained in Secs. 4.3 and 4.4.2. The correspondence between phase transitions and $\psi$ would be essentially perfect if the rate of QBO phase progression, $\partial \psi / \partial t$, were uniform. That is, a constant value of $\partial \psi / \partial t$ would imply that any horizontal band containing red or blue dots in panels (c)-(e) would contain no green dots. However, the variable spacing of Oct-Mar $\psi$ values in Fig. 4.17(a) shows that $\partial \psi / \partial t$ is not constant, and hence the two definitions of QBO phase do not agree exactly. We may therefore ask: based on the $\psi$-bin results of Sec. 4.4.2, does the decadal drift of $\psi$ values shown in Fig. 4.17 also support the notion – as the Sec. 4.3 results appeared to do – that the late winter HTE signal should weaken in the later part of the record, and in particular during the period 1977-1997 emphasized by L08?

Fig. 4.7 showed that the late winter HTE signal is optimized in the $\psi$-bins for which $\psi_1^G \approx -0.25$. Hence the late winter HTE should be strongest when $\psi$ lies within horizontal bands in Fig. 4.17 that are defined by $\psi \in [-0.25, 0]$ and $\psi \in [0.25, 0.5]$. Inspection of Fig. 4.17 shows, however, that there is not a clear bias towards $\psi$ being located outside of these ranges during the 1977-1997 period, although there clearly is a drift of $\psi$ away from these ranges in the later part of the record. The $\psi$-bin defined by $\psi_1^G \approx -0.25$ thus appears to correspond most closely to the phase bin defined by $\tau(30) \in [4, 9]$ (Apr-Sep), whose members are highlighted in Fig. 4.17(d), and (as already noted) appear evenly distributed throughout the record. The lack of a clear bias suggests that the $\psi$-bin results cannot explain the L08 result for the 1977-1997 period, although they do suggest that the late winter HTE should weaken in the later part of the record (just not during the specific 1977-1997 period).

Why this apparent discrepancy between the two techniques? Since neither phase transitions nor $\psi$ are perfect proxies for $\bar{u}_{EQ}(z)$, there is no particular reason to favour either technique – $\psi$-bins as opposed to phase bins – over the other. Examination of the standard deviation of $\bar{u}_{EQ}$ within the groups (W vs. E, or $\psi^G$ vs. $\tilde{\psi}^G$) defined by either technique indicates no difference in the amount of variance captured by $\psi$

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26 The diagnosed seasonality of the rate of descent of QBO phase, shown in Chap. 2, also demonstrated that the progression of QBO phase proceeds at a variable rate.
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Figure 4.18: (a) QBO phase angle $\psi$ (defined by tropical radiosonde winds) in Nov vs. ERA-40 Feb-Mar average $\bar{u}$ at the equator at 50 hPa. (b) Nov $\psi$ vs. ERA-40 Feb-Mar average $\bar{u}$ at 10 hPa, 55°N. In both panels, years in the range 1977-1997 are highlighted in yellow. Year labeling is defined as in Fig. 4.17.

To resolve this apparent discrepancy, we examine the L08 result for the 1977-1997 period in closer detail. L08 used ERA-40 data to demonstrate a drastically reduced correlation between $\bar{u}$ at 10 hPa, 55°N and equatorial $\bar{u}$ at 50 hPa, where both quantities were averaged over Feb-Mar (see their Fig. 4(a)). Fig. 4.18(a) shows the scatter of Nov $\psi$ vs. Feb-Mar equatorial $\bar{u}$ at 50 hPa, and Fig. 4.18(b) shows the scatter of Nov $\psi$ vs. Feb-Mar $\bar{u}$ at 10 hPa, 55°N. Fig. 4.18(a) demonstrates

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27 Note that while the use of $\psi$ does in principle allow for arbitrarily fine resolution of QBO phase, in practice the size of groups is limited by the length of the record, i.e., by the requirement that groups contain $\approx 10$ (or preferably more) members. This is the reason why neither method should be favoured, at least on the basis of the scatter (i.e., standard deviation) of $\bar{u}_{EQ}(z)$ profiles within the groups defined by each method.

28 $\psi$ is here defined by its Nov value because Nov is the month that defined the $\psi$-bins of Fig. 4.7.
succinctly the relationship between $\psi$ and 50 hPa QBO phase. Fig. 4.18(b) indicates the general sense of the QBO-vortex correlation: for $\psi$ in the middle range (roughly corresponding to QBO-W at 50 hPa) the vortex tends to be stronger, while it tends to be weaker for $\psi$ outside this range (roughly corresponding to QBO-E at 50 hPa).

The years 1977-1997 are highlighted in Fig. 4.18. From Fig. 4.18(b), many of these years are seen to represent intermediate vortex strengths. In particular, the clustering of years 1977, 1982, 1987, 1992 and 1994 into roughly the $\psi \in [-0.15, -0.3]$ range – which is associated with the decadal drift of $\psi$ shown in Fig. 4.17 – may contribute to this tendency. The decadal drift of $\psi$ in the 1977-1997 period leads to the yellow dots in Fig. 4.18 being more spread out over the whole range of $\psi$ than in the earlier (pre-1977) period, as indicated by the more pronounced clustering of the red dots. A succession of years with intermediate vortex strengths will not contribute to a strong HTE signal, but overall it cannot be said – based on the appearance of Fig. 4.18(b) – that the 1977-1997 years strongly buck the HTE trend (with the exception of the outlier 1996). However, it may be noted that two years in which strong negative vortex anomalies occur, 1983 and 1986 (Fig. 4.18(b), top left), occur for values of $\psi$ associated with very weak 50 hPa QBO phase (Fig. 4.18(a), top left). These years will therefore contribute negligibly to the QBO-vortex correlation as determined by L08, effectively degrading their HTE signal. The outlier 1996, for which an extremely strong vortex occurs during QBO-E at 50 hPa, will also degrade the signal. Thus Fig. 4.18 suggests, in summary, that the L08 result for the 1977-1997 period is due partly to the decadal drift of $\psi$, partly to the use of the single-level (50 hPa) QBO phase definition in defining the HTE signal, and partly due to luck (the outlier 1996).

Thus we conclude that there is indeed a reduction in late winter HTE strength in the later part of the observed record that is consistent with the decadal drift of $\psi$, but that this reduction may be less intense than that diagnosed by L08 for the 1977-1997 period. The notion that decadal-scale changes in the strength of the HTE can occur as a result of the decadal drift of $\psi$ (or equivalently, of QBO phase transition seasonality) is consistent with the diagnostics presented in Secs. 4.3 and 4.4. Further testing of this hypothesis requires either a longer observational record or modelling
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Low-frequency sampling variability

The decadal drift of $\psi$, combined with the fact that the HTE signal is optimized when the ERA-40 record is partitioned into subsamples spanning roughly 0.25 QBO cycles, suggests that sampling variability may complicate the reliable detection of a late winter HTE signal when the data are partitioned into temporally contiguous subsamples, as was done by L08. If $\psi$ clusters within the $\psi$-range that we suppose is optimal for exciting the HTE, as Fig. 4.17 suggests happens in the first half of the record, the HTE signal is likely to be diagnosed as strong. Clustering of $\psi$ outside the optimal range, as Fig. 4.17 suggests may occur towards the end of the record, is likely to yield a weak HTE signal. Periods of fairly uniform drift of $\psi$, as appears to occur during 1975-1990 period, will be an intermediate case. Given a temporally contiguous subsample of length $T$ that is used to attempt to detect the HTE correlation, a robust estimate of the HTE signal will require larger $T$ during periods when $\psi$ is predominantly found outside the optimal HTE $\psi$-ranges. Of course, this truth of this assertion is contingent upon our hypothesis regarding the sensitivity of the HTE to $\psi$ being correct. However, that is the hypothesis that we consider here.

Another possible subsampling consists of stratifying the data according to the phase of the 11-year solar cycle, in which case each solar phase yields several temporally contiguous groups, each with $T \approx 5$ years, and separated from the next group (belonging the same solar phase) by $\approx 5$ years. It has been shown that a strong late winter HTE appears to occur during solar minima, but that during solar maxima the HTE signal appears to reverse [Labitzke and van Loon, 1988] or perhaps simply disappear [Gray et al., 2001b]. However, since the observed record contains only a small number of solar cycles (the 44-year ERA-40 period contains four) it has been suggested that the apparent modulation of the HTE by the solar cycle may be a manifestation of sampling variability, rather than a causal effect of the solar cycle [Baldwin and Dunkerton, 1989].

Fig. 4.19(a) shows the relationship between $\psi$ and the Jan-Feb average vortex
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Figure 4.19: (a) QBO phase angle $\psi$ (defined by tropical radiosonde winds) in Nov vs. ERA-40 Jan-Feb average $\bar{u}$ at 50 hPa, 60°N. (b) As in (a), but for Feb instead of Jan-Feb. Year labeling is defined as in Fig. 4.17. Colouring denotes the phase of the 11-year solar cycle: maximum (red), minimum (green), or intermediate and hence unclassified (yellow); see text for details.

strength, using the 50 hPa, 60°N gridpoint to represent the vortex. The Jan-Feb average is chosen so that a direct comparison may be made with Fig. 2(d) of Naito and Hirota [1997]. The colouring of the dots indicates the phase of the 11-year solar cycle, with the classification performed in the same way as Naito and Hirota [1997]: if the Jan-Feb mean value of the 10.7 cm radio flux is less than $120 \times 10^{-22} \text{ W m}^{-2} \text{ Hz}^{-1}$ (greater than $140 \times 10^{-22} \text{ W m}^{-2} \text{ Hz}^{-1}$) then the winter is assigned to the minimum (maximum) group. Fig. 4.19(b) indicates the same thing as (a), but for the Feb average instead of the Jan-Feb average, which allows direct comparison with Fig. 4.7. (Recall that Feb is found to produce the strongest late winter HTE signal, while the Jan signal is of similar sign but lower significance than the Feb signal.) Fig. 4.7 diagnoses the Feb HTE signal using all data points in Fig. 4.19(b) (i.e., irrespective of colouring), in a series of subsamples based on the value of $\psi$ with each subsample.

Note that the convention used in Fig. 2(d) of Naito and Hirota [1997] is to label the year (on the horizontal axis) according to the year of the Jan/Feb in question, i.e. the winter of 1960/61 is denoted as 1961. The labeling of years in Fig. 4.19 employs the same convention as Fig. 4.17, using the first year of NH winter to label the year, i.e. the winter of 1960/61 is denoted as 1960.
containing approximately half of the available years. When the data are stratified according to the phase of the solar cycle, the HTE signal is diagnosed using only the green (solar min) or red (solar max) points. Since all but six years are classified as either solar min or max, each these groups also involves approximately half of the available years.

It is clear from Fig. 4.19 that the solar minimum years (green) appear to show a strong HTE signal, while the solar maximum years (red) appear to show little HTE signal; these facts are consistent with Fig 2(d) of Naito and Hirota [1997]. (This point was also made by Gray et al. [2001b].) Fig. 4.19 also shows that neither solar cycle phase has an obvious preference for being found in any particular subrange of $\psi$; hence it is unlikely that the strong HTE signal for solar min is due to the drift of $\psi$ into alignments that are favourable for the LW HTE (and vice versa for the negligible signal during solar max). But inspection of Fig. 4.19 indicates that for a given solar cycle phase, only $\approx 5$ years tend to fall into any quarter-cycle range of $\psi$ that gives a strong LW HTE signal (cf. Fig. 4.7), and moreover we may conclude from the results of Secs. 4.3 and 4.4 that stratification of the data by solar cycle phase is not required for detection of the LW HTE signal. These facts suggest that the apparent solar modulation of the LW HTE signal may be a manifestation of sampling variability, since a small number of years (i.e., $\approx 5$) fall within the range of $\psi$ values that gives a strong LW HTE signal for each solar cycle phase.

Again, we must reiterate that the observed record is of insufficient length to justify firm conclusions regarding the role of either $\psi$ or the solar cycle, and hence either a lengthened observational record or modelling studies are required to shed further light on this question. However, confidence in the hypothesis that decadal drifts in $\psi$ may have an impact on the late winter HTE strength is bolstered by the CMAM results presented in this chapter. As noted in Sec. 1.5.1, no external sources of interannual variability – including the 11-year solar cycle – are imposed in CMAM. However, Secs. 4.3 and 4.4 showed that the $\psi$-dependence of the CMAM HTE signal bore some resemblance to the $\psi$-dependence of the ERA-40 HTE signal. There is also a similarity between ERA-40 and CMAM in the decadal drift of $\psi$. Fig. 4.20 shows
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Figure 4.20: Interannual variability of the QBO phase angle $\psi$ for the CMAM QBO. As in Fig. 4.17, the values occurring in a given NH winter are displayed at the same location on the abscissa, i.e. they are stacked vertically. Oct-Mar values of $\psi$ in each year of the 150-year run are shown.

that $\psi$ tends to alternate, on decadal timescales, between drifting and clustering in a manner similar to that seen in Fig. 4.17. This behaviour is of course consistent with the decadal variations in QBO phase transition seasonality shown in Fig. 2.9. By the results of Secs. 4.3 and 4.4, the behaviour indicated by Figs. 2.9 and 4.20 will lead to decadal variations in the HTE strength.
Chapter 5

Conclusions

5.1 Summary of main results

Chapter 2 characterized the QBO in CMAM by comparing it with the observed QBO. The CMAM QBO was shown in Sec. 2.2 to be realistic in some respects (amplitude, mid-stratospheric width, seasonal clustering of phase transitions) and unrealistic in others (too long period, westerly bias in lower stratosphere, too narrow lower stratospheric width, too shallow vertical extent). Sec. 2.3 argued that the existence of a variable QBO period in CMAM suggested that similar behaviour in the real atmosphere could be a result of internal atmospheric variability, rather than external forcings (including SSTs). Seasonal modulation of the QBO phase descent rate was argued to be the main factor determining the seasonal clustering of QBO phase initiations, suggesting that this clustering is not an artefact of a short (50 years, or \( \approx 20 \) QBO cycles) observational record. Future work involving a detailed analysis of the CMAM QBO zonal momentum budget may reveal the mechanism(s) behind the seasonal clustering of phase initiations.

Chapter 3 compared the stratospheric climates of the CMAM QBO and control runs. Sec. 3.2 showed that changes in tropical stratospheric \( \bar{u} \) led to statistically significant changes in extratropical stratospheric \( \bar{u} \) and \( \bar{T} \) during NH winter. The timing and pattern of the extratropical changes were found to be consistent with the Holton-Tan hypothesis that the coupling between tropical and polar latitudes is due
to the effect of the low-latitude $\bar{u} = 0$ line on the dissipation of stationary planetary waves during winter. However, although the magnitude of the QBO-control cooling in the extratropical stratosphere was similar to the W-E cooling associated with the HTE in ERA-40, the CMAM cooling did not extend to the pole. One interpretation of this result is that the simple introduction of time-mean westerlies in the lower stratosphere in CMAM does not substantially affect the seasonal average NH polar vortex strength.

While the timing of QBO-control seasonal differences suggested consistency with the Holton-Tan hypothesis, it was useful to rule out the possibility that extratropical stratospheric $\Delta \bar{u}$ and $\Delta \bar{T}$ signals were strongly forced by parameterized gravity waves, since changes in tropical $\bar{u}$ may equally well affect both resolved and parameterized waves. Sec. 3.3 analyzed the extratropical momentum budget and determined that the extratropical stratospheric changes were predominantly forced by resolved waves, although a feedback due to parameterized orographic gravity waves was found to be important in the subtropical upper stratosphere.

Given the dominant role of resolved waves, Sec. 3.4 sought to gain insight into the nature of the $\Delta(\nabla \cdot \mathbf{F})$ response during NH winter. The main question of interest was the means by which low-latitude changes in $\bar{u}$ interact with $\bar{u}$ at high latitudes. For the ERA-40 W-E case, the response of $\Delta(\nabla \cdot \mathbf{F})$ extended to higher latitudes than in the QBO-control case. Sec. 3.4 showed that changes in $\Delta F^z(z)$ at mid and high latitudes could either reinforce (ERA-40) or counteract (CMAM) the changes in horizontal planetary wave propagation induced by $\Delta \bar{u}_{EQ}$. It was suggested that the $\Delta F^z(z)$ response is due to a wave-vortex feedback that is characteristic of natural polar vortex variability, and that the different ERA-40 and CMAM responses were (consequently) due to different climatological vortex strengths and/or geometries. Given that no statistically significant QBO-control tropospheric $\Delta(\nabla \cdot \mathbf{F})$ response was found (indicating no change in tropospheric wave sources), this result suggests that the stratosphere is able to control the vertical flux of wave activity into it, in response to a stratospheric perturbation ($\Delta \bar{u}_{EQ}$). This behaviour is consistent with mechanistic studies [Gray et al., 2003; Scott and Polvani, 2004].
5. Conclusions

Chapter 4 analyzed the NH winter Holton-Tan effect (HTE) in the CMAM QBO run and in ERA-40. Secs. 4.3 and 4.4 showed that in both datasets, the early and late winter HTE signals were found to occur for QBO phase orientations that were separated by roughly a quarter cycle in QBO phase (where the phase in both cases was defined by its November value). The polar responses in the two datasets occurred for $\Delta \bar{u}_{EQ}$ perturbations that were very similar in the 25-40 km layer, suggesting that these altitudes may be causal for the HTE. The apparent unimportance of the $z < 25$ km region is consistent with the fact that the QBO-control extratropical $\Delta \bar{u}$ signal (Chap. 3) did not extend to the pole even though a strong $\Delta \bar{u} > 0$ feature occurred in the tropical lowermost stratosphere (due to the westerly bias of the CMAM QBO, Sec. 2.2). Sec. 4.5 showed that although the overall amplitude of the CMAM HTE signal was roughly half that of the ERA-40 signal, its spatial structure was somewhat realistic. Additionally, it is notable that an early winter (Nov) HTE signal was found in CMAM, as most previous GCM studies have obtained a HTE only in mid or late NH winter (Sec. 1.3.2).

It was noted in Sec. 4.2 that if equatorial $\bar{u}$ at 50 hPa is used to define QBO phase in CMAM, the resulting HTE signal has the wrong sign and timing, whereas the two methods employed in Secs. 4.3 and 4.4 revealed that somewhat realistic HTE signals occurred in CMAM despite the CMAM QBO having some unrealistic aspects (Sec. 2.2). This shows that 50 hPa $\bar{u}$ is only a proxy for the vertical structure $\bar{u}(z)$, and the relationship of the proxy to the region of causal influence depends on the vertical wavelength of the QBO, which is different in ERA-40 and CMAM. Thus when diagnosing the HTE in GCMs, a more robust indicator of QBO phase (such as those considered here) needs to be employed.

Sec. 4.6.1 discussed the causality of the early and late NH winter HTE signals. It was argued that the Nov HTE signal is most likely to result from $\Delta \bar{u}_{EQ}$ in the mid-stratosphere (consistent with arguments made in Sec. 3.4). One reason for this is that the seasonal variation of low-latitude $\bar{u}$ is such that strong QBO modulation of the $\bar{u} = 0$ line’s location in the NH for $z < 30$ km does not occur until Dec-onwards. It is therefore likely that CMAM’s realistic representation of the seasonal cycle plays a key
role in the realism of the CMAM HTE; the seasonal cycle has often been neglected in mechanistic modeling studies of the HTE (Sec. 1.3.1). The causality of the late winter HTE signal is less clear, but it was hypothesized that the early winter HTE may bias the vortex state away from mid-range states that make a late winter HTE more likely. This would be consistent with the notion that the polar vortex is sensitive to $\Delta \bar{u}_{EQ}$ for only a middle range of vortex strengths (i.e., of wave forcings), which is a finding shared by virtually all mechanistic modelling studies of the HTE (Sec. 1.3.1). This hypothesis could explain the finding of Secs. 4.3 and 4.4 that the late winter HTE signal appears to be strongest when the early winter signal is weakest.

Finally, Sec. 4.6.2 discussed the possibility that the seasonal alignment of QBO phase transitions (Sec. 2.3) may lead to a modulation of HTE strength on decadal timescales, due to its effect on the structure of $\Delta \bar{u}_{EQ}$ that occurs during NH winter. Owing to the very high statistical significance of the early winter (Nov) HTE signal seen in ERA-40, it was suggested that this effect would be primarily of importance for the late winter HTE signal. It was argued that the results of Secs. 4.3 and 4.4, combined with the seasonal clustering of phase transitions shown in Sec. 2.3, suggested that the HTE strength should appear to be weakened in the second half of the ERA-40 record when the QBO phase was defined by the sign of the 50 hPa equatorial wind. This is consistent with the results of Lu et al. [2008]. It was also suggested that this effect might masquerade as an apparent solar cycle modulation of the HTE. Although no firm conclusion in this regard is possible until a longer observational record is obtained, the fact that similar behaviour occurs in CMAM – where it can only be due to internal atmospheric variability – is suggestive of this possibility.

## 5.2 Future work

Sec. 5.1 has highlighted a number of realistic aspects of the QBO-vortex interaction (i.e., HTE signal) that occur in CMAM despite the CMAM QBO having some unrealistic features. Naturally it would be interesting to know if these conclusions remain valid when the model exhibits a more realistic QBO. Obtaining a QBO in a
GCM that is both realistic and spontaneous is a non-trivial task (for reasons noted in Sec. 2.1), but it represents an essential extension of the work presented here. Indeed, perhaps the most salient point to be taken from our results is the remarkable fact that the extratropical response to the QBO in CMAM is as realistic as it is (as argued in Chap. 4), given the obvious deficiencies in the CMAM QBO itself. This is a promising result, as it suggests that a better QBO representation is worth pursuing not only for the sake of the QBO itself, but also for the extratropical climate.

An improved representation of the QBO in CMAM will allow some QBO-specific issues raised in Chap. 2 to be addressed. The partial seasonal synchronization of QBO phase transitions appears to be important for QBO-vortex coupling (as argued in Chap. 4), yet its mechanism is not understood. Simpler models may be useful for testing specific hypotheses about the origin of this behaviour, but analysis of GCM results will ultimately establish whether it is an important aspect of a realistic QBO. Another important question concerns how the QBO might respond to external forcings such as changing greenhouse gas concentrations. An accurate answer will require GCM studies in which the QBO is realistic and is forced by the correct distribution of tropical wave types.

It was argued in Chap. 3 that anomalies in Rossby-wave breaking (RWB) location and intensity are responsible for the poleward migration of the HTE signal. Empirical diagnostics of RWB support this view [Baldwin and Holton, 1988], but the exact connection between such diagnostics and $\nabla \cdot \mathbf{F}$ is unclear, and it would be interesting to see if it can be made more explicit. Hitchman and Huesmann [2007] presented a climatology of RWB events in the NCEP/NCAR reanalysis that provides a first step in this direction, and a possible extension of their work would be to derive similar climatologies for ERA-40 and CMAM. The exact relation of such empirical diagnostics to QBO modulation of the low-latitude $\bar{u} = 0$ position remains unclear. Attempting to quantify this relationship may provide an explicit test of the arguments made in Sec. 3.4.2.

The results of Chap. 4 suggest that further mechanistic modelling studies of the HTE, incorporating realistic forms of $\Delta \bar{u}_{EQ}$ suggested by our HTE diagnosis, would be
useful. Such studies would allow explicit testing of our assertion that distinct forms of \( \Delta \bar{u}_{EQ} \) are associated with the early and late winter HTE signals. Moreover, the role of the seasonal cycle could be tested in a controlled manner, so as to establish whether it is the radiative seasonal cycle per se, rather than the occurrence of radiative relaxation at all, that is responsible for the timing of the observed NH HTE signal. As noted in Chap. 1, most mechanistic studies to date have considered idealized forms of \( \Delta \bar{u}_{EQ} \) and neglected the seasonal cycle. These simplifying assumptions were well-motivated by the need to obtain a first-order understanding of QBO-vortex interactions, but the results of our analysis suggest that the addition of some extra complexity may now be useful and timely.

Finally, the results of Chap. 4 may be extended by considering alternate definitions of the polar vortex strength. Several are available: SSWs, annular modes, and the area enclosed by specified PV contours [Butchart and Remsberg, 1986]. Consideration of daily data (rather than the monthly averages employed herein) may shed further light on the timing of QBO influence. An extension of our analysis to the Southern Hemisphere winter is also desirable.
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