The Structure and Variability of the Tropical General Circulation

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Abstract

The Structure and Variability of the Tropical General Circulation

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This study explores the structure and variability of the tropical atmospheric circulation, with emphasis on the relationships between temperature, wind, mean meridional circulation (MMC), and the eddy fluxes of zonal momentum.

The patterns of MMC streamfunction, radiative heating rates, cloud fraction and mixing ratios of various trace species illustrate the clear distinction between the tropospheric Hadley circulation and the stratospheric Brewer-Dobson circulation (BDC), with regard to the meridional width of the tropical upwelling and the equatorial symmetry of the seasonal cycle structure.

Tropical upwelling in the BDC, inferred from the equatorially-even component of the zonally-averaged, lower-stratospheric temperature profiles, appears to be modulated by variations in eddy heat fluxes at high latitudes on both the intraseasonal and interannual time scales. The response of an idealized axisymmetric model to an externally imposed wave drag over a fixed region in the high latitudes does not resemble sudden warmings accompanied by a collapse of the polar night jet, and it does not extend into the tropics. Reconciling between our observational and model results will likely require the use of a more realistic forcing distribution in three-dimensional models capable of resolving the finer scale features emanating from high to low latitudes.
The nonseasonal variability of zonally-averaged zonal wind \( u \) is dominated by the equatorially-even component and strongly coupled to fluctuations in the equatorially-even MMC cells. Intraseasonal variations in \( u \) exhibit prominent maxima in the upper troposphere over the equator and ~25°N/S. The Madden-Julian Oscillation contributes to the variability over the equator, but explains only a modest fraction of the variance. Pulsations in convection over the Indo-Pacific warm pool region give rise to equatorial Kelvin and Rossby waves that conspire to produce a zonally-symmetric response. Subtropical \( u \) variability appears to be forced, in part, by variations in the extratropical eddy momentum fluxes. Instability of the jet at the jet exit regions over the Pacific may also contribute to the variability at subtropical latitudes and produce a Rossby wave couplet that straddles the equator, thereby inducing \( u \) anomalies over the equator.
# TABLE OF CONTENTS

<table>
<thead>
<tr>
<th>Chapter</th>
<th>Title</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>List of Figures</td>
<td></td>
<td>iii</td>
</tr>
<tr>
<td>List of Tables</td>
<td></td>
<td>viii</td>
</tr>
<tr>
<td>Chapter 1: Introduction</td>
<td></td>
<td>1</td>
</tr>
<tr>
<td>1.1</td>
<td>The Brewer-Dobson circulation</td>
<td>3</td>
</tr>
<tr>
<td>1.1.1</td>
<td>The downward control principle</td>
<td>5</td>
</tr>
<tr>
<td>1.2</td>
<td>Tropical troposphere</td>
<td>7</td>
</tr>
<tr>
<td>1.3</td>
<td>Outline of the dissertation</td>
<td>10</td>
</tr>
<tr>
<td>Figures and Tables</td>
<td></td>
<td>12</td>
</tr>
<tr>
<td>Chapter 2: Structure of the Tropical Mean Meridional Circulation</td>
<td></td>
<td>13</td>
</tr>
<tr>
<td>2.1</td>
<td>Introduction</td>
<td>13</td>
</tr>
<tr>
<td>2.2</td>
<td>Data</td>
<td>14</td>
</tr>
<tr>
<td>2.3</td>
<td>Annual mean</td>
<td>16</td>
</tr>
<tr>
<td>2.4</td>
<td>Seasonal cycle</td>
<td>18</td>
</tr>
<tr>
<td>2.5</td>
<td>Summary</td>
<td>20</td>
</tr>
<tr>
<td>Figures and Tables</td>
<td></td>
<td>22</td>
</tr>
<tr>
<td>Chapter 3: Tropical Stratospheric Circulation</td>
<td></td>
<td>25</td>
</tr>
<tr>
<td>3.1</td>
<td>Introduction</td>
<td>25</td>
</tr>
<tr>
<td>3.2</td>
<td>Data and methods</td>
<td>27</td>
</tr>
<tr>
<td>3.3</td>
<td>Structure and variability of the Brewer-Dobson circulation</td>
<td>29</td>
</tr>
<tr>
<td>3.3.1</td>
<td>Annual mean</td>
<td>29</td>
</tr>
<tr>
<td>3.3.2</td>
<td>The variability about the mean</td>
<td>30</td>
</tr>
<tr>
<td>3.3.3</td>
<td>Nonseasonal variability</td>
<td>33</td>
</tr>
<tr>
<td>3.4</td>
<td>Forcing of the tropical upwelling in the Brewer-Dobson circulation</td>
<td>35</td>
</tr>
<tr>
<td>3.4.1</td>
<td>The role of high-latitude wave forcing</td>
<td>35</td>
</tr>
<tr>
<td>3.4.2</td>
<td>The role of forcing from below</td>
<td>37</td>
</tr>
<tr>
<td>3.4.3</td>
<td>The role of wave forcing in the subtropical lower stratosphere</td>
<td>38</td>
</tr>
<tr>
<td>3.5</td>
<td>Model results</td>
<td>40</td>
</tr>
<tr>
<td>3.5.1</td>
<td>Introduction</td>
<td>40</td>
</tr>
<tr>
<td>3.5.2</td>
<td>Steady-state vs. periodic forcing</td>
<td>43</td>
</tr>
<tr>
<td>3.5.3</td>
<td>Response to time-varying forcing</td>
<td>44</td>
</tr>
<tr>
<td>3.5.3.1</td>
<td>Sensitivity to the latitude of the forcing</td>
<td>45</td>
</tr>
<tr>
<td>3.5.3.2</td>
<td>Sensitivity to the level of the forcing</td>
<td>46</td>
</tr>
<tr>
<td>3.5.3.3</td>
<td>Sensitivity to the amplitude of the forcing</td>
<td>46</td>
</tr>
<tr>
<td>3.5.3.4</td>
<td>Sensitivity to the presence of the stratospheric polar vortex</td>
<td>46</td>
</tr>
<tr>
<td>3.5.3.5</td>
<td>Sensitivity to the time scale of the radiative damping</td>
<td>47</td>
</tr>
<tr>
<td>3.5.3.6</td>
<td>Sensitivity to the stratospheric static stability</td>
<td>47</td>
</tr>
</tbody>
</table>
### 3.6 Summary and concluding remarks ................................................................. 48
Figures and Tables .............................................................................................. 51

#### Chapter 4: Tropical Tropospheric Circulation .............................................. 74
4.1 Introduction .................................................................................................... 74
4.2 Data and methods ....................................................................................... 76
4.3 General characteristics ............................................................................. 78
  4.3.1 Variability of the mean meridional circulations .................................... 78
  4.3.2 Variability of the meridional wind component .................................... 81
  4.3.3 Variability of the zonal wind component ......................................... 82
  4.3.4 Summary and discussion ...................................................................... 83
4.4 Intraseasonal variability of the zonal wind \([u]\) component ....................... 85
  4.4.1 Dominant space-time pattern of tropical \([u]\) variability .................... 86
  4.4.2 Variability of \([u]\) over the equator ...................................................... 90
  4.4.3 Variability of \([u]\) at 25°N ..................................................................... 94
  4.4.4 The role of the eddy momentum fluxes .............................................. 96
  4.4.5 Summary and discussion ...................................................................... 97
4.5 Model results ............................................................................................... 102
  4.5.1 Climatological mean structures .......................................................... 103
  4.5.2 Variability of the mean meridional circulations .................................. 104
  4.5.3 Variability of the meridional wind component .................................... 106
  4.5.4 Variability of the zonal wind component .......................................... 106
4.6 Summary and concluding remarks ............................................................. 107
Figures and Tables .............................................................................................. 110

#### Chapter 5: Summary and Future Work ......................................................... 156
Figures and Tables .............................................................................................. 160

#### References .................................................................................................... 161

#### Appendix A: Model Descriptions ................................................................. 173
  A.1 Dry dynamical core model ..................................................................... 173
    A.1.1 Adjustments to the model stratosphere ......................................... 175
    A.1.2 Forcing specifications ...................................................................... 177
  A.2 Moist dynamical core model .................................................................. 179
  A.3 Atmospheric Model 2 .............................................................................. 180
Figures and Tables .............................................................................................. 182
**LIST OF FIGURES**

<table>
<thead>
<tr>
<th>Figure Number</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>1.1 Climates of annual mean and r.m.s. amplitudes of (u), (v), (\Psi) and (u^<em>v^</em>) based on daily ERA-Interim data during 2000-08</td>
<td>12</td>
</tr>
<tr>
<td>2.1 Vertical cross-sections of the climatological annual mean and Jan-Feb minus Jul-Aug fields of (T), (\Psi), (v) and the sum of the shortwave and longwave radiative heating rates based on daily ERA-Interim and monthly JRA-25 data</td>
<td>22</td>
</tr>
<tr>
<td>2.2 Vertical cross-sections of the climatological annual mean fields of zonally-averaged cloud fractions and mixing ratios of WV, O(_3) and CO</td>
<td>23</td>
</tr>
<tr>
<td>2.3 As in Fig. 2.2 but for Jan-Feb minus Jul-Aug fields</td>
<td>24</td>
</tr>
<tr>
<td>3.1 Monthly mean and anomaly time series of global-mean (T)</td>
<td>51</td>
</tr>
<tr>
<td>3.2 Meridional profile of climatological annual mean (T)</td>
<td>51</td>
</tr>
<tr>
<td>3.3 Meridional profiles of total, odd and even components of monthly mean (T^*) for Jan/Feb and Jul/Aug</td>
<td>52</td>
</tr>
<tr>
<td>3.4 EOF/PC 1 and 2 of monthly mean, odd component of (T)</td>
<td>53</td>
</tr>
<tr>
<td>3.5 EOF/PC 1 and 2 of monthly mean, even component of (T)</td>
<td>54</td>
</tr>
<tr>
<td>3.6 Monthly mean time series of the even component of polar cap (T), zonal wind in the midlatitude stratosphere, tropical (T^*), and PC 1 of monthly mean, even component of (T)</td>
<td>54</td>
</tr>
<tr>
<td>3.7 Monthly mean time series (plotted as a function of calendar month) of polar cap (T), zonal wind in the midlatitude stratosphere, tropical (T^*), PC 1 of monthly mean, even component of (T), EP flux index, and EPW index</td>
<td>55</td>
</tr>
<tr>
<td>3.8 Meridional profiles of total and even component of monthly anomaly (T^*) for Jan/Feb and Jul/Aug</td>
<td>56</td>
</tr>
<tr>
<td>3.9 EOF/PC 1 and 2 of monthly anomaly, even component of (T)</td>
<td>57</td>
</tr>
<tr>
<td>3.10 Monthly anomaly time series of the even component of polar cap (T), zonal wind in the midlatitude stratosphere, tropical (T^*), and PC 1 of monthly mean, even component of (T)</td>
<td>57</td>
</tr>
<tr>
<td>3.11 EOF/PC 1 and 2 of monthly anomaly, even component of (T^*) based on high-pass and low-pass filtered data</td>
<td>58</td>
</tr>
<tr>
<td>3.12 Monthly mean and anomaly time series of PC 1 of even component of (T) and EP flux index</td>
<td>59</td>
</tr>
<tr>
<td>3.13 Monthly anomaly time series of PC 1 of even component of (T^*) and EP flux index based on high-pass and low-pass filtered data</td>
<td>59</td>
</tr>
<tr>
<td>3.14 Meridional profiles of regression and correlation coefficients between monthly anomaly, even component of (T^*) and monthly anomaly time series of EP flux index based on high-pass and low-pass filtered data</td>
<td>60</td>
</tr>
<tr>
<td>3.15 Sea surface temperature field regressed upon PC 1 of monthly anomaly, even component of (T^*) based on low-pass filtered data</td>
<td>61</td>
</tr>
</tbody>
</table>
3.16 Vertical cross-sections of $[v*T^*], \mu*v^*, \text{and EP flux divergence fields in the previous month regressed upon the monthly anomaly time series of tropical } T^*$. 62
3.17 Vertical cross-section of $[\mu*v^*]$ field in the current month regressed upon the monthly anomaly, Dec-Apr time series of tropical $T^*$ 63
3.18 Vertical cross-sections of the terms in the zonal momentum balance equation for a 3D simulation with steady easterly torque 64
3.19 As in Fig. 3.18 but using $T_{eq}$ profile with a polar vortex in the Northern Hemisphere 65
3.20 As in Fig. 3.18 but based on an axisymmetric model 66
3.21 Vertical cross-sections of the time-mean $\Psi$ field using $T_{eq}$ profile as in Held and Suarez (1994) and $T_{eq}$ profile with constant tropical temperatures 67
3.22 Vertical cross-sections of the transient $\Psi$ field at the four phases of the periodic forcing cycle 68
3.23 Transient meridional profiles of lower-stratospheric $T$ at the time of the peak forcing 69
3.24 Vertical cross-sections of the time-mean $\Psi$ field in Fig. 3.22 69
3.25 Time-mean meridional profiles of lower-stratospheric $T$ for the various simulations 70
3.26 As in Fig. 3.25 but for pressure velocity 71
4.1 Daily time series of 150-hPa $[u]_{ONS}$ and $[v]_{ONS}$ 110
4.2 Vertical cross-section of the correlation coefficients between the daily nonseasonal time series of $[u]$ at the same latitude of the Northern and Southern Hemispheres at each pressure level 110
4.3 Vertical cross-sections of $[\Psi], [\mu*v^*], [u]$ and 2-day $d[u]/dt$ regressed upon PC 1 of nonseasonal $[\Psi]$ 111
4.4 Vertical cross-sections of $[\Psi], [\mu*v^*], [u]$ and 10-day $d[u]/dt$ regressed upon PC 2 of nonseasonal $[\Psi]$ 111
4.5 EOFs 1 and 2 of nonseasonal $[v]$ 112
4.6 Vertical cross-sections of $[v]$ and $[\Psi]$ regressed upon IMFs of PC 1 of nonseasonal $[v]$ 113
4.7 EOF 1 of 30-day high-pass and low-pass Lanczos filtered nonseasonal $[\Psi]$ 113
4.8 Daily time series of the equatorially even and odd components of 150-hPa $[u]_{12NS}$ 114
4.9 MCA 1 and 2 of nonseasonal $[u]$ and 10-day $d[u]/dt$ 115
4.10 Meridional profiles of the 150-hPa $[u]$ of the 50-days with the highest and lowest equatorial $[u]$ during Dec-Mar (2000-08) 115
4.11 Vertical cross-sections of $[u]$ regressed upon PC 1 and 2 of the nonseasonal, zonally-averaged angular momentum field (45°N-45°S, 100-1000 hPa) 116
4.12 Vertical cross-sections of $[u]$ regressed upon PC 1-3 of the nonseasonal, zonally-averaged angular momentum field (90°N-90°S, 100-1000 hPa) 116
4.13 Lag-correlation functions between (a) PC 1 and 2 of the nonseasonal angular momentum field (45°N-45°S, 100-1000 hPa) and (b) time series of nonseasonal 150-hPa $[u]_{25N, 25S}$ and $[u]_{0NS}$ .......................................................... 117

4.14 Vertical cross-sections of $[u]$ regressed upon the linear combinations of PC 1 and 2 of the nonseasonal angular momentum field (45°N-45°S, 100-1000 hPa) .......................................................... 118

4.15 Vertical cross-sections of $[\Psi]$, $[u*v*]$, $[u]$ and 10-day $d[u]/dt$ regressed upon the daily time series of 150-hPa $[u]_{0NS}$ ........................................................................................ 119

4.16 As in Fig. 4.15 but regressed upon the daily time series of 150-hPa $d[u]/dt_{0NS}$ .................................................................................................. 119

4.17 As in Fig. 4.15 but regressed upon the daily time series of 150-hPa $[u]_{25N}$ during Nov-Mar, from which Nov-Mar means for each year have been subtracted out ........................................................................ 120

4.18 As in Fig. 4.15 but regressed upon the daily time series of 150-hPa $d[u]/dt_{25N}$ during Nov-Mar, from which Nov-Mar means for each year have been subtracted out ........................................................................ 120

4.19 As in Fig. 4.15 but regressed upon the daily time series of 150-hPa $[u]_{25S}$ during May-Sep ........................................................................ 121

4.20 As in Fig. 4.15 but regressed upon the daily time series of 150-hPa $[u]_{25S}$ during Nov-Mar, from which Nov-Mar means for each year have been subtracted out ........................................................................ 121

4.21 Vertical cross-sections of $[\Psi]$, $[u*v*]$, $[u]$ and 10-day $d[u]/dt$ regressed upon EC 1 of $[\Psi]$, derived from MCA of nonseasonal $[\Psi]$ and $[u*v*]$ .......................................................... 122

4.22 As in Fig. 4.21 but for the second MCA mode .......................................................... 122

4.23 Vertical cross-sections of $[\Psi]$, $[u*v*]$, $[u]$ and 10-day $d[u]/dt$ regressed upon EC 1 of $[\Psi]$, derived from MCA of nonseasonal, 10-day low-pass Lanczos filtered $[\psi]$ and unfiltered $[u*v*]$ fields .................................................................. 123

4.24 Vertical cross-sections of $[\Psi]$, $[u*v*]$, $[u]$ and 10-day $d[u]/dt$ regressed upon the linear combination of MJO 1 and 2, in phase with maximum 150-hPa $[u]_{0NS}$ .......................................................... 124

4.25 As in Fig. 4.24 but regressed upon the linear combination of MJO 1 and 2, in quadrature with maximum 150-hPa $[u]_{0NS}$ .......................................................... 124

4.26 Regression coefficients between various time series plotted on the axes of MJO 1 and 2, $[u]_{0NS}$ and 10-day $d[u]/dt_{0NS}$, and $P$ and 10-day $dP/dt$ .......................................................... 125

4.27 Time-latitude sections of 150-hPa $[u]$, $[\Psi]$ and $[u*v*]$ regressed upon the linear combinations of (a) MJO 1 and 2, and (b) 150-hPa $[u]_{0NS}$ and 10-day $d[u]/dt_{0NS}$ .......................................................... 126

4.28 Composite maps of with respect to the highest and lowest 200 days of MJO 1 and 2 ........................................................................................................ 127

4.29 Precipitation, 150-hPa geopotential height and wind fields regressed upon the linear combinations of MJO 1 and 2 ........................................................................................................ 128
4.30 Time-longitude sections of 150-hPa $u_{0NS}$ and $\Phi_{26NS}$ regressed upon the linear combinations of MJO 1 and 2, 150-hPa $[u]_{0NS}$ and 10-day $d[u]/dt_{0NS}$, and MJO-residual, 150-hPa $[u]_{0NS}$ and 10-day $d[u]/dt_{0NS}$ ........................................ 129

4.31 Precipitation, 150-hPa geopotential height and wind fields regressed upon the linear combinations of 150-hPa $[u]_{0NS}$ and 10-day $d[u]/dt_{0NS}$ ................................................... 130

4.32 MCA 1 and 2 of MJO-residual, nonseasonal $[u]$ and 10-day $d[u]/dt$ ........................................ 131

4.33 Vertical cross-sections of $[u]$ regressed upon PC 1-4 of the nonseasonal, zonally-averaged angular momentum field (45°N-45°S, 100-1000 hPa), based on MJO-residual data .................................................. 132

4.34 Vertical cross-sections of $[\Psi]$, $[u^*v^*]$, $[u]$ and 10-day $d[u]/dt$ regressed upon the daily time series of 150-hPa $[u]_{0NS}$ based on MJO-residual data ........................................ 133

4.35 As in Fig. 4.34 but regressed upon the daily time series of 150-hPa $d[u]/dt_{0NS}$ .......................................................... 133

4.36 Time-latitude sections of 150-hPa $[u]$, $[\Psi]$ and $[u^*v^*]$ regressed upon the linear combinations of 150-hPa $[u]$ and $d[u]/dt$ at (a) 25°N and (b) 25°S during Nov-Mar (from which Nov-Mar means for each year have been subtracted out) .......................................................... 134

4.37 Precipitation, 150-hPa geopotential height and wind fields regressed upon the linear combinations of 150-hPa $[u]_{25N}$ and 10-day $d[u]/dt_{25N}$ during Nov-Mar (from which Nov-Mar means for each year have been subtracted out) .......................................................... 135

4.38 Time-longitude sections of 150-hPa $u_{0NS}$ and $\Phi_{26NS}$ regressed upon the linear combinations of 150-hPa $[u]_{25N}$ and 10-day $d[u]/dt_{25N}$ during Nov-Mar (from which Nov-Mar means for each year have been subtracted out) .......................................................... 136

4.39 As in third panel of Fig. 4.37 but based on daily data with Nov-Mar means for each year not subtracted out ........................................ 137

4.40 The 150-hPa $u$ field regressed upon the linear combinations of 150-hPa $[u]_{25N}$ and 10-day $d[u]/dt_{25N}$ during Nov-Mar (from which Nov-Mar means for each year have been subtracted out) ........................................ 138

4.41 As in Fig. 4.37 but regressed upon the linear combinations of 150-hPa $[u]_{25S}$ and 10-day $d[u]/dt_{25S}$ during Nov-Mar ........................................ 139

4.42 As in Fig. 4.40 but regressed upon the linear combinations of 150-hPa $[u]_{25S}$ and 10-day $d[u]/dt_{25S}$ during Nov-Mar ........................................ 140

4.43 Vertical cross-sections of $[\Psi]$, $[u]$, 10-day $d[\Psi]/dt$ and $d[u]/dt$ regressed upon the daily time series of the equatorial convergence of 150-hPa $[u^*v^*]$ .................. 141

4.44 Precipitation, 150-hPa geopotential height and wind fields regressed upon the linear combinations of $P$ and 10-day $dP/dt$ ........................................ 142

4.45 Climatological annual mean and r.m.s. amplitudes of $[u]$, $[v]$, $[\Psi]$ and $[u^*v^*]$ based on the dry dynamical core model ........................................ 143

4.46 As in Fig. 4.45 but for the moist dynamical core model ........................................ 144

4.47 As in Fig. 4.45 but for the aqua-planet version of the AM2 model ........................................ 145
4.48 As in Fig. 4.45 but for the AM2 model with topography and climatologically-varying SSTs ................................................................. 146
4.49 Climatological annual mean precipitation and 157-hPa wind fields for the full version of the AM2 model .................................................. 147
4.50 Vertical cross-sections of $[\Psi]$, $[u*v*]$, $[u]$ and 2-day $d[u]/dt$ regressed upon PC 1 of nonseasonal $[\Psi]$ for the aqua-planet AM2 model .......... 148
4.51 Vertical cross-sections of $[\Psi]$, $[u*v*]$, $[u]$ and 10-day $d[u]/dt$ regressed upon PC 2 of nonseasonal $[\Psi]$ for the aqua-planet AM2 model .......... 148
4.52 As in Fig. 4.50 but for the full AM2 model ........................................ 149
4.53 As in Fig. 4.51 but for the full AM2 model ........................................ 149
4.54 EOFs 1-3 of nonseasonal $[v]$ for the full AM2 model ....................... 150
4.55 Vertical cross-sections of $[v]$ and $[\Psi]$ regressed upon IMFs of PC 1 of nonseasonal $[v]$ for the full AM2 model ....................................... 151
4.56 Vertical cross-sections of $[u]$ regressed upon PC 1 and 2 of the nonseasonal, zonally-averaged angular momentum field (46°N-46°S, 74-975 hPa) for the aqua-planet and full versions of the AM2 model ........................................... 152
4.57 Vertical cross-section of the correlation coefficients between the daily nonseasonal time series of $[u]$ at the same latitude of the Norther and Southern Hemispheres at each pressure level for the aqua-planet and full versions of the AM2 model ........................................................................... 153
5.1 Vertical cross-sections of the time-mean $[T]$ and $[u]$ for axisymmetric model simulations with and without a forcing ........................................... 160
A.1 Vertical cross-sections of the radiative equilibrium temperature ($T_{eq}$) field of the dry dynamical core model in sigma and pressure coordinates .......... 182
A.2 Number of levels within a pressure layer based on the formula used in the models of Held and Suarez (1994) vs. Polvani and Kushner (2002) .............. 183
A.3 Vertical cross-section of $T_{eq}$ with a polar vortex in the Northern Hemisphere .. 184
A.4 Vertical cross-section of $T_{eq}$ with constant tropical temperatures .................. 185
A.5 Vertical cross-sections of the time-mean $[T]$ for 3D and 2D simulations with latitudinally-varying and constant tropical $T_{eq}$ profiles .......................... 186
A.6 As in Fig. A.5 but for $[u]$ .................................................................. 187
A.7 Vertical profiles of $T_{eq}$ over the equator with varying static stability .............. 188
<table>
<thead>
<tr>
<th>Table Number</th>
<th>Description</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>3.1</td>
<td>Correlation coefficients between monthly mean time series of the various BDC indices and EP flux index</td>
<td>72</td>
</tr>
<tr>
<td>3.2</td>
<td>Correlation coefficients between monthly anomaly time series of the various BDC indices and EP flux index</td>
<td>73</td>
</tr>
<tr>
<td>3.3</td>
<td>As in Table 3.2 but for high-pass filtered data</td>
<td>73</td>
</tr>
<tr>
<td>3.4</td>
<td>As in Table 3.2 but for low-pass filtered data</td>
<td>73</td>
</tr>
<tr>
<td>4.1</td>
<td>Total variance of the sum of IMFs 1-11 and percent contribution of the IMFs to the total variance of the various time series</td>
<td>154</td>
</tr>
<tr>
<td>4.2</td>
<td>Ratio of the r.m.s. amplitudes of the two components of the total ( u<em>v</em> ) at 10°N/S, 150 hPa and 45°N/S, 300 hPa</td>
<td>155</td>
</tr>
<tr>
<td>A.1</td>
<td>A summary of the specifications of the models used in this study</td>
<td>189</td>
</tr>
</tbody>
</table>
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DEDICATION

To my grandmothers, 上山由恵 and 名方きみ子,
and in loving memory of my grandfathers, 上山善紀 and 名方大介.
Chapter 1
Introduction

The *general circulation* of the atmosphere, as defined in this study, is the collection of quantitative statistical properties of atmospheric fluid motions that are driven by the uneven distribution of heating in the atmosphere. Understanding the relationships between the climatological-mean and time-varying fields of winds, temperature and various other descriptors of the atmosphere is one of the primary objectives of general circulation studies. An accurate and detailed depiction of the spatial structure and temporal variability of these atmospheric fields is essential for achieving this goal. The division of the three-dimensional flow into “annular” (zonal mean) and “eddy” (departures from the zonal mean) components is a compact and useful approach for this purpose. Extensive efforts have been made over many decades to examine the structure and variability of the annular circulation and their relationships to the eddies, as summarized in the historical review of the atmospheric general circulation by Lorenz (1967).

Much of the work on the general circulation thus far have focused on the extratropics, owing to the relative abundance of observations over the Northern Hemisphere continents. Our understanding of the extratropical annular modes (e.g., Thompson and Wallace 2000) and the eddy-zonal mean flow interactions at those latitudes (e.g., Lorenz and Hartmann 2001, 2003), among others, has been well established for some time. The Tropics – hereafter defined as the region equatorward of $\sim 30^\circ$ latitude – is one of the remaining frontiers of general circulation studies because it is the region in which wind and temperature fields are least clearly defined. The analysis of the tropical circulation is also somewhat more complex than that of the extratropics, for example, because quasi-geostrophic theory becomes inapplicable as one approaches the equator (i.e., as the Coriolis parameter approaches zero and the Rossby number becomes large). We have yet to fully comprehend how the tropical general circulation is maintained, how it influences the global climate, and how it may change in the future.
This study aims to fill the gap in our understanding of the atmospheric general circulation by examining the structure and variability of the tropical circulation.

One of the first explorations of the dynamics of low-latitude annular variability was in the context of the quasi-biennial oscillation (QBO: Reed 1960; Ebdon 1960), which dominates the variability in the tropical stratosphere. Early investigations of the QBO established that (i) the zonal-mean zonal wind anomalies are in geostrophic balance to well within 1° of latitude of the equator (Reed 1960; Wallace and Holton 1968), (ii) time-varying mean meridional circulations (MMC) are instrumental in maintaining thermal wind balance between the QBO-related zonal wind and temperature fields (Reed 1960; Wallace and Holton 1968), and (iii) QBO-related zonal wind anomalies are forced by the convergence of eddy fluxes of zonal momentum associated with vertically-propagating planetary and gravity waves (Lindzen and Holton 1968; Holton and Lindzen 1972; Dunkerton 1997).

Less well understood is the dynamics of the variability in the Brewer-Dobson circulation. Discovered on the basis of water vapor (Brewer 1949) and ozone (Dobson 1956) measurements more than half a century ago, this thermally-indirect, global-scale circulation cell in the stratosphere, with ascent at low latitudes and descent at higher latitudes, has since been recognized as playing an important role in the stratospheric and global climate, as discussed in the next section of this chapter. However, due to the limited availability of reliable data sets of the global stratosphere, there remains a controversy over the forcing of variations in tropical upwelling in the Brewer-Dobson circulation on various time scales. We will present some new observational results relating to this issue in Chapter 3 (see also Ueyama and Wallace 2010).

The discovery of the thermally-direct circulation cell in the tropical troposphere dates back even further, in some sense, to the classical paper by Hadley (1735), although Hadley was not aware that the cell is restricted to latitudes equatorward of 30°. The climatological mean structure of the Hadley cells and their variability over the course of the seasonal cycle has been the topic of many subsequent studies, as discussed in Section 1.3. The less emphasized nonseasonal variability (i.e., the departures from the
climatological mean seasonal cycle) of the tropical tropospheric MMC is examined in Chapter 4, based on an analysis of daily data from global reanalysis and simulations of the tropical circulation using models of varying complexity.

With the exception of the studies on the zonal mean Hadley cells, much of the existing literature on the variability of the tropical tropospheric circulation is concerned with zonally-asymmetric features of the flow. Many of these studies are regional in scale, such as those focused on the monsoon systems that vary on a seasonal time scale, or in relation to a particular phenomenon or regime, such as the Madden-Julian Oscillation (MJO: Madden and Julian 1971, 1972) and the El Niño Southern Oscillation (ENSO: Walker 1924, Bjerknes 1969) that contribute to the variability on the intraseasonal and interannual time scales, respectively. Most of the studies that deal with the annular variability of the tropical general circulation have been framed in terms of variations in the atmospheric angular momentum (e.g., Kang and Lau 1990, 1994; Weickmann et al. 1992). There have been relatively few investigations of the tropical annular variability for its own sake, a topic we explore in Chapter 4.

In the remainder of this chapter, we provide brief descriptions and background on the stratospheric Brewer-Dobson circulation and tropical tropospheric annular variability that are relevant to this study.

1.1 The Brewer-Dobson circulation

The Lagrangian mean meridional circulation that ventilates the stratosphere, commonly referred to as the Brewer-Dobson circulation (BDC), is marked by low-latitude ascent and high-latitude descent (Brewer 1949; Dobson 1956; Andrews et al. 1987). The BDC exerts a large influence on the chemical make-up of the stratosphere. For example, the BDC not only redistributes ozone within the stratosphere by transporting it from its photochemical source region in the tropics to the polar regions, but the vertical motions in the BDC also control the cross-tropopause exchange of ozone-poor and ozone-rich air. The upward transport of chemically and radiatively active aerosols and trace gases (e.g., water vapor, carbon dioxide, methane, nitrogen oxide,
chlorofluorocarbons) through the tropical tropopause alters the thermal structure and trace gas distribution of the middle atmosphere.

Also of critical importance is the adiabatic cooling and warming associated with the ascending and descending branches of the BDC, respectively, that drive temperatures away from radiative equilibrium. Due to this dynamical effect of the BDC, temperatures in the tropical lower stratosphere are maintained below their radiative equilibrium, while temperatures in the polar winter stratosphere are warmer than would be expected on the basis of radiative equilibrium calculations (Andrews et al. 1987; Gettelman et al. 2004; Corti et al. 2005). Tropical lower-stratospheric temperatures control the stratospheric water vapor budget by “freeze-drying” the upwelling air to the saturation vapor pressure of the temperature minimum or cold point (Mote et al. 1996; Dessler 1998). The radiative heating by stratospheric water vapor, in turn, impacts stratospheric temperature (Forster and Shine 1999; Shine et al. 2003), possibly inducing a positive feedback in the tropical lower stratosphere (i.e., increased water vapor causes a warming of the tropical lower stratosphere, which allows for the upward transport of air with higher water vapor content). Water vapor concentration in the stratosphere also influences the rate of polar ozone loss through its effect on the formation of polar stratospheric clouds (Kirk-Davidoff et al. 1999). Long-term trends in the strength of the BDC and the associated rate of ventilation of the stratosphere have been the focus of studies of Butchart and Scaife (2001), Sigmond et al. (2004), Eichelberger and Hartmann (2005), Butchart et al. (2006), Li et al. (2008), Garcia and Randel (2008) and Fu et al. (2010).

Despite the importance of the tropical upwelling for global climate, our understanding of its behavior is far from complete. The traditional definition of the tropical tropopause at a single pressure level (either the cold point or a discontinuity in the lapse rate defined by the World Meteorological Organization) has, in recent decades, evolved into one that portrays the tropopause as an atmospheric layer of finite depth with properties of both tropospheric and stratospheric air. While this concept of the so-called “tropical tropopause layer” (TTL) is now widely accepted and studied extensively, as summarized in a review paper by Fueglistalter et al. (2009a), it is still unclear to what
extent the air within the TTL is influenced by the BDC from above and by the deep convection from below.

Recent satellite measurements of temperature, cloud, water vapor, and various chemical species, in conjunction with global reanalysis data sets that became available in the mid 1990s, allow us to examine the TTL region in the context of the structure and variability of the tropical general circulation. In Chapter 2, we show that there exists a clear distinction between the tropospheric Hadley circulation and the stratospheric BDC, with regard to the meridional width of the tropical upwelling and the equatorial symmetry of the structure of the seasonal cycle. We then explore the dominant forcing of variations in tropical upwelling in the BDC in Chapter 3. Our empirical results indicate that tropical upwelling is largely modulated by variations in the high-latitude wave forcing across a wide range of frequencies, but simulations in an idealized model fail to confirm this result. A brief description of the theory of the mechanical forcing of the thermally-indirect BDC is offered below.

1.1.1 THE DOWNWARD CONTROL PRINCIPLE

The tropical upwelling in the BDC is often interpreted in terms of an extratropical suction pump that operates in accordance with the downward control principle (Haynes et al. 1991; Holton et al. 1995). Wave drag upsets thermal wind balance by decelerating the westerlies in the dissipation region. Thermal wind balance is maintained by the development of a MMC cell, with ascent at lower latitudes and descent at higher latitudes.

The transformed Eulerian-mean (TEM) residual circulation is often used as a proxy for the Lagrangian-mean motion in the stratosphere (Andrews et al. 1987), which cannot be directly measured. Within this framework, tropical upwelling can be estimated in one of two ways (Holton 1990; Rosenlof and Holton 1993; Rosenlof 1995; Randel et al. 2002a; Randel et al. 2008): by solving (i) the TEM thermodynamic energy equation or (ii) the TEM zonal momentum balance equation. Both methods have advantages as well as limitations. For example, the former, based on quasi-geostrophic TEM thermodynamic
energy balance (Eqn. 1.1), requires an accurate radiative transfer model in which satellite and assimilated data are used to estimate the zonally-averaged radiative heating rate \( Q \). Accurate temperature measurements with high vertical resolution are also needed to correctly capture the shallow layer of the cold point tropopause. Vertical velocities calculated via this method are reasonably accurate at all latitudes above \( \sim 100 \) hPa, where radiation dominates the thermodynamic forcing. Following Eqn. 5 of Randel et al. (2008), the thermodynamic energy equation can be written as

\[
\frac{\partial [T]}{\partial t} + \left[ v^T \right] \frac{\partial [T]}{\partial y} + \left[ w^T \right] S = [Q] \]  

(1.1)

Square brackets ([ ]) denote zonal averages and superscript \( T \) represent TEM variables. \( S \) is the static stability term \( S = H N^2/R \) where \( H \) is scale height, \( N^2 \) is the Brunt-Väisälä frequency squared, and \( R \) is gas constant), and all other terms are in their standard notation.

The momentum balance approach can be implemented as follows. The zonally-averaged vertical velocity at height \( z \) averaged between latitudes \( \phi_1 \) and \( \phi_2 \) based on the TEM zonal momentum balance, following Eqn. 2 of Randel et al. (2008), is given by:

\[
\left\langle w^T \right\rangle = \int_{\phi_1}^{\phi_2} \frac{e^{z/H}}{a \cos \phi} \left\{ \cos \phi - \frac{\partial [u]}{\partial t} f \right\} D \Phi \left( X - \frac{\partial [u]}{\partial t} f \right) \left( e^{-z/H} d\zeta \right) \right\rangle_{\phi_1}^{\phi_2} \]  

(1.2)

Angle brackets (< >) denote an average between latitudes \( \phi_1 \) and \( \phi_2 \). \( a \) is the radius of the Earth, \( f \) is the Coriolis term, and \( X \) represents any unresolved forces (e.g., gravity wave drag). \( D \Phi \) is the scaled Eliassen-Palm (EP) flux divergence

\[
D \Phi = \frac{e^{z/H}}{a \cos \phi} \nabla \cdot \tilde{F} \]  

(1.3)

where \( \tilde{F} \) is the EP flux vector with the following horizontal and vertical components:
\[ F_\phi = e^{-z/H} a \cos \phi \left( [-u \ast v \ast] + \frac{\partial [u]}{\partial z} \left[ \frac{v \ast T \ast}{S} \right] \right) \]  

(1.4)

\[ F_z = e^{-z/H} a \cos \phi \left( f \left[ \frac{v \ast T \ast}{S} \right] + [u \ast w \ast] \right) \]  

(1.5)

This method of calculating the TEM residual vertical velocity has the advantage of requiring only the zonally-averaged zonal wind ([u]) and eddy covariance statistics ([u*v*], [v*T*], [u*w*]), which are well represented by the global reanalysis data sets and easily calculated. However, the crude approximation of subgrid-scale eddy forcing in X, often computed as the residual from all other terms in the momentum equation, may not be reliable in regions where small-scale waves (e.g., orographically-forced gravity waves) are prominent. In addition, vertical velocity estimates become problematic near the equator due to uncertainties in the eddy statistics and the proportionality to 1/f. Equatorial upwelling is thus estimated as an average over an equatorial belt of finite width, usually extending out to at least \( \sim 15^\circ \)N/S.

Equation 1.2 is a mathematical representation of the downward control principle. We note that the net integrated force above the level of forcing determines the vertical motion at that level, rather than the force per unit mass at that level: hence the term “downward control”.

### 1.2 Tropical troposphere

Hadley (1735) postulated the existence of a thermally-direct MMC with rising motion near the equator and sinking motion in the polar regions. In this seminal paper, he concluded that the circuit, completed by equatorward flow at low levels and poleward flow aloft, gives rise to the surface easterly winds in its lower branch. Observations almost a century later revealed that the predominant southwesterly winds in the Northern Hemisphere midlatitudes requires a secondary thermally-indirect cell in higher latitudes (Dove 1837; Maury 1855), which has come to known as the Ferrel cell (Ferrel 1856, 1859; Thompson 1857, 1892). As a result, the Hadley cell was redefined to represent the
dominant MMC in the tropics. The notion of the eddies influencing the zonal mean motion by transporting various properties from one latitude to the next did not emerge for another century or so. By then, estimates of the intensity of the observed Hadley cell (Palmén and Vuorela 1963; Vuorela and Tuominen 1964; Kidson et al. 1969; Oort and Rasmusson 1970) and its eddy-driven component (Gilman 1965) had led to a general consensus on the existence and structure of the tropical MMC.

Following these earlier works, more comprehensive data were acquired, allowing for a detailed depiction and diagnosis of the tropical MMC. Relevant to this study are the works of the past few decades concerning the seasonal cycle of the Hadley cells. Observational results of Oort and Rasmusson (1970) indicated the prevalence of a single, cross-equatorial cell throughout much of the year. Lindzen and Hou (1988) argued that the equatorial symmetry of the pair of cells straddling the equator, exemplified by those in the annual mean streamfunction pattern shown in Fig. 1.1, is largely an artifact of the time-averaging over the year, and that solsticial regimes with strong equatorial asymmetry are the favored configuration of the tropical MMC. Dima and Wallace (2003) showed that an equally valid interpretation of the observations involves the equatorially-symmetric cells that are present all year round with an amplitude comparable to that of the equatorially-asymmetric pattern, an interpretation supported by the subsequent theoretical study of Walker and Schneider (2005). We show in Chapter 4 that equatorially-symmetric and -asymmetric patterns of the seasonal variability in the tropical MMC are also the preferred patterns of the nonseasonal variability in the global MMC, and that it is the equatorially-symmetric cells that are strongly coupled to the annular variability of the zonal wind field.

The tropical tropospheric $[u]$ field exhibits interesting structures both in the annual mean and in the variability about the mean, such as the climatological mean easterlies over the equator in Fig. 1.1 and the poleward propagation of low-latitude $[u]$ anomalies documented by Riehl et al. (1950), Feldstein (1998), and others. A diagnosis of the structure and variability of the tropical $[u]$ field requires not only an understanding of
the role of the MMC, but also of the interrelationships between \([u]\), MMC and the eddy fluxes of zonal momentum.

Lee (1999) showed that the climatological mean easterlies in the equatorial upper troposphere result from the balance between the equatorial convergence of eddy momentum fluxes and the advection of easterly momentum by the seasonally-varying, equatorially-asymmetric MMC. The equatorial superrotation predicted in idealized models as a response to the equatoward momentum flux induced by equatorial planetary waves (e.g., Kraucunas and Hartmann 2005) is not observed in the presence of seasonally-varying, cross-equatorial MMC (Dima et al. 2005). A comprehensive climatology of the climatological-mean equatorial planetary waves presented by Dima and Wallace (2007) offers a clear representation of the eddies that are responsible for transporting momentum into the equatorial belt. On a horizontal plane in the upper troposphere, these waves assume the form of the canonical equatorial wave pattern identified by Matsuno (1966) and Gill (1980), which consists of equatorially-symmetric Kelvin and Rossby waves to the east and west of the equatorial heat source, respectively. In the zonal mean, these waves appear as a couplet of equatoward eddy momentum fluxes straddling the equator at \(\sim 150\) hPa level, as seen in Fig. 1.1.

Many of the mechanisms proposed for explaining the poleward propagation of \([u]\) anomalies invoke the role of the eddy momentum fluxes at extratropical latitudes: the role of the transient eddy momentum fluxes (Feldstein 1998), variations in the extratropical annular modes (Thompson and Lorenz 2004), and the equatorward propagation and breaking of midlatitude waves (Lee et al. 2007). The tropical \([u]\) field may indeed be modulated by forcing from higher latitudes as these studies suggest, but the possibility of locally-forced tropical variability cannot be entirely ruled out, given that the fluctuations in precipitation over the warm pool region or the equatorial cold tongue in the eastern Pacific associated with the seasonal cycle, the MJO, and ENSO, all affect the structure of the equatorial planetary waves (Dima and Wallace 2007).

Weickmann et al. (1997) interpreted the intraseasonal oscillations in atmospheric angular momentum in terms of the modulation of the stationary waves by the MJO.
Watanabe et al. (2002) identified a pattern of the tropical annular variability that projects strongly upon the ENSO pattern, but they concluded (on the basis of the robustness of the pattern in the ENSO-residual data) that their “tropical axisymmetric mode” pattern represents an intrinsic mode of atmospheric variability. It appears likely, as seen in a case study by Hsu et al. (1990), that the interaction between the tropics and the extratropics is an integral part of the MJO-related tropical intraseasonal oscillation. The relevant tropical-extratropical linkages however have yet to be fully elaborated. Our analysis in Chapter 4 attempts to add some insights on this issue.

We also show in Chapter 4 that intraseasonal variations in $[u]$ at low latitudes exhibit a tendency for equatorial symmetry. How might the modulation of the equatorial planetary waves, whose prominent equatorial symmetry has been noted (Dima and Wallace 2007) and explored (Kraucunas and Hartmann 2007), influence the symmetry properties of the tropical $[u]$ variability? The role of the eddy-driven MMC has been invoked to explain the symmetric variability of the climate system on the interannual time scale (Seager et al. 2003a), and in this study, we offer a dynamical interpretation of the role of the climatological-mean equatorial planetary waves in leveraging the intraseasonal variations in the equatorial planetary waves to produce large variations in the equatorward eddy fluxes of zonal momentum, which ultimately drive the equatorially-symmetric variability of $[u]$ in the equatorial upper troposphere.

### 1.3 Outline of the dissertation

We begin in Chapter 2 with a description of the zonal mean structures of the climatological annual mean and seasonal cycle of the tropical atmosphere, emphasizing the contrasting features of the thermally-direct tropospheric Hadley circulation and the thermally-indirect stratospheric Brewer-Dobson circulation. Chapter 3 examines the structure and variability of the stratospheric BDC on the intraseasonal to interannual time scales. In Chapter 4, we investigate the leading patterns of the nonseasonal variability in the tropospheric MMC, and use them as a basis for categorizing the nonseasonal annular variability of winds in the tropical troposphere. Although this study is mainly
observational, numerical models are used in several instances to test our interpretation of our empirical findings. Chapter 5 concludes with a summary of the major findings and an outlook for future research.
Fig. 1.1: Climatological-mean (a) annual mean and (b) root-mean-squared amplitudes of the zonally-averaged variables, as indicated in a, based on daily ERA-Interim data during 2000-08. Contour intervals in a and b are 5 and 1 m s\(^{-1}\) for \([u]\), 0.5 and 0.1 m s\(^{-1}\) for \([v]\), 2 x 10\(^{10}\) and 0.5 x 10\(^{10}\) kg s\(^{-1}\) for \([\psi]\), and 5 and 10 m\(^2\) s\(^{-2}\) for \([u*v*]\), respectively. Dashed contours denote negative values. Gray shading in b indicate values ≥ 5 m s\(^{-1}\) for \([u]\), ≥ 0.5 m s\(^{-1}\) for \([v]\), ≥ 3 x 10\(^{10}\) kg s\(^{-1}\) for \([\psi]\), and ≥ 50 m\(^2\) s\(^{-2}\) for \([u*v*]\).
Chapter 2

Structure of the Tropical Mean Meridional Circulation

2.1 Introduction

Vertical velocities above the level of main convective outflow are orders of magnitude smaller than mid-tropospheric values, resulting in large uncertainties in the structure and variability of the upwelling within the tropical tropopause layer (TTL). The TTL plays an important role in Earth’s radiative balance not only by setting up the chemical boundary conditions for the stratosphere, but also by modifying clouds at their base (Highwood and Hoskins 1998; Gettelman and Forster 2002; Fueglistaler et al. 2009a). Hence, clear depictions of the convectively-driven, thermally-direct tropospheric Hadley circulation and its transition to the eddy-driven, thermally-indirect stratospheric Brewer-Dobson circulation (BDC) are needed.

A distinctive feature of the TTL is the cold point tropopause that encompasses latitudes equatorward of ~20° both in the annual mean and the variability about the mean. Its pronounced seasonal cycle, first documented by Reed and Vlcek (1969), appears to be an adiabatic temperature (cooling) response to the seasonally-varying strength of tropical upwelling (e.g., Andrews et al. 1987). Air that ascends through the cold point is dehydrated, which gives rise to a water vapor “tape recorder” signal in the tropical lower stratosphere (Mote et al. 1996; Randel et al. 2001; Schoeberl et al. 2008).

The strength of tropical upwelling more directly affects the concentrations of chemical tracers with large vertical gradients. Seasonal cycles of the mixing ratios of tropical ozone (Shiotani and Hasebe 1994; Schoeberl et al. 2008) and carbon monoxide (Schoeberl et al. 2006; Schoeberl et al. 2008) have been explained in terms of seasonal variations in upwelling associated with high altitude convective outflow (Folkins et al. 2006) and by variations in the strength of the dynamically-forced upwelling in the BDC acting on the strong background vertical gradients (Randel et al. 2007).

Zonal-mean tropical tropospheric cloud fraction undergoes an annual cycle, with increased cloudiness in the summer hemisphere (e.g., Jakob 1999; Virts and Wallace...
2010a). Zhang (1993) demonstrated that, unlike convection and lower tropical cirrus (Liao et al. 1995), the seasonal cycle of tropical clouds with cloud-top temperatures below 200 K is in phase with that of tropical cold point temperature; this result has been confirmed for subvisual clouds (Wang et al. 1996) and TTL cirrus (Virts and Wallace 2010a).

In this chapter, we will examine the structures of the climatological-mean annual mean and seasonal cycle of the zonally-averaged tropical (40°N-40°S) circulation from ~50 hPa to 1000 hPa based on two reanalysis data sets. For comparison, we superimpose onto the circulation statistics the zonal-mean fields of satellite-derived cloud fraction and mixing ratios of water vapor (WV), ozone (O\textsubscript{3}) and carbon monoxide (CO) equatorward of 30°N/S. The TTL has been characterized by one or more of these fields, which are summarized in Fueglistaler et al. (2009a), but here we broaden the scope of the analysis beyond the TTL region.

### 2.2 Data

Temperature and meridional wind are based on the European Centre for Medium-Range Weather Forecasts (ECMWF) Interim reanalysis (ERA-Interim: Simmons et al. 2007) and the Japanese 25-year Reanalysis (JRA-25: Onogi et al. 2007). We analyzed once daily (00Z) ERA-Interim data gridded at 1.5° latitude by 1.5° longitude from 40°N to 40°S and 16 vertical levels between 50 hPa and 1000 hPa (i.e., 50, 70, 100, 125, 150, etc.) during 12 Jun 2006 – 11 Jun 2009. We also analyzed monthly mean JRA-25 data from Jan 1979 to Dec 2006 in the domain 40°N to 40°S at 1.1° latitude by 1.1° longitude resolution and 31 vertical levels from 42 hPa to 995 hPa (i.e., 42, 55, 71, 89, 109, 131, 155, etc.). Variables are averaged over latitude circles to compute the zonal means. The streamfunction field for the mean meridional circulation (MMC) is obtained from a vertical integration of the zonally-averaged meridional winds, as described in Dima and Wallace (2003).

Vertical velocities from ERA-Interim, JRA-25, and the 40-year ECMWF Reanalysis (ERA-40: Uppala et al. 2005) were compared to assess their reliability in
representing the vertical motions in the lower stratosphere. The climatological-mean, annual mean fields of JRA-25 and ERA-40 exhibit qualitatively similar magnitudes and spatial structures, but the corresponding field based on daily mean ERA-Interim exhibits a much less coherent pattern (not shown).

As a way of circumventing the uncertainties in the vertical velocity fields in the models, we infer the vertical velocities in the tropical lower stratosphere from radiative heating rate estimates. This inference is made possible by the fact that positive radiative heating is balanced by adiabatic cooling of ascending air, which maintains temperatures below their radiative equilibrium values, and vice versa (e.g., Andrews et al. 1987). A precise estimate of motions in the BDC requires Lagrangian mean statistics, which are virtually impossible to measure, but the Eulerian counterparts should be a reasonable approximation within the tropics. We make use of the sum of shortwave and longwave heating rates in the 50-150 hPa layer calculated every 12 hours by the reanalysis forecast models, available every 1° latitude at 8 levels from 1 Apr 2006 to 31 Dec 2009 for ERA-Interim and every 2.5° latitude at 4 levels from Jan 1979 to Dec 2007 for JRA-25. The latitudes and pressure levels of the zonal mean heating rates were linearly interpolated to the grid locations of the meridional wind data for the respective reanalyses.

Vertical profiles of clouds are retrieved globally by an active lidar instrument aboard the polar-orbiting Cloud-Aerosol Lidar and Infrared Pathfinder Satellite Observations (CALIPSO) satellite, one of five satellites in the current Afternoon or “A-train” constellation of the National Aeronautics and Space Administration. “A-train” satellites pass through the tropics ~29-30 times per day. CALIPSO observations are available at 5 km horizontal (along-track) and 60 m vertical resolution at the altitudes of interest (Winker et al. 2007). Cloud layer detection algorithms are documented at http://www-calipso.larc.nasa.gov. For this study, cloud profiles in the domain 30°N-30°S from 13 June 2006 to 12 June 2010 were combined to create climatological monthly mean “cloud fraction” on a 1.5° latitude by 200 m grid. The zonally-averaged cloud fraction field for a given calendar month (e.g., Jan) represents the fraction of CALIPSO
profiles acquired during that month (e.g., Jan 2006-10) that identified a cloud layer at a given latitude and level.

The Microwave Limb Sounder (MLS) aboard the Aura satellite in the “A-Train” constellation measures thermal emissions in spectral bands centered on 190 and 240 GHz from the limb of the Earth’s atmosphere. Further details regarding MLS retrieval methods can be found at http://mls.jpl.nasa.gov. MLS variables analyzed in this study are climatological monthly-mean profiles of WV mixing ratio (at 11 pressure levels between 316 and 46 hPa) and O\textsubscript{3} and CO mixing ratios (at 5 pressure levels between 215 and 46 hPa) from 30°N to 30°S, calculated using profiles from the same 4-year period as the CALIPSO data.

2.3 Annual mean

The climatological-mean annual mean streamfunction field for the MMC (contours) are superimposed on zonally-averaged temperature (colored shading), meridional wind and radiative heating rates (vectors) in the domain 40°N-40°S and 50-1000 hPa in Fig. 2.1a, based on data from ERA-Interim and JRA-25. The annual mean structure of the streamfunction field is dominated by a pair of Hadley cells. The Southern Hemisphere (SH) cell is slightly stronger and wider than the Northern Hemisphere (NH) cell such that the boundary between the cells is shifted a few degrees north of the equator, corresponding to the annual mean position of the intertropical convergence zone (ITCZ). Both cells, the SH cell in particular, appear to be stronger in the ERA-Interim than in the JRA-25 by up to 50%.

The meridional width of upwelling in the tropical troposphere, as indicated by the ascending branches of the annual mean Hadley cells, extends from ~10°N to ~15°S. The flow field above ~125 hPa, shown in vectors, indicates a much broader upwelling of the BDC, which extends poleward to ~30-35°N/S. The sum of the longwave and shortwave heating rates from ERA-Interim are roughly twice as large as those of JRA-25 in the 70-125 hPa layer (~0.4 vs. ~0.2 K d\textsuperscript{-1}), primarily from the disparity in the longwave component (not shown). The pronounced off-equator maxima in the ERA-Interim heating
rates at 70 hPa are in agreement with Fueglistaler et al. (2009b). The transition from a narrow and concentrated upwelling in the troposphere to a broad and more uniform upwelling in the lower stratosphere, depicted in both reanalyses, confirms the notion that tropical upwelling above and below ~150 hPa is forced by different processes.

The tropical tropopause, with temperatures as low as ~193 K in the annual mean, appears as an equatorially-even (symmetric) layer centered at 100 hPa with a meridional width comparable to that of the positive heating rate vectors. That the cold point is much wider than the extent of the upwelling of the tropospheric Hadley cells suggests that ascent above ~100 hPa is unlikely to be forced from below by deep convection, but rather that it is forced from above by the BDC.

Next, we examine the annual mean structures of zonally-averaged cloud fraction and trace species superimposed on the circulation features of ERA-Interim, transcribed from those the left panel of Fig. 2.1a. The climatological-mean annual mean cloud fractions equatorward of 30°N/S, documented by Virts et al. (2010; their Fig. 2a) and reproduced here in colored shading in Fig. 2.2a, exhibit a pronounced maximum of ~30% in the upper troposphere above the plume of rising air over the latitude of the ITCZ (~7°N). The band of enhanced cloud fraction appears to spread out meridionally as it approaches the statically-stable stratosphere.

Zonally-averaged mixing ratios of WV in the annual mean, shown in Fig. 2.2b, exhibit a distinct minimum (~3.5 ppmv) at 68 hPa, presumably due to the upward transport of air imprinted with low mixing ratios at the cold point tropopause (Mote et al. 1996). We note that in-situ data with finer vertical resolution than those of the reanalyses show the cold point tropopause at ~90 hPa (e.g., Fueglistaler et al. 2009a), closer to the level of the WV minimum. There is little meridional gradient in the WV mixing ratios throughout the tropical lower stratosphere, as would be expected from the zonal mean temperature structure (Fig. 2.1a). The increase in WV above 68 hPa is believed to be due to the moistening by methane oxidation (e.g., Bates and Nicolet 1950; Le Texier et al. 1988).
The annual mean structures of zonally-averaged mixing ratios of O$_3$ and CO are shown in Figs. 2.2c and d, respectively. The large positive and negative vertical gradients of O$_3$ and CO reflect their stratospheric and tropospheric sources, respectively. Meridional gradients are generally small for both species at all levels in the annual mean.

2.4 Seasonal cycle

The structures of the climatological-mean seasonal cycles of the respective fields are represented in terms of differences between the Jan-Feb and Jul-Aug means. The peak-to-peak difference in the streamfunction field for the MMC, shown in Fig. 2.1b, is dominated by a single cell centered over the equator. The cells in the two reanalyses are roughly of the same magnitude (~4.5 x 10$^{11}$ kg s$^{-1}$). Radiative heating rates below 100 hPa (not shown) indicate anomalous upwelling in the SH and anomalous downwelling in the NH, mirroring the equatorially-odd pattern of the MMC in the troposphere. At higher levels, positive Jan-Feb minus Jul-Aug heating differences prevail across ~30°N-40°S, which appear to be associated with a broad, equatorially-even temperature pattern with peak-to-peak amplitude of ~7 K at the 70-hPa level.

Figure 2.3 shows vertical cross-sections of Jan-Feb minus Jul-Aug fields of cloud fraction and mixing ratios of WV, O$_3$ and CO, superimposed on the corresponding streamfunction field based on the ERA-Interim data. To enhance the contrasting features in panels b-d, seasonal amplitudes are displayed as percentages of their respective annual mean values in Fig. 2.2. The seasonal cycle in tropical cloud fraction, adapted from Virts and Wallace (2010; their Fig. 6a) and shown in colored shading in Fig. 2.3a, exhibits anomalously high cloud fractions in the rising branch of the Hadley circulation and anomalously low cloud fractions in the sinking branch. Cloud amounts decrease rapidly above the 100-hPa level, but the meridional spreading of the signal in the 70 hPa to 100 hPa layer suggests that cloud fraction above the level of main convective outflow is at least partly influenced by tropical upwelling in the BDC.

Consistent with the pattern of cloud fraction profiles in Fig. 2.3a, WV mixing ratios in Fig. 2.3b indicate anomalously moist air in the rising branch of the anomalous
cell, and vice versa in the sinking branch. Above ~120 hPa, the WV signature becomes primarily equatorially even, in agreement with the heating rate profiles in the lower stratosphere. It appears that air imprinted with low WV mixing ratios at ~100 hPa travels upward over the course of several months, causing a phase shift of the annual cycle with increasing altitude, such that the variations become out-of-phase at 56 hPa.

Seasonal cycles of the zonally-averaged mixing ratios of O$_3$ and CO, shown in Fig. 2.2c and d, generally agree with the results of other fields with respect to the contrasting equatorial symmetry in the troposphere and stratosphere. Anomalously low O$_3$ and high CO, characteristic of tropospheric air, are observed in the rising branch of the Hadley cell, and vice versa in the sinking branch. Meridional profiles above 100 hPa are broader and more uniform across latitudes; anomalies of like sign are observed in the tropical belt at any given level.

The largest seasonal amplitudes of O$_3$ are observed in the 68-100 hPa layer (Fig. 2.3c), which is higher than would be expected solely on the basis of the height of the anomalous tropospheric cell. Consistent with the findings of Randel et al. (2007), this feature appears to be due to the anomalous upwelling acting on the large vertical gradient of O$_3$ which peaks at ~68 hPa (not shown). Similarly, the large negative vertical gradient of CO above 100 hPa (not shown) is likely to be responsible for the peak seasonal amplitudes in Fig. 2.3d. Evidence of the CO tape recorder noted by Schoeberl et al. (2006) accounts for the out of phase relationship between the annual cycles at 68 hPa and 44 hPa. The skewness of the seasonal peaks in O$_3$ and CO towards the NH may be a reflection of the slightly stronger upwelling at ~10°N than ~10°S at the 68-hPa level, which is implied by the magnitude of the heating rates. Chemistry appears to be playing a role in determining the pattern of the mixing ratios of O$_3$ and CO in the upper troposphere (~147 hPa), as acknowledged in previous studies (Folkins et al. 2006; Schoeberl et al. 2006).
2.5 Summary

The zonal mean structure of the climatological-mean annual mean and seasonal cycle of the tropics was examined based on temperature, meridional wind and radiative heating rates from two reanalysis data sets and satellite-derived measurements of cloud, WV, O$_3$ and CO. The agreement between the patterns derived from multiple data sources illustrates the clear distinction between the tropospheric Hadley circulation and the stratospheric BDC, with regard to the meridional width of the tropical upwelling and the equatorial symmetry of the Jan-Feb minus Jul-Aug structure.

In the annual mean, ascent within the tropical troposphere, indicated by the streamfunction field for the MMC, is confined to the latitude belt ~15°N-10°S. The influence of the vertical motions in the MMC on the background constituents is clearly discernible up to ~125 hPa, especially in the zonally-averaged cloud fraction field, which shows high cloud amounts in the plume of rising air that appear to spread out to form an anvil just below the 100-hPa level, where the vertical velocities are weakening substantially with height. The relatively narrow tropical upwelling branch of the Hadley circulation is in striking contrast to the broad tropical upwelling in the BDC, inferred from the sum of the longwave and shortwave radiative heating rates of the reanalysis models. Positive heating rates in the lower stratosphere, indicative of ascent, encompass latitudes equatorward of ~30-35°N/S, corresponding to the broad meridional extent of the tropical cold point centered at 100 hPa. Air that ascends through the cold point is dehydrated and transported upward, giving rise to minimum WV mixing ratios at 70 hPa. Annual mean mixing ratios of O$_3$ and CO increase and decrease more or less monotonically with height, respectively.

The seasonal cycle, represented here as the difference between Jan-Feb and Jul-Aug means, of the Hadley circulation is accompanied by coherent changes in cloud fraction and mixing ratios of WV, O$_3$ and CO up to 147 hPa; the rising branch of the anomalous cell in the SH tropics is associated with features that are characteristic of tropospheric air (i.e., cloudy, moist air with low O$_3$ and high CO mixing ratios), whereas the sinking branch of the anomalous cell in the NH tropics exhibits features that are
characteristic of stratospheric air (i.e., cloud-free, dry air with high O\textsubscript{3} and low CO mixing ratios).

Seasonal cycles in the lower stratosphere exhibit patterns with even equatorial symmetry. The pronounced seasonal cycle in temperature at 70 hPa extends across the entire tropics along the latitudes of positive annual mean heating rates. Clouds appear to spread meridionally near the base of the statically-stable layer above the cold point. A broad, equatorially-even layer of anomalously low WV mixing ratios is observed at 100 hPa. The large seasonal cycles of O\textsubscript{3} and CO in the lower stratosphere also exhibit broad meridional structures and are collocated with the level of their strongest vertical gradients in their mixing ratios. The phase shift of the annual cycle with increasing altitude, indicative of the tape recorder phenomenon, is observed for WV and CO, but not for O\textsubscript{3}. Although the influence of quasi-biennial oscillation on the mixing ratios of WV and O\textsubscript{3} is discernible at 46 hPa, its effect on the climatological-mean annual mean and seasonal cycle patterns presented here is small (not shown). The lack of pronounced signatures of El Niño Southern Oscillation in tropical mean TTL cloud fraction (Virts and Wallace 2010) further suggests that the results of this study are robust, despite the fact that they are based on only four years of data.
Fig. 2.1: Vertical cross-sections of the climatological-mean (a) annual mean and (b) Jan-Feb minus Jul-Aug fields of zonally-averaged temperature (colored shading; K), streamfunction (contours; $2 \times 10^{10}$ kg s$^{-1}$ in a, $4 \times 10^{10}$ kg s$^{-1}$ in b), meridional wind component (horizontal vectors; m s$^{-1}$) and the sum of the shortwave and longwave radiative heating rates (vertical vectors; K d$^{-1}$) based on (left) once daily ERA-Interim data during 12 Jun 2006 – 11 Jun 2009 (1 Apr 2006 – 31 Dec 2009 for the heating rates) and (right) monthly JRA-25 data during Jan 1979 – Dec 2006 (Jan 1979 – Dec 2007 for the heating rates).
Fig. 2.2: As in left panel of Fig. 2.1a but for (a) CALIPSO cloud fractions and MLS volume mixing ratios of (b) water vapor, (c) ozone and (d) carbon monoxide in colored shading during 13 Jun 2006 – 12 Jun 2010.
Fig. 2.3: As in left panel of Fig. 2.1b but for (a) CALIPSO cloud fractions and MLS volume mixing ratios of (b) water vapor, (c) ozone and (d) carbon monoxide in colored shading during 13 Jun 2006 – 12 Jun 2010. Volume mixing ratios (b–d) are presented as percentages of the annual mean mixing ratios shown in Fig. 2.2.
Chapter 3

Tropical Stratospheric Circulation

3.1 Introduction

The existence of pronounced annual cycles in tropical cold point temperature (Reed and Vlcek 1969) and in the mixing ratios of water vapor, ozone, methane and other chemical species in the tropical lower stratosphere (Russell et al. 1993; Mote et al. 1995, 1996; Niwano et al. 2003; Folkins et al. 2006; Schoeberl et al. 2006; Randel et al. 2007; Schoeberl et al. 2008) all serve to indicate that tropical upwelling in the Brewer-Dobson circulation (BDC) is substantially stronger in January than in July. The BDC also exhibits more subtle nonseasonal variability on intraseasonal and interannual time scales, as discussed in the following sections.

The annual-mean BDC is thermally indirect, with upwelling of cold air in the tropics and descent of warmer air at high latitudes, and hence must be mechanically driven. Haynes and McIntyre (1987) and Haynes et al. (1991) argued that the BDC is driven by the breaking of vertically-propagating Rossby waves and gravity waves in the extratropics. It is widely believed that wave breaking in the extratropical stratosphere plays an important role in driving the BDC, but it is not clear how much of that forcing derives from the wintertime polar night jet region and how much of the forcing is due to wave breaking at lower latitudes. The downward control principle, described in Section 1.1.1, predicts that the time-mean response to wave forcing is local in the latitude domain, but that a wave forcing can exert a non-local control under transient conditions whereby the induced MMC extends horizontally away from the forcing region (Haynes et al. 1991; Holton et al. 1995). Thus it is conceivable that transient variations in the strength of the BDC and the associated variations in tropical upwelling could largely be driven by high-latitude wave forcing.

Randel et al. (2002a,b) showed observational evidence that tropical upwelling, as inferred from temperature and ozone perturbations, varies in response to week-to-week fluctuations in the high-latitude EP flux, in accordance with diagnostics based on the
downward control principle. Results of a concurrent observational study by Salby and Callaghan (2002) suggest that tropical upwelling varies in synchrony with high-latitude EP flux on the interannual time scale as well. Dhomse et al. (2008) showed evidence of a year-to-year modulation of tropical lower-stratospheric water vapor by mid- to high-latitude planetary wave driving as inferred from eddy heat fluxes at 50 hPa averaged from 45°N/S to 75°N/S. Iwasaki (1992), Yulaeva et al. (1994) and Chae and Sherwood (2007) argued that the annual cycle in tropical upwelling is associated with enhanced high-latitude planetary wave forcing during the northern winter. The statistically significant anti-correlations between high-latitude EP flux and tropical temperature tendencies documented in Fig. 5 of Salby and Callaghan (2002) and those between extratropical eddy heat fluxes and tropical lower-stratospheric water vapor mixing ratios documented in Fig. 2 of Dhomse et al. (2008) lend credence to this hypothesis.

Diagnostics of the forcing of tropical upwelling based on the downward control principle by Rosenlof (1995), Boehm and Lee (2003), and Kerr-Munslow and Norton (2006), however, give the general impression that high-latitude wave forcing does not influence tropical upwelling on a time scale of seasons or longer. To maintain and/or perturb tropical upwelling with a realistic magnitude and structure in numerical simulations of the BDC, it appears to be necessary to prescribe some form of wave drag that extends into subtropical latitudes (Plumb and Eluskiewicz 1999; Semeniuk and Shepherd 2001; Scott 2002; Zhou et al. 2006; Geller et al. 2008). As an alternative to high-latitude wave driving, Norton (2006) argued that seasonally-varying equatorial planetary waves and their associated deep convection could force an annual cycle in tropical upwelling. Extending this line of argument, Deckert and Dameris (2008) and Rosenlof and Reid (2008) have suggested that a secular trend toward higher tropical sea surface temperatures (SSTs) and more vigorous tropical convection could act to strengthen the BDC.

In the following sections of this chapter, we will show additional evidence that variations in high-latitude wave forcing affect the strength of tropical upwelling in the BDC on intraseasonal, annual, and interannual time scales. This work can also be found
in Ueyama and Wallace (2010). To simplify the analysis of the extratropical forcing of tropical upwelling, irrespective of the hemisphere in which the forcing occurs, we decompose the BDC-related fields into equatorially symmetric and asymmetric components. We show that equatorward of ~45°N/S, both the annual-mean temperature profile and the seasonal and nonseasonal variability about the mean project almost exclusively onto the equatorially-symmetric component, for which the distribution of upwelling and downwelling in the BDC can be inferred quite simply from the thermodynamic energy balance (i.e., from variations in the departures of the meridional temperature profiles from the respective radiative equilibrium temperature profiles). The inference is straightforward because temporal variations in radiative forcing (apart from volcanic eruptions) are strictly seasonal and project almost exclusively onto the equatorially asymmetric component. For example, the equatorially symmetric component of the meridional profiles of January and July radiative heating are identical, apart from the small effect of the eccentricity of the Earth's orbit. Hence, the rate of upwelling and downwelling in the equatorially symmetric component of the BDC can be determined from the observed temperatures without reference to the time-varying distribution of radiative equilibrium temperature.

3.2 Data and methods

The analysis is based on monthly, zonally-averaged brightness temperature ($T$) fields derived from the lower-stratospheric channel of Version 3.2 Microwave Sounding Unit (MSU) / Advanced Microwave Sounding Unit (AMSU) carried aboard National Oceanic and Atmospheric Administration (NOAA) satellites. The data are gridded at 2.5° x 2.5° resolution and extend to 87.5°N/S. The period of record 1979-2007 is used in this study. The weighting function of the lower-stratospheric channel is concentrated mainly in the 15 to 19 km (30 to 150 hPa) layer, as detailed in the Remote Sensing Systems Web site (see http://www.ssmi.com/msu/).

The monthly mean time series of global-mean $T$ (weighted by cosine of latitude) shown in Fig. 3.1a exhibits a well-defined annual cycle that is highly reproducible from
year to year, with the exception of those following the eruptions of El Chichón (1982) and Mt. Pinatubo (1991). As documented by Spencer et al. (1990), Christy and Drouilhet (1994) and Yulaeva et al. (1994), the climatological-mean annual cycle of global-mean $T$ exhibits a maximum between August and September and a minimum between January and February, with peak-to-peak amplitude of ~0.5 K. The post-eruption intervals stand out even more clearly in the anomaly time series (i.e., the departures from the climatological-mean annual cycle) shown in Fig. 3.1b. The anomaly time series also exhibits a cooling trend due to stratospheric ozone depletion, changes in water vapor content, and the buildup of greenhouse gases (Shine et al. 2003; Thompson and Solomon 2005). A similar plot appears as Fig. 3.17 of Trenberth et al. (2007). Apart from these features, the nonseasonal variability of global-mean $T$ is very small.

In the analysis that follows, we will make use of the variable $T^*$ defined as the local $T$ minus the global-mean $T$ (i.e., the departures from each month’s global-mean $T$). Using $T^*$ in place of $T$ effectively eliminates the signatures of the volcanic eruptions and the cooling trend of the global stratosphere in the interannual variability while having only a very small influence on the local $T$.

We also make use of global European Centre for Medium-Range Weather Forecasts (ECMWF) Reanalysis (ERA-40: Uppala et al. 2005) and National Centers for Environmental Prediction (NCEP) – National Center for Atmospheric Research (NCAR) Reanalysis (Kalnay et al. 1996) data for the 23-year period of record 1979-2001 to investigate the relationship between lower-stratospheric temperature and the forcing of the mean flow by wave activity. The variables analyzed were zonally-averaged zonal wind $[u]$, meridional eddy heat flux $[v' T']$, meridional eddy momentum flux $[u' v']$, and the vertical eddy flux of zonal momentum $[u' w']$. The zonal wind and vertical eddy fluxes of zonal momentum are based on monthly mean ERA-40 data, the highest temporal resolution of the ECMWF Reanalysis data that was available to us at the time the study was conducted. In order to include the contribution of the variability with periods shorter than a month, the eddy heat and momentum fluxes were computed based on non-overlapping, 5-day (pentad) mean NCEP-NCAR reanalysis data rather than the monthly
mean ERA-40 data. The meridional eddy heat fluxes in the two datasets were compared and found to be very similar, so the reliability of the data set used to represent them is not an issue. To compute the eddy fluxes, each variable is first decomposed into zonal means and departures from the zonal mean, and the departures of the two variables are multiplied to yield a flux time series. The climatological (23-year) monthly and pentad means are then subtracted from their respective time series to obtain time series of the nonseasonal variability. Finally, the eddy fluxes are zonally averaged, and in the case of \([v'T']\) and \([u'v']\), successive pentads are averaged to create monthly mean data.

### 3.3 Structure and variability of the Brewer-Dobson circulation

#### 3.3.1 Annual mean

The meridional profile of the annual-mean \(T\) is shown in Fig. 3.2, superimposed on a reference line of the annual-mean, global-mean \(T\) (208.6 K). Consistent with prior results of Christy and Drouilhet (1994), \(T\) increases with latitude up to \(\sim 55^\circ\text{N/S}\), beyond which it decreases toward the pole in both hemispheres. Annual-mean \(T\) within the deep tropics (equatorward of \(\sim 15^\circ\text{N/S}\)) are uniform at \(\sim 203.8\) K, approximately 5 K below the global mean. The coolness of the tropical belt relative to subpolar latitudes must be a consequence of the eddy-driven, thermally indirect BDC that drives temperatures away from the radiative equilibrium temperature profile.

In the annual mean, the radiative equilibrium temperature profile in the lower stratosphere is almost perfectly equatorially symmetric. The annual-mean \(T\) mirrors this symmetry, except at latitudes poleward of \(\sim 55^\circ\text{N/S}\). The weaker subpolar-to-polar temperature gradient in the Northern Hemisphere (NH) compared to the Southern Hemisphere (SH) reflects the larger land-sea temperature contrasts and rougher orography of the NH, which give rise to more vigorous wave breaking in the stratosphere, resulting in a stronger poleward mass flux in the BDC with stronger descent and consequently stronger adiabatic warming over the northern polar cap region (Iwasaki 1992; Chae and Sherwood 2007). The more pronounced polar minimum in the SH may also be related to the radiative impact of the loss of stratospheric ozone in recent decades.
3.3.2 The Variability About the Mean

The top panel of Fig. 3.3 shows meridional profiles of $T^*$ for the individual months of January and February (hereafter Jan/Feb) and July and August (hereafter Jul/Aug) during the 29-year period of record. The Jan/Feb and Jul/Aug profiles cross at $\sim 45^\circ$ latitude in both hemispheres. Equatorward of this crossing latitude, temperatures are consistently lower in Jan/Feb than Jul/Aug and their profiles all exhibit the same nearly-parabolic shape that resembles the BDC-induced structure in the annual-mean $T$ (Fig. 3.2). The compactness of the Jan/Feb and Jul/Aug bundles of profiles indicates that the seasonal variations in $T^*$ are much larger than the year-to-year variations, except at high northern latitudes during winter. Removing the equatorial stratospheric quasi-biennial oscillation signal by linear regression (not shown) further reduces the year-to-year variability in $T^*$ within the equatorial belt, but only slightly. Poleward of $\sim 45^\circ$N/S, winters are colder than summers with smaller seasonal contrast in the NH than in the SH. The anomalously high temperatures over the Arctic in Jan/Feb of some years (e.g., 1987) are due to the occurrence of sudden warming events.

We now decompose the meridional temperature profiles into even and odd components with respect to symmetry about the equator. The even (equatorially symmetric) component is defined as the mean of zonally-averaged temperatures at corresponding latitudes in the Northern and Southern Hemispheres, $(T_{\text{NH}} + T_{\text{SH}}) / 2$. The odd (equatorially asymmetric) component is defined as half the difference between zonally-averaged temperatures at corresponding latitudes in the Northern and Southern Hemispheres, which are $(T_{\text{NH}} - T_{\text{SH}}) / 2$ and $(T_{\text{SH}} - T_{\text{NH}}) / 2$, respectively. By construction, the odd component of the meridional profiles is identically equal to zero on the equator at all times. In fact, the odd component of the Jan/Feb and Jul/Aug $T^*$ profiles shown in Fig. 3.3b are remarkably flat out to beyond $20^\circ$N/S; the low-latitude variability in $T^*$ is almost entirely captured by the even component shown in Fig. 3.3c.

Given that the annual cycle in the radiative equilibrium temperature projects almost exclusively onto the odd component, as explained in Section 3.1, it follows that the seasonal variability in the even component of temperature must be almost exclusively
dynamically driven;\(^1\) the difference between the Jan/Feb and Jul/Aug meridional profiles of the even component of \(T^*\) in Fig. 3.3c is a response to differences between the eddy-driven, equator-to-pole BDC in those months. The equator-to-subpolar temperature contrast in the even component of \(T\) and \(T^*\) is almost twice as large in Jan/Feb as in Jul/Aug. The two sets of profiles intersect at \(\sim 45^\circ\text{N/S}\); Jan/Feb is colder (warmer) than Jul/Aug equatorward (poleward) of this crossing latitude. Hence, the annual cycle in \(T^*\) can be characterized as an enhancement of the equator-to-subpolar temperature contrast during the northern winter, which is presumably caused by the strengthening of the BDC. Tropical upwelling in the BDC, as inferred from the concave meridional profiles of the even component of \(T\) in Fig. 3.3c, is stronger during Jan/Feb than in Jul/Aug at latitudes extending from the equator to as far poleward as \(\sim 45^\circ\text{N/S}\).

To place these results concerning the annual cycle in the context of the total variability, Empirical Orthogonal Function (EOF) analysis is performed on the odd and even components of \(T\) (including the global mean), weighing the data in proportion to their contributions to the area-weighted variance (i.e., by the square root of the cosine of latitude). The EOF patterns displayed in the paper are obtained by regressing the odd and even components of the temperature field upon the respective standardized principal component (PC) time series. Thus the amplitudes of the patterns provide an indication of the typical amplitudes of the anomalies observed in association with fluctuations in the respective EOF modes.

The variability of the odd component of \(T\), shown in Fig. 3.4, is almost entirely dominated by the annual cycle of temperatures poleward of \(\sim 45^\circ\text{N/S}\). The hemispheric seesaw in mid-to-high latitude \(T\), following the seasonal migration of the sun, accounts for 97\% of the total month-to-month variance of the odd-component \(T\) field. The second

\(^1\) Nonseasonal variations in \(T^*\) at high latitudes of the Northern and Southern Hemispheres are virtually uncorrelated and hence project equally on the even and odd components of the variability. The even component is emphasized in this study because it captures virtually all of the anti-correlation between low- and high-latitude temperatures that exists by virtue of the equator-to-pole BDC. The nonseasonal variability in the odd component of the high-latitude temperature is largely unrelated to the nonseasonal variability in tropical temperature.
mode, which explains 2% of the total variance, represents the modulation of the annual cycle at midlatitudes. Both modes of the odd component of $T$ exhibit small amplitudes in the tropics (zero on the equator) by construction, and therefore, are not of interest for the purpose of this study. The remaining analyses focus on the variability of the even component.

The leading mode of variability, shown in the top panels of Fig. 3.5, accounts for 89% of the total month-to-month variance of the even-component $T$ field. The corresponding standardized PC 1 time series is dominated by the annual cycle, with extrema in Jan/Feb and Jul/Aug, consistent with the profiles in Fig. 3.3c. During positive excursions of PC 1 observed in Jan/Feb of each year, temperatures equatorward of the crossing latitude at 45°N/S are below the annual mean by ~2°C over most of this latitudinal range. The fluctuations in PC 1 are almost perfectly sinusoidal in shape and quite reproducible from one cycle to the next.

Most of the remaining variance of the even component of $T$ is accounted for by the second EOF shown in the lower panel of Fig. 3.5, which has positive loadings at all latitudes. The corresponding PC time series closely mirrors the behavior of global-mean $T$ anomaly (Fig. 3.1b). The bulge at ~30°N/S resembles the meridional profile of the observed temperature trend from 1979 to 2005 documented by Fu et al. (2006), and may indicate the poleward shift of the subtropical jet streams. When EOF analysis is performed on the $T^*$ field from which the variance associated with fluctuations in global-mean $T$ is removed, a virtually identical leading EOF/PC is obtained that accounts for 93% of the variance (not shown).

In view of the dominance of the leading EOF of the even component of $T$, its PC time series, repeated in Fig. 3.6, is a useful index of the fluctuations in the strength of the equator-to-pole BDC. Also shown in Fig. 3.6 are time series of the even component of tropical (equator to 30°N/S) $T^*$, polar cap (60°N/S to the pole) $T$, and zonal wind in the midlatitude stratosphere (40-70°N/S, 10-30 hPa). The tropical $T^*$ and zonal wind time series are inverted to make them easier to compare with the other series. Correlations between each of these time series are shown in Table 3.1. The variability in all four of the
time series is dominated by the annual cycle, which is quite smooth and regular from year to year, even in the presence of the quasi-biennial oscillation and El Niño Southern Oscillation. The similarity between the polar cap and inverted tropical temperature time series confirms the high degree of cancellation between low- and high-latitude temperature variations noted by Yulaeva et al. (1994). The midlatitude stratospheric zonal winds are strongly correlated with temperature indices of the BDC, as required by thermal wind balance; months of strong equator-to-pole temperature gradient are marked by strong (westerly) vertical shear and vice versa. Upon careful inspection of Fig. 3.6, it is evident that the similarity between the time series extends beyond the annual cycle. Subtle differences in the shapes of the seasonal evolution from one year to the next are also similar in the four time series. These relationships are explored in further detail in Section 3.3.3.

Figure 3.7 shows the same time series plotted as a function of calendar month, with individual years superimposed on the same time axis. This mode of presentation emphasizes the shape of the annual march, and it reveals more clearly the seasonal dependence of the year-to-year variability of the time series. The four time series are remarkably similar, both with respect to the mean and the relative amplitude and seasonality of the interannual variability. Nonseasonal variability, as manifested in the diversity of the annual marches, tends to be much larger during Jan/Feb than Jul/Aug.

3.3.3 Nonseasonal Variability

The top panels of Fig. 3.8 show meridional profiles of monthly $T^*$ anomalies (departures from climatological-mean annual cycle) for Jan/Feb and Jul/Aug of each calendar year in the 29-year period of record. Compared with the monthly mean profiles of $T^*$ (Fig. 3.3a), the amplitudes of the anomaly profiles are smaller by about a factor of two within the tropics. It is evident from an inspection of the even component of the profiles, shown in the bottom panels of Fig. 3.8, that anomalously low tropical temperatures are not always accompanied by anomalous high-latitude warmth (and vice
versa) as in Fig. 3.3c. However, there is still a tendency for anti-correlation between low- and high-latitude temperatures.

Modes that reflect the structure of the anomalies about the seasonally-varying climatological mean are obtained by performing EOF analysis upon the even component of the monthly $T$ anomaly field. The leading EOF of the even component of $T$ anomaly, shown in the top panel of Fig. 3.9, accounts for 55% of the total variance. This mode is dominated by the annual cycle and resembles the profile of EOF 1 of the monthly mean variability (Fig. 3.5), except that the crossing latitude is 47.5°N/S, a few degrees poleward of the crossing latitude in Fig. 3.5. Its counterpart for $T^*$ (not shown) is virtually identical and accounts for 72% of the variance. The second mode shown in the bottom panels of Fig. 3.9 is dominated by variations in global-mean temperature, as reflected in the exclusively positive loadings of EOF 2 as well as in the resemblance between its PC 2 time series and global-mean $T$ anomaly time series (Fig. 3.1b). Monthly anomaly time series of the various indices of the BDC are shown in Fig. 3.10. The correlations between these anomaly time series (Table 3.2) are not as strong as those for the time series that include the annual cycle (Table 3.1), but they are statistically significant at above the 99% confidence level, based on the student $t$ statistic for a one-tailed test with the number of degrees of freedom computed from the autocorrelations of the time series.

The meridional structure of the intraseasonal and interannual variability in the BDC is examined separately by applying a successive centered 5- and 3-month running mean smoothing operators to the even component of the monthly anomaly fields to form the low-pass filtered version, which represents the interannual variability. To exclude the contribution of volcanic eruptions and the long-term cooling trend in lower-stratospheric temperatures to the interannual variability, the analysis is performed on the $T^*$ field. The corresponding high-pass filtered series, obtained by subtracting the low-pass filtered time series from the unfiltered series, represents the intraseasonal variability.

The leading EOFs of the even component of high-pass and low-pass filtered $T^*$, shown in Fig. 3.11, both explain large fractions of the total variance (81% and 70%,
respectively). The high-pass filtered pattern is characterized by remarkably uniform amplitudes equatorward of the crossing latitude at 50°N/S and much larger temperature anomalies of opposing sign poleward of that latitude. In the corresponding low-pass filtered pattern, the crossing latitude is 37.5°N/S and the amplitude equatorward of this latitude is not as uniform as that in the high-pass filtered pattern. These results indicate that the dominant structures of the seasonal and nonseasonal variability in the equator-to-pole BDC are similar, both consisting of “seesaws” with crossing latitudes of 37.5°N/S for the intraseasonal variability, 45°N/S for the annual cycle, and 50°N/S for the interannual variability.

### 3.4 Forcing of the tropical upwelling in the Brewer-Dobson circulation

#### 3.4.1 The role of high-latitude wave forcing

In this section, we examine the direct connection between high-latitude wave forcing and tropical (equator to 30°N/S) lower-stratospheric temperature. As a measure of high-latitude wave forcing, we use an EP flux index based on the zonally-averaged eddy heat flux $v'T'$ averaged over 50-80°N/S (weighted by cosine of latitude) and the 50-, 30-, 20- and 10-hPa levels, which roughly correspond to the region occupied by the zonally-averaged polar night jet. Since eddy heat flux is directly proportional to the vertical component of the EP flux, high-latitude mean $v'T'$ is commonly used to indicate the amount of wave activity entering the extratropical stratosphere. Our wave-forcing index is a linear combination of the current and the previous month’s values, where the relative weights are determined by a least-squares best fit designed to maximize the correlation with the PC 1 time series shown in Figs. 3.5, 3.6, 3.9 and 3.10; the weights for the anomaly time series are 0.31 for the current month and 0.60 for the previous month. The resulting wave-forcing index has a 1-month autocorrelation of 0.35. Similar results are obtained when the analysis is based on a wave-forcing index constructed from the current, −1 and −2 month lagged $v'T'$ values (not shown).

Monthly mean and anomaly time series of the EP flux index are shown in Fig. 3.12a,b together with the corresponding PC 1 time series, repeated from Figs. 3.5, 3.6,
The annual cycle in the EP flux index is nearly in phase with the annual
cycle in PC 1, with strongest forcing during the northern winter. Shown in Tables 3.1 and
3.2 are the correlation coefficients between the EP flux index and various indices of the
strength of the BDC. The correlations between the EP flux index and PC 1 are quite
strong for both the mean and anomaly time series (0.82 and 0.65, respectively), as are the
corresponding correlations between the EP flux index and tropical $T^*$ (–0.83 and –0.65,
respectively). Hence, it is evident that variations in high-latitude wave forcing account for
an appreciable fraction of the temporal variance of the strength of the tropical upwelling
in the BDC.

Figure 3.13 shows monthly anomaly time series of the EP flux index and PC 1
time series based on high-pass and low-pass filtered data, as defined in Section 3.3.3. The
resemblance is striking at both frequencies; correlations are both in the range 0.6 to 0.7,
as shown in Tables 3.3 and 3.4. High-pass and low-pass filtered EP flux indices are both
well correlated with tropical temperature as well as with the other filtered BDC indices.
To show that these high correlations are not artifacts of the filtering process, filtered
versions of the various indices were recomputed based on annualized time series
consisting of seasonal November through March means, which represent the time of year
when the EP flux indices are highest (Fig. 3.7). The correlation between the tropical $T^*$
and the EP flux index time series based on these data is –0.72. To further substantiate this
result, Fig. 3.14 shows meridional profiles of regression coefficients (top panel) and
correlation coefficients (bottom panel) between the even component of $T^*$ and the EP
flux index based on high-pass and low-pass filtered data. The profiles of the regression
coefficients resemble the meridional structure of the leading EOFs (Fig. 3.11). The
correlation coefficients in the tropical belt are comparable for the high-pass and low-pass
filtered data, with values on the order of –0.6. The slight dip in the correlation in the
equatorial belt relative to the subtropics for the low-pass filtered data may be due to the
presence of the quasi-biennial oscillation.

The EP flux index plotted as a function of calendar month in Fig. 3.7 clearly
shows the extended “active season” of the planetary waves, which extends from the
month of the breakdown of the Antarctic polar night jet in November through the northern winter. During this season, the fluxes are large and highly variable from month to month and from year to year.

3.4.2 The role of forcing from below

Modeling studies of Boehm and Lee (2003) and Norton (2006) showed that the upward propagation of equatorial planetary waves (EPW) from the troposphere is capable of influencing the rate of upwelling across the tropical tropopause. An enhancement of the EPW leads to enhanced wave breaking in the lower stratosphere, which induces an easterly acceleration of the mean flow at that level. Enhanced equatorial upwelling in the layer below is required to maintain thermal wind balance. Norton (2006) proposed that the annual cycle in the strength of the BDC is induced by a weakening of the EPW during the northern summer, when the core of deep tropical convection shifts away from the equator due to enhanced heating over the Tibetan plateau. Indeed, the vertical momentum flux in the equatorial belt (10°S-10°N, 100-300 hPa), shown in Fig. 3.7, exhibits a pronounced minimum during Jul/Aug that is no less reproducible from one year to the next than the EP flux indices. On the other hand, it seems unlikely that the seasonality of the EPW could be the primary cause of the pronounced annual cycle or the nonseasonal variability in the BDC; nonseasonal variability in these fluxes is virtually uncorrelated with nonseasonal BDC variability \( r = -0.03 \) as represented by PC 1 of the even component of monthly \( T \) anomaly field (Figs. 3.9 and 3.10). It is conceivable that vertical momentum flux as resolved by daily data could yield a correlation that is higher than that based on monthly data. Even so, the role of the EPW in modulating the strength of the BDC is questionable since the EPW are restricted to much lower latitudes than the BDC-related upwelling, as inferred from the crossing latitudes of the leading EOFs (Figs. 3.5, 3.9, 3.11). In numerical experiments of Norton (2006), the response of the tropical upwelling to fluctuations in the EPW forcing was found to be comparable to the width of the equatorial waveguide itself \( \sim 15°N/S \), whereas the crossing latitude of the \( T \) profile in the annual cycle is \( \sim 45°N/S \).
Rosenlof and Reid (2008) suggested that the downward trends in tropical lower-stratospheric temperature and water vapor content over the past few decades could be occurring in response to rising SSTs in the western Pacific warm pool. In way of evidence, they noted a maximum anti-correlation \( r = -0.44 \) between monthly tropical cold point temperature anomalies averaged over a region slightly to the west of the western Pacific warm pool and SST anomalies averaged over the warm pool region. To determine whether this correlation is indicative of a coherent pattern of SST variations that is linked to variations in the strength of the BDC, we regressed gridded data of the extended reconstruction SST version 3 (ERSST.v3: Smith et al. 2008) upon PC 1 of the low-pass filtered, nonseasonal variability. The regression pattern is shown in Fig. 3.15. While positive regression coefficients prevail over most of the western Pacific warm pool region, the values range only up to 0.15 and the regression pattern lacks spatial coherence. Variability of the BDC (PC 1 time series) and SST anomalies averaged over the warm pool region \( (10^\circ \text{S}-10^\circ \text{N}, 120^\circ \text{E}-180^\circ) \) are weakly positively correlated \( r = 0.17 \) and 0.16 for unfiltered and low-pass filtered time series, respectively. The stronger local anti-correlations between SST and cold point temperature reported by Rosenlof and Reid (2008) may have more to do with the influence of SST anomalies on the equatorial planetary waves than on the zonally-symmetric, equator-to-pole BDC; deep convection, large-scale ascent, elevated geopotential heights in the upper troposphere and depressed cold point temperatures all tend to overlie regions of positive SST anomalies. Lanzante (2009) has argued that the cooling trend in the tropical lower stratosphere documented by Rosenlof and Reid (2008) and Rosenlof and Reid (2009) is an artifact related to a change in the radiosonde instrument type during the last decade or so.

3.4.3 The role of wave forcing in the subtropical lower stratosphere

Rosenlof (1995) concluded that much of the forcing of the climatological-mean annual cycle in tropical lower-stratospheric temperatures is due to wave breaking at latitudes around 30° latitude. Randel et al. (2008) showed evidence supporting this view based on climatological-mean eddy flux statistics. They noted that substantial EP flux
divergences at northern subtropical latitudes extend as high as 70-hPa level during the northern winter, inducing strong tropical upwelling, while wave activity at corresponding southern latitudes during the southern winter does not extend as high. If the subtropical momentum flux convergence by these waves does indeed play an important role in the annual cycle in the BDC, it is reasonable to expect that nonseasonal variability in the EP fluxes associated with these waves might force some of the nonseasonal variability in the BDC as well. To test this hypothesis, we regressed the total (even and odd components combined) field of \(v'T'\) for the previous month upon the monthly anomaly time series of tropical \(T^*\) (Fig. 3.10). Consistent with the results of the previous section, the regression pattern shown in Fig. 3.16a exhibits prominent maxima in the vicinity of the polar night jet, with Northern and Southern Hemispheres contributing in roughly equal measure to the forcing of the tropical upwelling. However, no feature of comparable strength is evident at subtropical latitudes, as would be expected if tropospheric waves penetrating into the lower stratosphere in this region were playing a prominent role in forcing the nonseasonal variability in the BDC.

We further tested the hypothesis by regressing the total eddy momentum field \(u'v'\) upon the monthly anomaly time series of tropical \(T^*\) at a one-month lag. The resulting regression pattern, shown in Fig. 3.16b, resembles the pattern for \(v'T'\) with statistically significant features in the high latitude stratosphere, but not in the subtropical lower stratosphere. Similar results are obtained in the corresponding regression cross-section of the EP flux divergence field, shown in Fig. 3.16c.

The high-latitude features in Fig. 3.16 are symmetric about the equator because the analyses are based on data for all months of the year. Regression cross-sections based on the boreal and austral cold seasons exhibit statistically significant features only in their respective hemispheres (not shown). Furthermore, the correlations between the tropical temperature time series and the high-latitude forcings are enhanced when data are stratified by season; for example, the maximum correlation of the high-latitude \(v'T'\) increases from 0.43 for all months of the year (Fig. 4.16a) to 0.56 for December through April (Dec-Apr) and 0.59 for June through October (Jun-Oct).
The \( [u'v'] \) field in the current month (i.e., no lag) regressed upon the monthly anomaly time series of tropical \( T^* \) exhibits a distinct seasonal signature in the midlatitude troposphere. A prominent maximum centered at 60°N and 300 hPa is observed in the regression cross-section of \( [u'v'] \) based on Dec-Apr data, as shown in Fig. 3.17. A SH counterpart is not observed during Jun-Oct (not shown). This result may be interpreted as an indirect effect of the stratospheric sudden warming events upon the tropospheric circulation, via the changes in the refraction of waves dispersing upward from the troposphere. Anomalies in the strength of the NH wintertime stratospheric polar vortex tend to be followed in a few weeks by anomalies of the same sign in the troposphere (Baldwin and Dunkerton 1999; Thompson et al. 2002) and persist for about a month for the duration of the stratospheric anomalies (Baldwin and Dunkerton 1999). The anomalous poleward flux of westerly momentum in the midlatitude upper troposphere associated with anomalous warming of tropical lower-stratospheric temperatures (i.e., weak tropical upwelling) in Fig. 3.17 is consistent with the findings of Hartmann et al. (2000) and Shindell et al. (2001) that anomalous westerly winds in the extratropical stratosphere (i.e., weak forcing of the BDC) tend to favor the equatorward propagation of waves in the upper troposphere and lower stratosphere.

### 3.5 Model results

#### 3.5.1 Introduction

Holton et al. (1995) performed an idealized numerical experiment using a quasigeostrophic model with spherical geometry to examine the circulation that develops in response to a westward force applied in the extratropical stratosphere (their Fig. 4). Under steady-state conditions (i.e., \( \partial [u]/\partial t = 0 \) in Eqn. 1.2), the non-local control of the mass flux is purely downward and the induced circulation is meridionally confined within the latitudinal boundaries of the forcing region. In a transient state in which the forcing is modulated at a given frequency, the induced circulation encompasses a region wider than the forcing. Their results also seem to suggest that a circulation induced by a westward
force with an annual frequency extends farther into and across the tropics, than a higher frequency forcing.

We used the Held and Suarez (1994) dry dynamical core model to test whether a prescribed forcing in the high-latitude stratosphere can induce an appreciable response at low latitudes that can be understood in terms of the downward control principle. Our approach is similar to that of Holton et al. (1995). An easterly forcing, as described in Appendix A.1.2, is applied over a region in the model stratosphere to represent the effect of wave breaking on the zonal wind field. We then observe the response of the MMC to the prescribed forcing. The zonal momentum balance is given by the following

\[
\frac{\partial [u]}{\partial t} - f [v^T] = \nabla \cdot \vec{F} + \alpha
\]

(3.1)

where notation is as defined in Section 1.1.1 and \(\alpha\) represents all residual terms. Our objective is to determine the extent to which a high-latitude forcing \(\partial [u] / \partial t\) can induce a MMC \(f [v^T]\) with a structure similar to the observed BDC. All results are presented as differences between the 3650-day means (after 730-day spin up) of simulations with and without a forcing.

Before we examine the strength and structure of the induced MMC, we first seek to confirm that the prescribed easterly forcing is predominantly balanced by the Coriolis force induced by the mean meridional flow, rather than by the EP fluxes \(\nabla \cdot \vec{F}\) and/or other residual terms \(\alpha\). Using the \(T_{eq}\) profile with highly-resolved isothermal stratosphere (Fig. A.1), we studied the response to a steady easterly torque of amplitude \(0.6 \text{ m s}^{-1} \text{ d}^{-1}\) applied in the region 50-80°N and 0-50 hPa level. Figure 3.18 shows vertical cross-sections of the four terms in Eqn. 3.1, plotted as the pressure-weighted departures from the mean with an identical background state but no forcing. We find that much of the applied easterly torque (Fig. 3.18a) is balanced by the EP flux divergence term (Fig. 3.18d), rather than by the Coriolis force induced by the mean poleward flow (Fig. 3.18b). When the same forcing is applied in a model with a polar vortex in the NH stratosphere using \(T_{eq}\) profile shown in Fig. A.3, the response of the mean meridional flow is greatly enhanced, as shown in Fig. 3.19b. The contribution of the EP flux
divergence term is diminished but not negligible (Fig. 3.19d) when compared to the smaller residual term in simulations with or without a background polar vortex (Figs. 3.18c and 3.19c). In both of these simulations, strong baroclinic eddies are evidently interfering with the downward control mechanism.

Since we are interested in examining the downward control without the effects of baroclinic eddies, we removed the eddies entirely by running an axisymmetric version of the model. We note that the mean states of the three-dimensional (3D) and axisymmetric models differ substantially (panels a and c of Figs. A.5 and A.6). However, this should not be a concern for the purpose of our study, provided that we examine the difference between the perturbed and non-perturbed fields within the same models. Vertical cross-sections of the three terms in Eqn. 3.1 for an axisymmetric simulation with the same forcing as in the 3D simulations and no polar vortex are shown in Fig. 3.20. Except at the poleward edge of the forcing region near the model boundary where patches of large residual forces are observed (Fig. 3.20c), the prescribed easterly forcing is almost entirely balanced by the mean poleward flow (Fig. 3.20b). The streamfunction field for the induced MMC, shown in Fig. 3.21a, is reminiscent of the circulation response to a steady-state westward force documented in Holton et al. (1995; their Fig. 4c). Consistent with the downward control principle, the induced MMC has a meridional width equal to that of the forcing (50-80°N) and extends downwards from the level of the forcing (50 hPa). The tropics is dominated by pancake-like structures that are presumably caused by instabilities. These features can be problematic for our investigation of the tropical upwelling as they may artificially influence the structure and amplitude of the induced MMC that encompasses the tropics. As a way of suppressing them, we prescribed a $T_{eq}$ profile with constant temperatures within the tropics, as discussed in Appendix A.1.1 and shown in Fig. A.4. The resulting streamfunction field, shown in Fig. 3.21b, indeed exhibits a much quieter tropics.

We now have a suitable setting in which to examine the response of the tropical upwelling to high-latitude forcing. For the remainder of our modeling study, unless otherwise noted, we used the axisymmetric version of the Held and Suarez (1994) dry
dynamical core model whose temperatures relax to a $T_{eq}$ profile with flattened tropics and highly-resolved isothermal stratosphere (i.e., no polar vortex). The radiative damping time scale is specified as $(60 \text{ d})^{-1}$ in the free troposphere and stratosphere. All other parameters are as described in Appendix A.1.

3.5.2 Steady-state vs. Periodic Forcing

The model results of Holton et al. (1995) and observational findings of Randel et al. (2002a) seem to suggest that the low-latitude response to extratropical wave forcing depends on the frequency of the forcing, where a high-frequency forcing induces a broader MMC with substantially stronger tropical upwelling than a MMC induced by a low-frequency forcing. These results imply that only the high-frequency variability of tropical upwelling can be explained by extratropical forcing. On the other hand, our observational results suggest that tropical upwelling is modulated by high-latitude wave forcing across a wide range of frequencies.

We have already confirmed that a steady-state forcing (i.e., the low-frequency limit) can only produce a localized response in the latitude domain (Fig. 3.21). Conceptually, this means that under steady-state conditions (i.e., $\partial[u]/\partial t = 0$), mean meridional motions ($f[v]$) can exist only where there is nonzero wave forcing ($\nabla \cdot F$); away from the forcing region ($\nabla \cdot F = 0$), there can be no meridional flow ($f[v] = 0$). In the presence of a time-varying forcing that accelerates or decelerates the mean flow (i.e., $\partial[u]/\partial t \neq 0$), as in the case of a propagating wave energy, the balance between the nonzero $\partial[u]/\partial t$ and $f[v]$ can be maintained outside of the forcing region at any given time. This non-local transient response to a time-varying forcing is the key to understanding the tropical response to remote forcing.

In the next section, we will demonstrate that while the transient response of the MMC to a time-varying forcing can indeed be felt away from the forcing region, the time-mean response is equivalent to the response to a steady-state forcing, provided that the time-varying forcing integrated over a full period of the oscillation is equal to the steady-state forcing. Hence, hereafter we investigate whether a time-varying forcing in
the extratropical stratosphere can induce a transient response of the MMC that is similar to the observed structure of the BDC.

3.5.3 RESPONSE TO TIME-VARYING FORCING

To illustrate the MMC response to a time-varying forcing, the model is forced with a 180-day periodic easterly forcing of amplitude 0.3 m s\(^{-1}\) d\(^{-1}\), as formulated in Eqn. A.4, in the region 30-60\(^\circ\)N and 0-50 hPa layer. This value of the peak amplitude of the prescribed forcing is an order of magnitude smaller than that in Plumb and Eluszkiewicz (1999) and deduced by Rosenlof (1995) because the model, under the current setting, cannot sustain the instabilities generated by a forcing greater than 1.5 m s\(^{-1}\) d\(^{-1}\). The conclusions remain valid for a forcing in the range, 0.3 to 1.5 m s\(^{-1}\) d\(^{-1}\) (not shown). Although we are ultimately interested in studying the response to a forcing with a period of a year or longer, 180-day period was chosen to increase the sample size and the statistical significance of the results.

Figure 3.22 shows the centered 5-day mean streamfunction fields during the four phases of the forcing cycle, where phases 1 and 3 are defined as the time of minimum and maximum strength of the induced MMC (i.e., value at 60\(^\circ\)N, 660 hPa), respectively, and phases 2 and 4 are midpoints of phases 1 and 3. When the forcing is near zero, a weak cell appears in the forcing region along with a secondary cell of the opposite polarity on the equatorward side (Fig. 3.22\(a\)). As the forcing amplifies, the low-latitude cell reverses direction (Fig. 3.22\(b\)). During the phase of peak easterly torque (Fig. 3.22\(c\)), the main cell at the forcing latitude appears to merge with the secondary low-latitude cell to form a single cell that is horizontally elongated towards the tropics, in accordance with the downward control principle. However, the cell does not extend into and across the tropics, as implied by the observations.

The smallness of the changes in the magnitudes of the tropical upwelling when forced with an easterly torque at high latitudes is in contradiction to the observational results presented in Section 3.4. The response of the meridional profile of lower-stratospheric (54-96 hPa) temperature at the time of the peak forcing, shown in Fig. 3.23,
tends to be concentrated at the edges of the band of forcing, where the induced vertical motions are strongest. Tropical temperatures are virtually unchanged. This profile bears little resemblance to the observed profile characterized by uniform temperatures equatorward of ~30°N/S (e.g., the annual mean profile of $T$ in Fig. 3.2).

Averaged over a complete cycle of the periodic forcing, the time-mean MMC is strictly confined to the latitudes of the forcing, as shown in Fig. 3.24. In fact, the structure is nearly identical to that of the cell forced by a steady-state forcing (Fig. 3.21b). The corresponding time-mean profile of lower-stratospheric temperatures, shown in Fig. 3.25a, is consistent with this MMC structure. The fact that the profiles at the time of peak forcing (Fig. 3.23) are virtually undistinguishable from those averaged over the cycle of the periodic forcing (Fig. 3.25a), and presumably from those of a steady-state forcing, confirms the weakness of the tropical response to time-varying forcing at extratropical latitudes.

Next we evaluate the sensitivity of the tropical upwelling response to various parameters of the prescribed forcing and the background field. To compare with observations, we examine the time-mean meridional profiles of lower-stratospheric (54-96 hPa) temperature for the various simulations.

3.5.3.1 Sensitivity to the latitude of the forcing

The responses to a forcing at 50-80°N and 30-60°N are compared in Fig. 3.25a. Lower-stratospheric temperature anomalies of the two simulations are of comparable magnitude equatorward of ~20°N and differ only within their respective forcing regions. In both simulations, the tropics appears to be unperturbed by the extratropical forcing. When the equatorward edge of the forcing is shifted from 30°N to 20°N and 10°N (Fig. 3.25b), temperatures over the equator progressively decrease. For the set of parameter values used in these simulations, the forcing must lie within ~10°N of the equator to induce an appreciable response extending across the equator. Tropical lower-stratospheric temperatures are not affected by shifting the poleward edge of the forcing.
3.5.3.2 Sensitivity to the level of the forcing

The downward control principle predicts that a forcing at a low altitude induces a stronger MMC than a forcing at a higher altitude. Figure 3.25c contrasts the meridional profiles of lower-stratospheric temperature for simulations with an easterly forcing prescribed above the 50, 70 and 100-hPa levels. The meridional temperature gradient across the forcing region increases as the bottom level of the forcing region is progressively lowered, indicative of the strengthening of the MMC. The increase is nonlinear since the total zonal force is a function of pressure-weighted zonal force. The structures of the meridional profiles outside of the forcing region are nearly identical. Thus, the level of the forcing appears to influence the strength of the induced MMC but not its meridional structure.

3.5.3.3 Sensitivity to the amplitude of the forcing

We applied an easterly torque of various strengths to examine the sensitivity to the magnitude of the forcing. Figure 3.25d illustrates the response to a doubling in the peak amplitude of the periodic forcing, from 0.3 to 0.6 m s\(^{-1}\) d\(^{-1}\). The cooling and warming that result from increased vertical motions at the equatorward and poleward edges of the forcing region, respectively, exhibit an approximately linear relationship to the change in forcing amplitude. Temperatures outside of the forcing region, however, are largely unaffected.

3.5.3.4 Sensitivity to the presence of the stratospheric polar vortex

The sensitivity results of the foregoing sections are based on simulations in which temperatures are relaxed to a \(T_{eq}\) profile with an isothermal stratosphere (Fig. A.4). To examine the sensitivity of the forced response to the prescribed \(T_{eq}\) profile, we repeated the experiments using a \(T_{eq}\) profile with a polar vortex in the NH stratosphere. The strength of the polar vortex is determined by the prescribed value of the lapse rate \(\gamma\) (Fig. A.3). A stronger polar vortex tends to be associated with a more equatorward latitude of the downwelling branch of the induced MMC, as suggested by the shifts in the
temperature maxima in Fig. 3.25e. In contrast, the upwelling branch in the lower latitudes is virtually unchanged.

3.5.3.5 Sensitivity to the time scale of the radiative damping

It is widely accepted that the long radiative relaxation time in the tropical lower stratosphere is responsible for the distinct structure of the cold point tropopause (e.g., Randel et al. 2002a). We examined the sensitivity of the lower-stratospheric temperature profile to the prescribed radiative relaxation time scale of the stratosphere. While the basic states of the unperturbed simulations with (40 d)$^{-1}$ and (60 d)$^{-1}$ damping time scales in the stratosphere ($\sigma < 0.1$) are different, the relative responses to the same prescribed forcing are nearly identical (not shown).

Bearing in mind that the radiative relaxation time scales in the real stratosphere are not uniform in latitude or height, it is conceivable that this factor could explain some of the difference between the structure of the lower-stratospheric temperature profile of the model and observations. On the other hand, the sensitivity to the value of the radiative damping time scale in the dry model does not appear to be large enough to account for the discrepancy. A longer radiative damping time scale merely increases the time the atmosphere takes to reach a steady state and does not increase the meridional extent of the forced circulation.

3.5.3.6 Sensitivity to the stratospheric static stability

The horizontal-vertical scale constraint of the downward control, as discussed in Section 3.5.2, is also a function of static stability. Hence we examined the tropical response to a forcing under varying degrees of stratospheric static stability, using $T_{eq}$ profiles shown in Fig. A.7. Relative to their respective unperturbed states, a forcing in an atmosphere with greater static stability ($dT/dz = +3$ K km$^{-1}$) causes larger departures in the lower-stratospheric temperature than in an atmosphere with smaller static stability ($dT/dz = -3$ K km$^{-1}$), as shown in Fig. 3.25f. The transient response of the induced MMC
is also slightly wider in a more stable stratosphere (not shown). However, the tropical response is subtle.

3.6 Summary and concluding remarks

In our analyses of the MSU/AMSU data in Sections 3.3 and 3.4 (Ueyama and Wallace 2010), variations in the zonally-averaged, lower-stratospheric temperatures \( T \) were decomposed into even and odd components with respect to their symmetry about the equator. Seasonal variations in radiative equilibrium temperature are dominated by the odd component. We have focused on variations in the even component, whose variability is mainly in response to variations in the eddy-driven Brewer-Dobson circulation (BDC). The distribution of upwelling and downwelling in the BDC is qualitatively inferred from the observed seasonal and nonseasonal variations in the meridional profiles of the even component of temperature. Whereas most previous studies have considered the eddy driving in the Northern and Southern Hemispheres separately, here they are combined in the “even component”.

We find that nearly 90% of the month-to-month variance of the temporal variations about the annual-mean reference profile, with ascent at low latitudes and descent at high latitudes, is accounted for by a single mode of variability dominated by the annual cycle. The equator-to-subpolar temperature contrast roughly doubles from Jul/Aug to Jan/Feb, implying an approximate factor of two doubling of the BDC. The crossing latitude in the annual cycle (i.e., the latitude poleward of which \( T \) is warmer and the descent in the BDC is presumably stronger in Jan/Feb than in Jul/Aug) is located at 45°N/S. The leading mode of the nonseasonal variability in the BDC, which accounts for over half of the total variance about a seasonally-varying reference profile, exhibits a meridional structure similar to that of the climatological-mean annual cycle. The dominance of a planetary-scale structure of the BDC variability, with a single crossing latitude and relatively uniform amplitudes equatorward of that latitude, prevails on intraseasonal, annual, and interannual time scales. The only notable frequency-dependent
feature is the crossing latitude, which is 12.5° farther poleward in the intraseasonal variability than in the interannual variability.

Three different indices of the strength of the BDC have been shown to vary in synchrony with one another in the annual cycle and in the nonseasonal variability across a wide range of frequencies: (i) inverted tropical temperature, (ii) polar cap temperature, and (iii) inverted zonal wind in the midlatitude stratosphere. These time series all exhibit a strong lagged response to month-to-month variations in the intensity of high-latitude wave forcing represented by the EP flux index constructed from a weighted sum of current and 1-month previous eddy heat fluxes in the region 50-80°N/S and 10-50 hPa. Equally strong relationships are observed on intraseasonal and interannual time scales, as evidenced by visual inspection of the time series and supporting calculations of the temporal correlation coefficients. The meridional structures of the leading EOFs of the even component of $T^*$ based on the filtered nonseasonal time series closely mirror the corresponding profiles of $T^*$ regressed on an index of high-latitude eddy forcing; in particular, crossing latitudes for high-pass and low-pass filtered profiles are located at 50°N/S and 37.5°N/S, respectively.

We have shown in Section 3.4.1 that year-to-year variations in tropical upwelling in the BDC are significantly correlated with variations in the high-latitude EP flux index, thereby confirming and extending previous findings of Randel et al. (2002a,b), Salby and Callaghan (2002) and Dhomse et al. (2008). Fu et al. (2010) offer additional evidence, showing that the 30-year trends in tropical $T$ are highly correlated with dynamically-induced trends in high-latitude temperatures on a calendar month-by-calendar month basis. The strong similarity between the meridional structure of the leading EOFs of the even component of $T^*$ (Fig. 3.11) and the pattern derived by regressing the even component of $T^*$ upon the time series of the EP flux index (Fig. 3.14) suggests that high-latitude eddy forcing strongly influences the equator-to-pole BDC on both intraseasonal and interannual time scales. These results support the notion that the pronounced annual cycle in the strength of the BDC is a consequence of the larger land-sea temperature contrasts and rougher orography in the NH compared to the SH, which give rise to
stronger eddy forcing at high latitudes during the northern winter and thereby induce stronger tropical upwelling.

We have also considered the possible roles of equatorial planetary waves and wave forcing in the subtropical lower stratosphere in influencing the variability of the BDC, but found no significant temporal correlations between the eddy forcing and tropical upwelling on either interannual or intraseasonal time scales. The locations of the crossing latitudes identified in this study would be difficult to explain on the basis of tropical or subtropical eddy forcing alone.

Following the conclusions based on the empirical data, we tested the response of the tropical upwelling to a high-latitude forcing using the dry dynamical core model of Held and Suarez (1994). We first clarified in Section 3.6 that although a time-varying forcing can induce a meridionally-elongated MMC cell at a given time, its time-mean response is identical in structure to the MMC induced by a steady forcing (i.e., the case of a pure downward control). Hence the sensitivity of the transient response of lower-stratospheric temperatures to a prescribed easterly torque in the model’s stratosphere was examined (Fig. 3.25). Meridional profiles of vertical velocity are consistent with the corresponding temperature profiles in the model and are shown in Fig. 3.26 for reference. Tropical temperatures appear to be most sensitive to the equatorward latitude of the forcing region and largely insensitive to the poleward latitude, the bottom-most pressure level, and the amplitude of the forcing. To force tropical upwelling with an appreciable magnitude in this model, the easterly torque must be prescribed at latitudes as equatorward as ~10°N/S, contrary to observations. Neither the radiative damping time scale nor the static stability of the stratosphere seem to influence tropical lower-stratospheric temperatures significantly. The presence or absence of the stratospheric polar vortex also has little effect in the tropics. Future work will be needed to reconcile these model results with the observations.
Fig. 3.1: Monthly (a) mean and (b) anomaly (the departure from climatological-mean annual cycle) time series of global-mean $T$. ©American Meteorological Society. Reprinted with permission.

Fig. 3.2: Meridional profile of climatological (1979-2007) annual-mean $T$. The gray line represents climatological annual mean of global-mean $T$ (208.6 K). ©American Meteorological Society. Reprinted with permission.
Fig. 3.3: Meridional profiles of (a) total, (b) odd and (c) even components of monthly mean $T^*$ (* denotes departure from global mean) for the individual months of Jan and Feb (black) and Jul and Aug (gray) during 1979-2007. ©American Meteorological Society. Reprinted with permission.
Fig. 3.4: The two leading EOFs and their corresponding standardized PC time series of monthly mean, odd component of $T$. EOFs 1 and 2 explain 97% and 2% of the total variance, respectively.
Fig. 3.5: As in Fig. 3.4 but for the even component of $T$. EOFs 1 and 2 explain 89% and 5% of the total variance, respectively. ©American Meteorological Society. Reprinted with permission.

Fig. 3.6: Monthly mean time series of the even component of polar cap (60°N/S to pole) $T$, zonal wind (inverted) in the midlatitude stratosphere (40-70°N/S, 10-30 hPa) based on monthly ERA-40 data, and tropical (equator to 30°N/S) $T^*$ (inverted), with standardized PC 1 of monthly mean, even component of $T$ (see Fig. 4). Peak-to-peak amplitudes are indicated on the left of each time series. ©American Meteorological Society. Reprinted with permission.
Fig. 3.7: Monthly mean time series plotted as a function of calendar month (with Jan-Jun repeated): polar cap (60°N/S to pole) $T$; zonal wind (inverted) in the midlatitude stratosphere (40-70°N/S, 10-30 hPa) based on monthly ERA-40 data; standardized PC 1 of monthly mean, even component of $T$ (see Fig. 4); tropical (equator to 30°N/S) $T^*$ (inverted); EP flux index constructed from lag 0 and –1 month eddy heat fluxes (50-80°N/S, 10-50 hPa) based on pentad NCEP-NCAR reanalysis data; and EPW index (vertical momentum flux averaged over 10°S-10°N, 100-300 hPa) based on monthly ERA-40 data. Peak-to-peak amplitudes are indicated on the left of each time series. ©American Meteorological Society. Reprinted with permission.
Fig. 3.8: Meridional profiles of (top) total and (bottom) even component of monthly anomaly $T^*$ for the individual months of (left) Jan/Feb and (right) Jul/Aug during 1979-2007. Profiles with equatorial $T^*$ below (above) 0°C are in black (gray). ©American Meteorological Society. Reprinted with permission.
Fig. 3.9: The two leading EOFs and their corresponding standardized PC time series of monthly anomaly, even component of $T$. EOFs 1 and 2 explain 55% and 25% of the total variance, respectively. ©American Meteorological Society. Reprinted with permission.

Fig. 3.10: Monthly anomaly time series of the even component of polar cap (60°N/S to pole) $T$, zonal wind (inverted) in the midlatitude stratosphere (40-70°N/S, 10-30 hPa) based on monthly ERA-40 data, and tropical (equator to 30°N/S) $T^*$ (inverted), with standardized PC 1 of monthly anomaly, even component of $T$ (see Fig. 8). Peak-to-peak amplitudes are indicated on the left of each time series. ©American Meteorological Society. Reprinted with permission.
Fig. 3.11: The leading EOFs and their corresponding standardized PC time series of monthly anomaly, even component of $T^*$ based on (top) high-pass filtered and (bottom) low-pass filtered data. EOF 1 of high-pass and low-pass filtered data explain 81% and 70% of their respective total variance. ©American Meteorological Society. Reprinted with permission.
Fig. 3.12: Monthly (a) mean and (b) anomaly time series of standardized PC 1 of even component of $T$ (see Figs. 4 and 8) and EP flux index constructed from lag 0 and –1 month eddy heat fluxes (50-80°N/S, 10-50 hPa) based on pentad NCEP/NCAR reanalysis data. Peak-to-peak amplitudes are indicated on the left of each time series. ©American Meteorological Society. Reprinted with permission.

Fig. 3.13: Monthly anomaly time series of standardized PC 1 of even component of $T^*$ and EP flux index constructed from lag 0 and –1 month eddy heat fluxes (50-80°N/S, 10-50 hPa) based on (a) high-pass filtered and (b) low-pass filtered pentad NCEP/NCAR reanalysis data. Peak-to-peak amplitudes are indicated on the left of each time series. ©American Meteorological Society. Reprinted with permission.
Fig. 3.14: Meridional profiles of (top) regression and (bottom) correlation coefficients between monthly anomaly, even component of \( T^* \) and monthly anomaly time series of EP flux index constructed from lag 0 and \(-1\) month eddy heat fluxes (50-80°N/S, 10-50 hPa) based on high-pass (solid) filtered and low-pass (dashed) filtered pentad NCEP-NCAR reanalysis data. ©American Meteorological Society. Reprinted with permission.
Fig. 3.15: Sea surface temperature field regressed upon the standardized PC 1 time series of monthly anomaly, even component of $T^*$ based on low-pass filtered data.
Fig. 3.16: The zonally-averaged (a) eddy heat flux [$v'T'$], (b) eddy momentum flux [$u'v'$], and (c) EP flux divergence fields in the previous month regressed upon the monthly anomaly time series of tropical (30°S-30°N) $T^*$. The contour intervals are 1 m s$^{-1}$ K for [$v'T'$], 1 m$^2$ s$^{-2}$ for [$u'v'$], and 0.2 m s$^{-1}$ d$^{-1}$ for the EP flux divergence. Positive (negative) regression coefficients are indicated by solid (dashed) lines. Correlations $> 0.3$ are shaded in gray.
Fig. 3.17: The zonally-averaged eddy momentum flux \([u'v']\) field in the current month regressed upon the monthly anomaly time series of tropical (30°S-30°N) \(T^*\) based on data from December through April months. The contour interval is 1 m\(^2\) s\(^{-2}\). Positive (negative) regression coefficients are indicated by solid (dashed) lines. Correlations > 0.3 are shaded in gray.
Fig. 3.18: Vertical cross-sections of the four terms in the zonal momentum balance equation (Eqn. 3.1) for a 3D simulation with steady easterly torque of amplitude 0.6 m s$^{-1}$ d$^{-1}$ in the region 50-80°N and 0-50 hPa level, weighted by the pressure interval ($\delta$Pa): (a) prescribed easterly torque $d[u]/dt$, (b) Coriolis force induced by the mean meridional flow $f[v']$, (c) residual $\alpha$, and (d) EP flux divergence. Solid contour represents $1.5 \times 10^2$ m s$^{-1}$ d$^{-1}$ Pa.
Fig. 3.19: As in Fig. 3.18 but using $T_{eq}$ profile with a polar vortex in the Northern Hemisphere stratosphere (Fig. A.3).
Fig. 3.20: As in Fig. 3.18 but based on an axisymmetric model.
Fig. 3.21: Vertical cross-sections of the time-mean streamfunction field for an axisymmetric simulation in Fig. 3.19 using tropospheric $T_{eq}$ profile (a) as in Held and Suarez (1994) (Fig. A.1) and (b) with constant temperatures within the tropics (Fig. A.4). Contour intervals are $1 \times 10^8$ kg s$^{-1}$. Red (blue) contours indicate positive (negative) values.
Fig. 3.22: Vertical cross-sections of the transient (centered 5-day mean) streamfunction field in response to 180-day periodic forcing of $0.3 \text{ m s}^{-1} \text{ d}^{-1}$ in the region $50-80^\circ\text{N}$, 0-50 hPa in the four phases of the forcing cycle: (a) no forcing, (b) quarter of a cycle after $a$, (c) peak forcing, (d) quarter of a cycle after $c$. Contour intervals are $2 \times 10^7 \text{ kg s}^{-1}$. Red (blue) contours indicate positive (negative) values.
Fig. 3.23: Transient (centered 5-day mean) meridional profiles of lower-stratospheric (54-96 hPa) temperature at the time of the peak forcing (180-day periodic forcing of 0.3 m s$^{-1}$ d$^{-1}$) in the specified regions, plotted as departures from the profile of the unperturbed simulation.

Fig. 3.24: Vertical cross-sections of the time-mean streamfunction field in Fig. 3.22. Contour intervals are $5 \times 10^7$ kg s$^{-1}$. Red (blue) contours indicate positive (negative) values.
Fig. 3.25: Time-mean meridional profiles of lower-stratospheric (54-96 hPa) temperature plotted as the departure from the profile without forcing: (a) 180-day periodic forcing of 0.3 m s$^{-1}$ d$^{-1}$ at 50-80°N and 30-60°N, 0-50 hPa with (60 d)$^{-1}$ radiative damping, neutral static stability and no stratospheric polar vortex; (b) 30-60°N forcing in a with varying latitudes of the forcing; (c) 30-60°N forcing in a with varying levels of the forcing; (d) 30-60°N forcing in a with varying amplitudes of the forcing; (e) 30-60°N forcing in a with NH stratospheric polar vortex of varying strengths; (f) 30-60°N forcing in a with varying stratospheric static stability.
Fig. 3.26: As in Fig. 3.25 but for pressure velocity ($\omega$).
Table 3.1: Correlation coefficients (> 99% significance level) between monthly mean time series of the even components of tropical $T^*$ (equator to 30°N/S); polar cap $T$ (60°N/S to pole); midlatitude stratospheric zonal wind based on monthly ERA-40 data (40-70°N/S, 10-30 hPa); standardized PC 1 of monthly mean, even component of $T$ (see Fig. 3.5); and EP flux index constructed from lag 0 and –1 month eddy heat fluxes (50-80°N/S, 10-50 hPa) based on pentad NCEP-NCAR reanalysis data. ©American Meteorological Society. Reprinted with permission.

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Table 3.2: Correlation coefficients (> 99% significance level) between monthly anomaly time series of the even components of tropical $T^*$ (equator to 30°N/S); polar cap $T$ (60°N/S to pole); midlatitude stratospheric zonal wind based on monthly ERA-40 data (40-70°N/S, 10-30 hPa); standardized PC 1 of monthly anomaly, even component of $T$ (see Fig. 3.9); and EP flux index constructed from lag 0 and −1 month eddy heat fluxes (50-80°N/S, 10-50 hPa) based on pentad NCEP-NCAR reanalysis data. ©American Meteorological Society. Reprinted with permission.

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Table 3.3: Same as Table 3.2 but for high-pass filtered data. ©American Meteorological Society. Reprinted with permission.

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Table 3.4: Same as Table 3.2 but for low-pass filtered data. ©American Meteorological Society. Reprinted with permission.

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Chapter 4
Tropical Tropospheric Circulation

4.1 Introduction

Nonseasonal annular variability of the tropical wind field is much less prominent in the troposphere than in the stratosphere, but it is nonetheless appreciable. Time series of zonally-averaged zonal \([u]\) and meridional \([v]\) wind components at the 150-hPa level over the equator are shown in Fig. 4.1. The \([u]\) time series varies on the intraseasonal (~20-80 d) time scale with a root-mean-squared (r.m.s.) amplitude of ~2.5 m s\(^{-1}\). Anderson and Rosen (1983) showed that the \([u]\) variations on this time scale are significantly correlated with the Madden-Julian Oscillation (MJO: Madden and Julian 1971, 1972).

Kang and Lau (1994) investigated the structure of the dominant modes of tropical tropospheric annular variability in the atmospheric angular momentum and MMC streamfunction \([\psi]\) fields by performing Empirical Orthogonal Function (EOF) analysis on 5-day mean data from the European Centre for Medium-Range Weather Forecasts (ECMWF) operational analysis. Maximum Covariance Analysis (MCA) was also applied upon the same two fields, including a version that had been high-pass filtered to eliminate the interannual variability. They found that the two leading modes of zonally-averaged angular momentum and \([\psi]\) fields are coupled with one another and explain much of the variability in globally-integrated atmospheric angular momentum. Similar patterns are recovered in Section 4.3.1 of our analysis.

Weickmann et al. (1997) analyzed variations in \([u]\) and the terms in the vertically-integrated zonal momentum balance in the context of the latitudinally-varying, intraseasonal variations in atmospheric angular momentum using daily data from the National Centers for Environmental Prediction (NCEP) operational analysis. They showed evidence of pronounced poleward propagation of anomalies in angular momentum and the convergence of eddy fluxes of zonal momentum \([u^*v^*]\), which they attributed to the movement of convection anomalies across the Indo-Pacific warm pool.
region associated with the MJO. They inferred that the MMC undergoes cyclic variations in association with \([u]\) and \([u^*v^*]\) perturbations, with new cells forming in the equatorial upper troposphere that subsequently widen and propagate poleward. Feldstein (1998) also noted a tendency for poleward propagation of \([u]\) anomalies at tropical and subtropical latitudes.

Watanabe et al. (2002) studied the dynamics of the “tropical axisymmetric mode (TAM)” based on an analysis of the monthly mean NCEP reanalysis data, numerical experiments with an atmospheric general circulation model (AGCM), and a theoretical analysis using a linear barotropic model. Their TAM is defined on the basis of the leading EOF of the 300-hPa global streamfunction field, which yields a pattern similar to the 200-hPa pattern obtained by Kang and Lau (1994). They showed that this pattern is robust with respect to the removal of variability associated with El Niño Southern Oscillation (ENSO).

On the basis of numerical experiments with an idealized AGCM, Lee et al. (2007) suggested that tropical annular variability on the intraseasonal time scale could arise from wave-mean flow interactions involving waves propagating into the tropics from higher latitudes. The observational study of Thompson and Lorenz (2004) provides a concrete example of how \([u]\) anomalies in the tropical upper troposphere develop on a time scale of a week or two in response to extratropical forcing, in this case, wintertime variations in the extratropical annular modes.

In this chapter, we document in more detail the structure and evolution of nonseasonal tropical annular variability on time scales ranging from days to weeks. Data sources and analysis techniques are described in the next section. Section 4.3 uses the equatorial symmetry properties of the leading EOFs of the \([\psi]\) field for the MMC as a basis for characterizing the tropical nonseasonal annular variability of \([v]\) and \([u]\). Variations in \([u]\) on the intraseasonal time scale are discussed in further detail in Section 4.4. Section 4.5 examines the extent to which the fundamental descriptors of the observed tropical annular variability are represented in four models of varying complexity. A summary of the results and some concluding remarks are presented in Section 4.6.
4.2 Data and methods

The analysis is based on once daily (00Z) ECMWF Interim reanalysis (ERA-Interim: Simmons et al. 2007) data for the 9-year period from 1 Jan 2000 to 31 Dec 2008 provided by the Research Data Archive at the National Center for Atmospheric Research (data set number ds627.0 at http://dss.ucar.edu). The main advances in ERA-Interim over its predecessors include a four-dimensional variational assimilation scheme with 12-hourly cycling, a new variational bias correction of satellite radiance data, and an overall improved numerical weather prediction model (Simmons et al. 2007). We analyzed data gridded at approximately 0.7° latitude by 0.7° longitude from 89.5°N to 89.5°S and 29 pressure levels from 50 to 1000 hPa. Except where specifically indicated, the study focuses on the “nonseasonal” component defined as the departures from the climatological-mean seasonal cycle computed from the climatological daily means (i.e., 1 Jan 2000-08, 2 Jan 2000-08, ... 31 Dec 2000-08). February 29 data of the leap years (2000, 2004 and 2008) have been omitted.

The four primary variables considered in this study are $u$, $v$, $\psi$, and $u*v*$. Zonal averaging is denoted by square brackets [()] and asterisks denote departures from the zonal mean. $\psi$ for the MMC is computed from a vertical integration of $v$, as described in Dima and Wallace (2003). The eddy flux term is computed by averaging the daily products of $u*$ and $v*$ around latitude circles and subtracting the climatological daily means at each latitude and pressure level (i.e., $[u*v*]$ minus the daily climatological-mean value of $[u*v*]$). Time derivative fields $d([()])/dt$ are estimated as centered differences. For example, what we refer to as a “10-day time rate of change” is computed by subtracting the value 5 days prior from the value 5 days following a given day in the time series. Similar results are obtained if the average of the 5 days for the intervals following and preceding the given day are used as a basis for estimating the 10-day rate of change.

In addition to the reanalysis data, we make use of the global 1° x 1° daily merged precipitation rates from Global Precipitation Climatology Project Version 1DD (GPCP V1.1: Huffman et al. 2001) for the same time period. The tropical annular variability was
examined in relation to the MJO by analyzing the pair of Real-time Multivariate MJO daily time series defined in Wheeler and Hendon (2004). The pair of orthogonal indices, hereafter referred to as MJO 1 and 2, is obtained by projecting the observed daily anomaly data onto the EOFs of a multivariate field of near-equatorial, outgoing longwave radiation (OLR) and zonal winds at 850-hPa and 200-hPa levels. These time-varying indices together describe the eastward propagation of MJO-related patterns along the equator in terms of an eight-phase cycle.

EOF and MCA analyses of the nonseasonal, zonally-averaged fields were performed in the conventional manner, weighting the zonal means along various latitude circles in proportion to their contribution to the area-weighted variance (i.e., by the square root of the cosine of latitude) and weighting the zonal means for various levels in proportion to their contribution to the mass-weighted variance (i.e., by the square root of the pressure interval represented by each of the model levels). The EOF patterns are obtained by regressing the fields upon the standardized principal component (PC) time series. Hence, the amplitudes of the patterns represent typical amplitudes of the anomalies associated with fluctuations in the respective EOF modes. Unless otherwise noted, pairs of MCA patterns are obtained by regressing the left field upon the standardized expansion coefficient (EC) time series of the right field, and vice versa. The resulting heterogeneous regression maps represent amplitudes of the anomalies in the left (right) field associated with fluctuations in the ECs of the right (left) field with an amplitude of one standard deviation. Tests for statistical significance of the regression fields are based on the student-t test with the number of temporal degrees of freedom (or “effective sample size”) estimated using Eqn. 31 in Bretherton et al. (1999).

Daily time series are decomposed into high and low frequency components using Ensemble Empirical Mode Decomposition (EEMD: Wu et al. 2007; Wu and Huang 2009), which yields a set of nearly orthogonal “intrinsic mode functions (IMF time series)” representing the variability in different frequency ranges that can be summed to recover the original time series. The product of the mean period of the IMFs, which approximately doubles between adjacent IMFs, and the energy density of the IMFs is a
constant for a white noise time series (Wu and Huang 2004); thus the energy (or amplitude) distribution of white noise follows an approximate halving from one IMF to the next. The variances and covariances of the IMFs provide the same kind of information as conventional power spectra and cross spectra.

4.3 General characteristics

In this section, we consider not only the intraseasonal variations in \([u]\), but also the weaker, higher-frequency variations in \([v]\), as illustrated in the sample time series shown in Fig. 4.1b. We will show that it is useful to characterize the variations in \([u]\) and \([v]\) in terms of their equatorial symmetry; fluctuations in \([u]\) with even symmetry about the equator tend to be accompanied by variations in \([v]\) with odd equatorial symmetry, and vice versa. Most of the kinetic energy of the nonseasonal annular variability in the tropical troposphere resides in the equatorially-even \([u]\) perturbations. The tendency for the equatorially-even component of \([u]\) to prevail over much of the tropical troposphere is illustrated in Fig. 4.2, which shows correlation coefficients between nonseasonal perturbations in \([u]\) at corresponding latitudes of the Northern and Southern Hemispheres (e.g., 10°N and 10°S) at each pressure level. Near the equator at any given level, \([u]\) perturbations of the two hemispheres are strongly positively correlated with one another, approaching unit correlation on the equator by convention. In the upper troposphere, positive correlations extend outward from the equator to nearly 30° latitude, even wider than the correlations associated with QBO-related variations in the lower stratosphere. To place these correlations in context, consider \([u]\) variations in the two hemispheres that are of comparable amplitude. A correlation coefficient of +0.33 would imply an amplitude ratio (even component / odd component) of 1.33 / 0.67 = 2, in which case the even component would account for 80% of the zonal kinetic energy \([u]^2/2\).

4.3.1 Variability of the mean meridional circulations

The leading EOF of the nonseasonal variability in \([\psi]\) in the global domain (i.e., 90°N-90°S) is shown in Fig. 4.3. This mode, which accounts for 23% of the nonseasonal
variance of the global $\psi$ field (29% in the tropical belt, 30°N-30°S), consists of a single cell centered on the equator that encompasses the entire tropics, much like the dominant pattern of seasonal variability of the $[\psi]$ climatology (e.g., Peixoto and Oort 1992; Dima and Wallace 2003). The peak-to-peak amplitude of the nonseasonal fluctuations in this mode, as inferred from the contours in Fig. 4.3, is ~13% as large as the departures of the climatological-mean Jan-Feb or Jul-Aug $[\psi]$ from the annual mean $[\psi]$ in the ERA-Interim. The cross-equatorial $[v]$ at the 150-hPa level observed in association with an amplitude of unit standard deviation of PC 1 time series is ~0.3 m s$^{-1}$.

The second EOF of the nonseasonal $[\psi]$ field, shown in Fig. 4.4, accounts for 12% of the global nonseasonal variance (16% in the tropical belt) and resembles the climatological annual mean $[\psi]$ field, with well-defined equatorially symmetric Hadley- and Ferrel-like cells (Figs. 1.1 and 2.1a). Peak-to-peak amplitudes per unit standard deviation of these cells are ~15% and ~18% as large as the climatological annual mean strengths of the Hadley and Ferrel cells, respectively. At the 150-hPa level, poleward $[v]$ across 15°N/S observed in association with unit standard deviation of PC 2 time series is ~0.1 m s$^{-1}$.

These leading EOFs of $[\psi]$ are similar to those obtained by Kang and Lau (1994; their Fig. 7) based on 5-day mean ECMWF data in the domain 40°N-40°S. We will refer to the $[\psi]$ pattern in EOF 1 as exhibiting odd equatorial symmetry and the pattern in EOF 2 as exhibiting even equatorial symmetry. The variability with even equatorial symmetry tends to be characterized by lower frequencies than the variability with odd equatorial symmetry, as indicated by the redder frequency spectrum of the PC 2 time series compared to that of the PC 1 time series, as shown in Table 4.1.

Next, we examine the patterns of variations in $[u]$, $d[u]/dt$ and $[u* v^*]$ observed in association with the leading modes of nonseasonal variability in $[\psi]$. These patterns are obtained by regressing the respective nonseasonal fields onto the standardized PC time series of the nonseasonal $[\psi]$ field. The polarity of the EOFs of $[\psi]$ is defined such that positive values of PC 1 and 2 time series denote an enhancement of the climatological mean Jan-Feb and annual mean cells, respectively.
The variations in $[u]$ and 2-day time rate of change of $[u]$ observed in association with excursions in PC 1 of $[\psi]$ (Fig. 4.3) exhibit odd equatorial symmetry. The patterns as a whole appear to be spatially coherent, but only the features in the lower branch of the cell meet the criterion for statistical significance at the 99% level. More prominent are features in $[u^*v^*]$ associated with variations in PC 1 of $[\psi]$. Positive excursions in PC 1, indicative of an enhancement of the climatological mean Jan-Feb cell, are accompanied by anomalous southward cross-equatorial $[u^*v^*]$ in the tropical upper troposphere, along with anomalous northward fluxes in the extratropics of both hemispheres. The cross-equatorial $[u^*v^*]$ and $[v]$ of the MMC in the tropical upper troposphere are in opposite directions, as in the seasonally-varying, climatological mean tropical circulation (Dima et al. 2005). The Coriolis force acting on $[v]$ of the MMC and the meridional convergence of $[u^*v^*]$ induce opposing tendencies in the zonal wind field, thus accounting for the smallness of $d[u]/dt$ (and consequently, of $[u]$) at those levels in the regression patterns in Fig. 4.3. In contrast, statistically significant zonal wind tendencies are observed in the lower branch of the cell, where the Coriolis force induced by the MMC is unopposed by the convergence of $[u^*v^*]$; an easterly acceleration is observed at $\sim12^\circ$N where the anomalous mass flux is equatorward, and a westerly acceleration is observed at $\sim12^\circ$S where the mass flux is poleward.

Positive excursions in PC 2 of nonseasonal $[\psi]$ are accompanied by well-defined $[u]$ anomalies that are nearly symmetric about the equator, with easterly anomalies centered at $\sim25^\circ$N/S and weaker westerly anomalies centered at $\sim55^\circ$N/S (Fig. 4.4). Kang and Lau (1994) obtained similar patterns when they performed MCA upon $[u]$ and $[\psi]$ fields. The Coriolis force acting on the poleward $[v]$ in the upper branches of the Hadley cells tends to damp the easterly wind anomalies centered at subtropical latitudes. It follows that these $[u]$ anomalies must be maintained by the eddy fluxes of zonal momentum. Positive excursions in PC 2 are indeed marked by anomalous divergence of eddy momentum fluxes (i.e., eddy forcing of an easterly acceleration) in the subtropical upper troposphere.
The regression pattern for \([u^*v^*]\) in Fig. 4.4 is dominated by a modulation of the climatological mean poleward fluxes of westerly momentum centered at \(\sim 35^\circ\text{N/S}\). Positive anomalies in PC 2 of \([\psi]\), indicative of a strengthening of the Hadley cells, are accompanied by enhanced poleward eddy transports at these latitudes. It should be noted that the strong equatorial symmetry of the extratropical \([u^*v^*]\) pattern is imposed by the analysis scheme; day-to-day variations in nonseasonal times series of \([u^*v^*]\) at 35°N and 35°S are virtually uncorrelated \((r = 0.02)\). A weaker dipole structure straddling the equator, with centers at \(\sim 10^\circ\text{N/S}\) is also evident in Fig. 4.4c. In the Section 4.4, we show that the fluxes in the inner tropics play an important role in forcing zonal wind anomalies on the equator.

Positive excursions in PC 2 of nonseasonal \([\psi]\) are accompanied by westerly accelerations in the tropical upper troposphere extending out to \(\sim 20^\circ\text{N/S}\). The existence of strong, statistically significant patterns in \([u]\) and in the 10-day time rate of change of \([u]\) indicates that the zonal wind anomalies propagate meridionally, with a new center of action of a given polarity appearing first in the equatorial belt and propagating poleward in both hemispheres.

The distinctive and sharply contrasting features of the \([\psi], [u], d[u]/dt \text{ and } [u^*v^*]\) fields associated with the dominant EOFs of nonseasonal variability in \([\psi]\) (Figs. 4.3 and 4.4) provide a basis for categorizing the nonseasonal annular variability of the tropics. Nonseasonal variability in \([\psi]\), and hence \([v]\), is dominated by a pattern of odd equatorial symmetry (Fig. 4.3) which is attended by weak variations in nonseasonal \([u]\). Stronger nonseasonal variability in \([u]\) is observed in association with an equatorially-even pattern of \([\psi]\) variability (Fig. 4.4). The following two sections elaborate further on the characteristics of the nonseasonal \([v]\) and \([u]\) variability.

### 4.3.2 Variability of the Meridional Wind Component

The leading EOFs of the nonseasonal \([v]\) field, shown in Fig. 4.5, exhibit patterns of odd equatorial symmetry with pronounced equatorial maxima, consistent with EOF 1 of the nonseasonal \([\psi]\) field (Fig. 4.3). However, the pancake-like structures in EOFs of
[v] have shorter vertical scales than the vertical scales of [v] associated with EOF 1 of [ψ]. The frequency dependence of the vertical structure of the MMC is explored in Fig. 4.6, in which the [v] and [ψ] fields are regressed upon two sets of IMFs of PC 1 of [v]: the highest-frequency IMF, which accounts for variations with periods of ~3-4 days (Fig. 4.6a), versus the sum of the IMFs from the third onward, which account for variations with periods longer than ~14 days (Fig. 4.6b). It is evident that the MMC associated with the low-frequency variability in [v] tends to be narrower and deeper, in accord with the *downward control principle* (Haynes et al. 1991; Holton et al. 1995), which predicts that the response of the zonally-symmetric circulation to a localized forcing should become more meridionally confined and vertically elongated as the period of the forcing lengthens. A similar distinction is evident in the leading EOFs of high-pass and low-pass filtered [ψ], shown in Fig. 4.7, in which filtered versions of nonseasonal [ψ] were generated by applying a 30-day Lanczos filter to the daily streamfunction data, then subtracting the daily climatology of the filtered time series.

### 4.3.3 Variability of the Zonal Wind Component

It is evident from Fig. 4.1 that nonseasonal variations of [u] in the upper troposphere are much larger than those of [v], even on the equator. The frequency spectrum of [u] is also much redder than that of [v], as shown in Table 4.1. Furthermore, results presented in Figs. 4.2-4.4 suggest that most of the kinetic energy resides in the equatorially-even component of [u]. The dominance of the equatorially-even component is highlighted in Fig. 4.8, which contrasts the equatorially even and odd components of [u] variability at 12° latitude at the 150-hPa level computed as the sum and difference of the time series at 12°N and 12°S, respectively. Nonseasonal fluctuations are more clearly apparent in the time series of the equatorially-even component (Fig. 4.8a). Furthermore, the EEMD spectrum of the equatorially-even component of [u] is slightly redder than the spectrum of the equatorially-odd component, as seen in Table 4.1.

Patterns remarkably similar to those of [u] and d[u]/dt in Fig. 4.4 can be obtained by performing MCA on the global fields of [u] and 10-day time rate of change of [u]. The
two leading MCA modes of \([u]\) and \(\frac{d[u]}{dt}\), which together account for 52% of the squared covariance, are shown in Fig. 4.9. The pair of patterns in quadrature with one another can be recovered from a variety of other methods, as discussed in Section 4.4.1.

4.3.4 SUMMARY AND DISCUSSION

We have documented the general characteristics of the tropical annular variability, with emphasis on the amplitudes, space and time scales, and equatorial symmetry of nonseasonal perturbations in \([\psi]\), \([v]\) and \([u]\), and the role of the eddy flux convergence and Coriolis force in the time-varying, zonally-averaged zonal momentum budget.

We have shown that tropical tropospheric MMC exhibits two distinctive kinds of nonseasonal variability that can be distinguished by their equatorial symmetry:

- cells that straddle the equator (Fig. 4.3), analogous in meridional structure to the seasonally-varying, climatological mean monsoonal circulation, and
- cells that are symmetric about the equator (Fig. 4.4), analogous in meridional structure to the annual mean, climatological mean Hadley circulation.

We refer to the former cells as exhibiting odd equatorial symmetry and to the latter as exhibiting even equatorial symmetry. The variability with even equatorial symmetry tends to be characterized by lower frequencies (Table 4.1) and much stronger associated \([u]\) variations (compare Figs. 4.3 and 4.4).

We have also shown that, for the nonseasonal variability,

- most of the kinetic energy of the tropical annular circulation resides in the zonal wind component (Fig. 4.1),
- most of the zonal kinetic energy per unit mass \([u]^2/2\) throughout most of the tropical upper troposphere resides in the equatorially-even component (Fig. 4.2),
- the frequency spectrum of \([u]\) is much redder than that of \([v]\) (Table 4.1), and
- the frequency spectrum of the equatorially-even component of \([u]\) is much redder than that of the equatorially-odd component (Fig. 4.8; Table 4.1).

We hypothesize that some of these characteristics follow directly from the symmetry properties of the motions: in particular, by the fact that \([v]\) in an equatorially-
even pattern of MMC approaches zero towards the equator, such that, by default, the
equatorially-odd pattern of MMC dominates at low latitudes. This kinematic constraint,
which is reflected in the ordering of the leading EOFs of the nonseasonal $[\psi]$ field, serves
to limit the strength of the Coriolis term induced by equatorially-even MMC cells in the
inner tropics. In the upper branch of equatorially-odd MMC cells, which are not subject
to this constraint, nonseasonal variations in the Coriolis term appear to be effective in
canceling the zonal wind accelerations induced by the convergence of the $[u^*v^*]$, thereby
accounting for the relative steadiness of the equatorially-odd component of $[u]$, apart
from the thermally-induced seasonal variations (Fig. 4.8b). In contrast, the equatorially-
even component of $[u]$ in the inner tropics should respond more or less passively to the
eddy forcing, giving rise to the much higher amplitude of the nonseasonal fluctuations
(Fig. 4.8a) and the greater redness of their frequency spectrum (Table 4.1).

In the lower branch of equatorially-odd MMC cells (Fig. 4.3), where $[v]$ should be
roughly equal and opposite to that in the upper branch and the eddy fluxes can be
assumed to be negligible, the Coriolis term should induce substantial off-equatorial zonal
wind accelerations in the same sense as the eddy forcing in the upper troposphere.
Accelerations in this sense are, in fact, observed at $\sim$12°N/S, but their amplitudes
($\sim$0.1 m s$^{-1}$ d$^{-1}$) are about an order of magnitude smaller than that of the inferred Coriolis
term. To account for this disparity by invoking linear damping of $[u]$ anomalies would
require one to assume a damping time of less than a day, which is substantially shorter
than estimates for the tropical Pacific Ocean (Deser 1993). It is conceivable that pressure
gradients across the Andes and other tropical mountain ranges could be playing a role in
damping the low-level zonal wind response to the anomalous MMC.

The dominant patterns of nonseasonal variability in $[v]$ exhibit patterns of odd
equatorial symmetry, consistent with the leading EOF pattern of $[\psi]$ but with shorter
vertical scales. We have shown that the aspect ratio of tropical MMC is frequency
dependent. As the time scale of the variations lengthens, the cells become vertically
elongated and meridionally confined, in accordance with the downward control principle.
The nonseasonal variability of \([u]\) tends to be dominated by the equatorially-even component, with most of the variance on the intraseasonal (~20-80 d) time scale. It is characterized by alternating episodes of equatorial easterly and westerly anomalies that widen with time and propagate poleward. A new regime of opposing polarity develops over the equator around the time that the previous regime reaches subtropical latitudes. The nonseasonal variations in \([u]\) associated with these patterns are strong enough to reverse the zonal wind direction in the equatorial belt and the sign of the meridional wind shear \(d[u]/dy\) at 10°N/S, as shown in Fig. 4.10. These fluctuations in \([u]\) are associated with variations in the strength of the equatorially-even MMC cells and the equatorward fluxes of westerly momentum in the inner tropics. In the following section, we will consider the extent to which these patterns of \([u]\) and \([\psi]\) vary in response to (i) extratropical forcing and (ii) variations in deep convection over the Indo-Pacific warm pool region.

### 4.4 Intraseasonal variability of the zonal wind \([u]\) component

This section confirms and extends the results of the prior investigations cited in Section 4.3 concerning the role of the nonseasonal variations in \([u*v*]\) and the MMC in forcing poleward propagating bands of \([u]\) anomalies, and clarifies the contributions of extratropical forcing and the MJO to this nonseasonal variability.

Before proceeding to our results, we will briefly review the relevant literature relating to the MJO. Hendon and Salby (1994) described the life cycle of the MJO by constructing a set of lagged regression maps based on a reference time series of the OLR over the equator at 84°E. In order to extract the features relevant to the MJO, wind and temperature fields were spatially smoothed, then temporally windowed and bandpass filtered to isolate zonal wavenumbers 1-3 and eastward periods of 35-95 days. The composites show eastward propagating equatorial Kelvin and Rossby waves, which have come to be regarded as the defining features of the MJO cycle.

Hsu et al. (1990) and Hsu (1996) have argued that the eastward propagation of tropical intraseasonal oscillation is not as pervasive as it appears in the velocity potential
field. In the 1996 study, Hsu performed singular value decomposition (SVD) analysis on the covariance matrix of the 200-hPa streamfunction and OLR anomalies to extract the recurrent coupled patterns of intraseasonal variability during boreal winter. The dominant pattern in the tropics consists of eastward propagating signal over the Indian and western Pacific oceans that is prominent in the SH but weak in the NH. East of the dateline, the pattern is dominated by a standing oscillation in both hemispheres. The phase relationship between the geopotential height, wind and streamfunction fields further revealed that these features are predominantly associated with Rossby wave type circulation in the subtropics, rather than with equatorial Kelvin waves. Hsu (1996) also noted the tendency for the poleward propagation of the zonally-symmetric component of the tropical intraseasonal oscillation. Kayano and Kousky (1998) identified an oscillating mode in the tropics related to the MJO with similar poleward propagating signals, based on an extended EOF analysis of the 20-70 day bandpass filtered, 200-hPa streamfunction field.

4.4.1 Dominant space-time pattern of tropical \([u]\) variability

The dominant patterns of the nonseasonal \([u]\) variability characterized by equatorially-even alternating bands of anomalous easterlies and westerlies can be recovered by regressing the \([u]\) field upon the standardized PC 1 and 2 time series obtained from EOF analysis of the angular momentum field \([(u)\cos\phi]\), where \(\phi\) is latitude) in the 45°N-45°S and 100-1000 hPa domain, as shown in Fig. 4.11. The leading EOF pattern with subtropical extrema is reminiscent of the leading mode of the zonal mean angular momentum anomalies documented by Kang and Lau (1994; their Fig. 6), the tropical axisymmetric mode pattern identified by Watanabe et al. (2002; their Fig. 3a), and the subtropical \([u]\) and \(d[u]/dt\) patterns in Fig. 4.9. The second mode, with an extremum in the equatorial upper troposphere, resembles the lagged tropical response to

\(^2\) EOF analysis performed on the global angular momentum field (i.e., 90°N-90°S, 100-1000 hPa) yields two subtropical patterns as the first and second modes and an equatorial pattern as the third mode (Fig. 4.12).
variations in the extratropical annular modes, as documented in Thompson and Lorenz (2004; their Fig. 1), and the equatorial \([u]\) and \(d[u]/dt\) patterns in Fig. 4.9.

The lag-correlation function between PC 1 and 2 time series, shown in Fig. 4.13a, confirms the tendency for cyclic evolution of the \([u]\) pattern; variations of the subtropical pattern lag variations of the equatorial pattern by about two weeks, consistent with prior evidence of poleward propagation of \([u]\) anomalies presented by Hsu (1996), Weickmann et al. (1997), Feldstein (1998), Kayano and Kousky (1998), and Thompson and Lorenz (2004). The lag-correlation functions between the nonseasonal time series of \([u]\) over the equator and \([u]\) at 25°N and 25°S at the 150-hPa level, shown in Fig. 4.13b, are similar in shape to that of the PCs, further confirming the tendency for poleward propagation, though it should be noted that the correlations are quite modest, in this case. The evolution of the propagating part of the \([u]\) pattern is illustrated in Fig. 4.14, which shows the \([u]\) field regressed upon the linear combinations of PC 1 and 2 time series, where the relative weights are determined by the sines and cosines of the phase angle of the \([u]\) cycle consisting of the two EOF patterns in Fig. 4.11. The equatorial westerly anomaly splits and propagates poleward towards the subtropics of both hemispheres, while a new center of the opposite sign develops in the equatorial upper troposphere.

Paired patterns of \([u]\) with structures somewhat similar to the leading EOFs can be recovered by regressing the \([u]\) and \(d[u]/dt\) fields upon the standardized daily nonseasonal time series of equatorial \([u]\) at the 150-hPa level, shown in Fig. 4.15. They can also be recovered by regressing the same fields upon the time series of equatorial \(d[u]/dt\) at the 150-hPa level, shown in Fig. 4.16. Time derivative fields \(d[(\ )]/dt\) in Section 4.4 are estimated as 10-day centered differences (i.e., the value 5 days following a given day minus the value 5 days prior to a given day). Unless otherwise noted, all subsequent analyses using the time series of \([u]\) or \(d[u]/dt\) will be the time series at 150 hPa, the level exhibiting the largest amplitude in tropical \([u]\) variability. By construction, the \([u]\) pattern in Fig. 4.15 resembles the \(d[u]/dt\) pattern in Fig. 4.16, but the similarity of the reciprocal patterns (with reversed sign) in the two figures confirms the tendency for cyclic behavior, with the patterns in Figs. 4.15 and 4.16 appearing in quadrature with one another.
Regressions of the $[\psi]$ field upon the same two time series are in the sense to cancel the off-equatorial $[u]$ anomalies. Easterly anomalies in the subtropics, for example, tend to be damped by the westerly Coriolis torque associated with the enhanced Hadley cells (Fig. 4.16), but the correlations are weak ($r \sim 0.2$). Equatorial $[u]$ anomalies appear to develop in response to the convergence of $[u^*v^*]$ in the inner tropics (Fig. 4.16), followed by a broadening of the fluxes across the tropics, as the equatorial $[u]$ maximum widens (Fig. 4.15). The patterns in Fig. 4.16 resemble those in Fig. 4.4, which are obtained by regressing the $[u]$ field upon the standardized PC 2 time series of nonseasonal $[\psi]$.

Regressing the various fields upon a reference time series of $[u]$ at subtropical latitudes also yields regression cross-sections of $[u]$ and $d[u]/dt$ that resemble the paired patterns of $[u]$ (e.g., Fig. 4.11). Figures 4.17 and 4.18 show the $[\psi]$, $[u^*v^*]$, $[u]$ and $d[u]/dt$ fields regressed upon the standardized daily nonseasonal time series of $[u]$ and $d[u]/dt$ at $25^\circ$N during the boreal cold season, November through March (Nov-Mar), respectively. To emphasize the variability on the intraseasonal time scale, year-to-year variability is eliminated by subtracting out the Nov-Mar means from each year in all subsequent plots (unless otherwise noted) restricted to this season.

Even though the fields are regressed upon the Northern Hemisphere (NH) time series, the paired patterns of $[u]$ and $d[u]/dt$ in Figs. 4.17 and 4.18 exhibit a tendency for equatorial symmetry; the $[u]$ anomalies at the same latitude in the two hemispheres are positively correlated as far poleward as $\sim30^\circ$N/S. The features in $[\psi]$ and $[u^*v^*]$ are largely confined to the hemisphere of the reference time series; the Southern Hemisphere (SH) features in these fields are not statistically significant.

The $[u]$ anomaly centered at $25^\circ$N is flanked by an anomaly of the opposite sign on its poleward side, centered at $\sim50^\circ$N (Fig. 4.17). This dipole structure in $[u]$ is reminiscent of the signature of the Northern Annular Mode (NAM), but at a slightly lower latitude. Subtropical westerly anomalies tend to be accompanied by suppressed Hadley and Ferrel cells in the NH and equatorward eddy fluxes of westerly momentum in the belt extending from $\sim15^\circ$N to $\sim50^\circ$N. A similarly strong, broad feature in the $[u^*v^*]$
field is observed at the time of maximum $\frac{du}{dt}$ at 25ºN (Fig. 4.18) near the climatological annual mean position of $[u^*v^*]$ (Fig. 1.1), indicating that an enhancement of subtropical zonal winds is accompanied by a weakening of the poleward eddy fluxes of westerly momentum. Variations in low-latitude $[u^*v^*]$ are also observed, albeit with marginal statistical significance (Fig. 4.17), which is consistent with the lagged tropical $[u]$ response to NAM documented by Thompson and Lorenz (2004).

We investigated whether the SH counterparts to Fig. 4.17 are observed with comparable strength. Figure 4.19 shows the respective fields regressed upon the standardized daily nonseasonal time series of $[u]$ at 25ºS during the austral cold season, May through September (May-Sep). Variations in the Hadley and Ferrel cells, extratropical $[u^*v^*]$, and bands of $[u]$ anomalies are reminiscent of their NH counterparts and with the structures associated with the Southern Annular Mode (SAM). However, there is no indication of equatorial eddy fluxes, and the modulation of $\frac{du}{dt}$ over the equator is weaker than in Fig. 4.17. When the various fields are regressed upon the time series of $[u]$ at 25ºS during the boreal cold season, the equatorial features in $[u^*v^*]$, $[u]$ and $\frac{du}{dt}$ are amplified, as shown in Fig. 4.20. It appears that the equatorial eddy fluxes of zonal momentum, which exhibit a strong signature only in Nov-Mar, play a key role in the tropical annular variability.

An alternative way of examining the structures related to the extratropical annular modes is by regressing the various fields upon the standardized EC time series of $\psi$ based on MCA of the nonseasonal $\psi$ and $[u^*v^*]$ fields. The paired patterns of $\psi$ and $[u^*v^*]$ for the leading mode, which accounts for 37.8% of the squared covariance, are shown in Fig. 4.21, and the corresponding patterns for the second mode, which accounts for 26.9% of the squared covariance, are shown in Fig. 4.22. Note that the $\psi$ field in these figures are homogeneous regression maps for the MCA. The cross-sections in Figs. 4.21 and 4.22 are roughly mirror images of one another, with the main centers of action for the leading mode in the SH and the centers for the second mode in the NH. Bands of $[u]$ anomalies of the opposing sign are observed in their respective hemispheres, which resemble the structures of the extratropical annular modes and the $[u]$ and $\frac{du}{dt}$
structures associated with the fluctuations in subtropical $[u]$ (Figs. 4.17 and 4.18). Both annular mode signatures appear to force weak accelerations over the equator, as indicated in the $d[u]/dt$ fields. It is interesting to note that a simple global pattern in which the NAM and SAM appear together, in combination with equatorial zonal wind signatures, reminiscent of those in Figs. 4.15-4.18, can be obtained by repeating the MCA of the $[\psi]$ and $[u*v*]$ fields to which a 10-day low-pass Lanczos filter has been applied to the $[\psi]$ field. The patterns in this leading mode, shown in Fig. 4.23, which accounts for 43.3% of the squared covariance, are very similar to those in Fig. 4.4 based on EOF analysis of the $[\psi]$ field. Perhaps we should note that the high degree of interhemispheric coupling implied by this pattern is illusionary; the NAM and SAM are not really coupled.

4.4.2 VARIABILITY OF $[u]$ OVER THE EQUATOR

The EEMD spectra of the MJO 1 and 2 indices of Wheeler and Hendon (2004) exhibit a peak on the intraseasonal time scale (i.e., in IMFs 3-4), as shown in Table 4.1. The nonseasonal variability in $[u]$ over the equator, whose spectrum is also shown in Table 4.1, is dominated by fluctuations in the similar ~20-50 d period range, which is suggestive of a connection with the MJO. Many of the low-latitude features in the $[\psi]$, $[u*v*]$, $[u]$ and $d[u]/dt$ fields associated with variations in equatorial $[u]$ (Figs. 4.15 and 4.16) are evident in the corresponding patterns for the MJO cycle, shown in Figs. 4.24 and 4.25. These patterns were derived by regressing the various fields upon the linear combinations of MJO 1 and 2, where the relative weights are determined by the sines and cosines of the phase angle of the MJO cycle. Figure 4.24 corresponds to the time of peak equatorial $[u]$, and Fig. 4.25 corresponds to the phase a quarter of a cycle later. These cross-sections bear a strong resemblance to the respective fields regressed upon the MJO 2 and 1 indices, respectively (not shown). In these figures, the dominant space-time pattern of tropical $[u]$ variability documented in Section 4.4.1 appears as coupled patterns of $[u]$ and $d[u]/dt$.

The modulation of tropical $[u]$ by the MJO cycle is summarized in the phase space diagram shown in Fig. 4.26a, which was constructed by regressing the time series
of equatorial $[u]$ and $d[u]/dt$ upon the MJO 1 and 2 indices. Also shown are the corresponding values for the PC 2 time series of $[\psi]$, equatorial convergence of $[u^*v^*]$ at the 150-hPa level computed as the difference between the time series of 150-hPa $[u^*v^*]$ at $10^\circ$N and $10^\circ$S, the global atmospheric angular momentum (GAM) computed in the same manner as Kang and Lau (1994), precipitation ($P$) averaged over the Indo-Pacific warm pool region ($15^\circ$N-15$^\circ$S, 75$^\circ$E-165$^\circ$E), and the 10-day time rate of change of $P$ ($dP/dt$). The MJO-related, tropical annular variability is revealed in greater detail in the time-latitude sections of $[u]$, $[\psi]$ and $[u^*v^*]$ at the 150-hPa level, shown in Fig. 4.27a. These sections were constructed by regressing the respective fields upon the linear combinations of MJO 1 and 2 in accordance with the sines and cosines of the phase angles in Fig. 4.26a.

In the following interpretation of Figs. 4.26a and 4.27a, phases are expressed in terms of a nominal 40-day period of the MJO cycle. The peak in PC 2 of $[\psi]$, which denotes an enhancement of the climatological mean Hadley circulation (Fig. 4.4), occurs around the time of maximum westerly acceleration on the equator and precedes the peak convergence of $[u^*v^*]$ by $\sim$4 days. The peak in equatorial convergence of eddy momentum fluxes is followed by peak equatorial westerlies $\sim$7 days later, followed by a peak in GAM another $\sim$3 days later. The convergence of $[u^*v^*]$ into the equatorial belt at the 150-hPa level occurs presumably due to a strengthening of the equatorial planetary waves around the time of peak MJO 1, which can be deduced from the composite charts of the 300-hPa vertical velocity and 150-hPa geopotential height and wind fields, based on MJO 1 and 2 (Fig. 4.28). The patterns in Fig. 4.28a,c project strongly upon the climatological mean equatorial planetary wave signature (Dima and Wallace 2007).

In Fig. 4.27a, the evolution of $[\psi]$ exhibits substantial equatorial asymmetry with a prevalence of northward propagation of $[\psi]$ anomalies, but these features are not statistically significant at the 99% level. In contrast, the low-latitude features in the $[u]$ and $[u^*v^*]$ fields, which exhibit correlations on the order of 0.3 or higher, are much more equatorially symmetric; both exhibit only small biases toward the SH. It is apparent from Figs. 4.26a and 4.27a that both the MMC and the eddy flux terms in the zonal
momentum balance project onto and presumably contribute to the westerly acceleration in the equatorial belt.

Phase space plot and time-latitude section analogous to those for MJO can be generated using the time series of \([u]\) and \(d[u]/dt\) over the equator. Figure 4.26b shows the phases of the variables plotted in a two-dimensional phase space in which \(d[u]/dt\) is the \(x\) axis and \([u]\) as the \(y\) axis. Apart from a counterclockwise phase shift of \(~20^\circ\), the results are remarkably similar to those based on the MJO indices, both with respect to the relative timing of the peaks (i.e., phase angle) and the correlations with the indices (i.e., amplitude).

The time-latitude sections based on equatorial \([u]\) and \(d[u]/dt\), shown in Fig. 4.27b, also bear striking resemblance to their MJO counterparts. By construction, the \([u]\) anomalies are stronger and more symmetric about the equator. Like their MJO counterparts, the \([\psi]\) anomalies appear to propagate poleward across the equator, but only the anomalies in the NH are statistically significant. The \([u*\nu^*]\) section shows couplets of equatorial eddy fluxes straddling the equator and mirrors its MJO counterpart with regard to the equatorial symmetry.

The resemblance between the equatorial \([u]\) cycle and the MJO cycle suggests that equatorial wave features of the MJO, as those documented by Hendon and Salby (1994), may also be observed in relation to equatorial \([u]\) variability. Many of the zonally-varying features identified by Hendon and Salby (1994) based on filtered and windowed data are observable in Fig. 4.29, which was constructed by regressing the total fields of precipitation, 150-hPa wind and geopotential height (\(\Phi\)) upon the linear combinations of MJO 1 and 2 used to construct Figs. 4.26a and 4.27a. In all four panels, which represent one half of the MJO cycle (the other half represented in reversed sign), active/suppressed convection is centered over the maritime continent and a swath of strong \(u\) anomalies is observed over the equator, flanked by anomalies in the \(\Phi\) field at subtropical latitudes. The evolution of the zonally-varying component of the anomalies in equatorial precipitation, equatorial \(u\), and \(\Phi\) at 26°N/S over the course of the MJO cycle is summarized in the time-longitude sections in Fig. 4.30a. Eastward propagation is
evident in all three variables. Although the MJO-related precipitation anomalies are mainly confined in the eastern hemisphere, the propagating signals of the $u$ and $\Phi$ anomalies can be traced across all longitudes.

Regression maps analogous to those in Fig. 4.29, but based on the time series of $[u]$ and $d[u]/dt$ over the equator, are shown in Fig. 4.31, and the corresponding time-longitude sections are shown in Fig. 4.30b. At the time of maximum westerly acceleration over the equator (Fig. 4.31a), anomalous upper-level winds diverge out of the region of enhanced precipitation over the Indo-Pacific warm pool region. The main convective region shifts eastward into the Pacific sector an eight of a cycle later, and the equatorial westerly anomalies strengthen in a belt extending from the central Pacific to the Indian Ocean. This feature, reminiscent of an eastward propagating Kelvin wave signature, is also evident in the time-longitude section of equatorial $u$ in Fig. 4.30b. The equatorial westerly wind anomalies continue to intensify as they propagate eastward from the Pacific to Atlantic sector and across Africa. These equatorial wind anomalies are flanked by lobes of negative $\Phi$ anomalies centered at ~20°N/S, which widen slightly as they propagate eastward with the equatorial winds (right panel of Fig. 4.30b). This sequence of events can be interpreted as follows: Rossby waves are initiated to the east of precipitation in the equatorial belt, then spread eastward in tandem with the westerly wind anomalies behind the Kelvin wave front. These waves continue to propagate eastward, and the wind anomalies eventually reverse direction as the cycle enters its phase of suppressed convection (Fig. 4.31 with reversed sign).

Whereas the MJO-related waves can be followed as they propagate all the way around the equator, the waves associated with the equatorial $[u]$ variability are clearly apparent only over the regions remote from the maritime continent (compare Figs. 4.30a and b). When the analysis is repeated using data from which the MJO indices have been linearly regressed out (Fig. 4.30c), the zonally-varying fields do not propagate eastward. As in the total $[u]$ field, the anomalies are strongest over the eastern Pacific and Atlantic sectors, far from the convectively active region.
The paired patterns of \([u]\) are robust with respect to the removal of MJO, but the amplitudes within the tropical belt are reduced. For example, the equatorial and subtropical maxima in the MCA patterns of \([u]\) and \(d[u]/dt\) in Fig. 4.9 are less than half as strong when the analysis is repeated based on MJO-residual data, as shown in Fig. 4.32. Tropical features are similarly weakened in the EOF patterns of angular momentum \((45^\circ\text{N}-45^\circ\text{S}, 100-1000 \text{ hPa})\) based on MJO-residual data, shown in Fig. 4.33. However, a comparison of these EOFs to those in Fig. 4.11 for the total field is not as straightforward since the paired patterns of \([u]\) are not represented in the first two modes, but are mixed among the four leading modes in the MJO-residual field.

The EEMD spectrum of the residual time series of equatorial \([u]\) from which the MJO indices have been linearly regressed out, shown in Table 4.1, is similar in shape to the spectrum of total \([u]\) and the variance is lower by only 25%. Regression cross-sections analogous to those in Figs. 4.15 and 4.16, but based on MJO-residual data, are shown in Figs. 4.34 and 4.35, respectively. The patterns are similar to their counterparts based on the total field, with \(\sim 30\%\) reduction in amplitude.

In summary, the MJO clearly plays a role in the intraseasonal variability of \([u]\), and regressing out the MJO indices from the zonally-averaged data substantially reduces the correlations among the various fields. However, it does not eliminate them altogether, nor does it substantially change the phases in a relative sense.

### 4.4.3 Variability of \([u]\) at 25°N

The structure and evolution of the zonal-symmetric and zonally-varying fields associated with the intraseasonal variability in \([u]\) at subtropical latitudes are investigated in this section. Time-latitude sections analogous to those in Fig. 4.27b, but based on the \([u]\) and \(d[u]/dt\) time series at 25°N and 25°S during the boreal cold season Nov-Mar, are shown in Fig. 4.36. The \([u]\) anomalies at 25°N and 25°S are correlated at \(r = 0.36\). At the time of peak \([u]\) anomalies at 25° latitude, anomalies of the opposite sign develop over the equator, which subsequently widen and propagate poleward. The northward propagation of \([\psi]\) anomalies that was observed in association with the MJO and
equatorial $[u]$ variability (Fig. 4.27) is also evident in these sections. In both hemispheres, a pulse of equatorward eddy fluxes of westerly momentum from the extratropics precedes the maximum in subtropical $[u]$. The weaker $[u^*v^*]$ centers in the inner tropics and the opposite hemisphere are not statistically significant.

Figure 4.37 shows the precipitation, 150-hPa level wind and $\Phi$ fields regressed upon the linear combinations of the time series of $[u]$ and $d[u]/dt$ at 25°N during Nov-Mar. These panels are analogous to those of Fig. 4.31, but constructed on the basis of the subtropical zonal wind variability based on Nov-Mar data from which the Nov-Mar means for each year have been subtracted out. At the time of maximum westerly acceleration at 25°N (Fig. 4.37a), weak westerly anomalies are observed over the equator at longitudes except over the western and central Pacific. An eighth of a cycle later, lobes of positive $\Phi$ anomalies develop over the central Pacific at $\sim$15°N/S, separated by a swath of easterly anomalies straddling the equator. The equatorial easterly anomalies that intensify over the central Pacific is clearly depicted in the time-longitude section of equatorial $u$ shown in Fig. 4.38. The westerly anomalies to the west of the maritime continent appear to be associated with equatorial Kelvin waves, with a prominent maximum in $\Phi$ south of India. The subtropical $\Phi$ anomalies along that longitude are strongly negative (also see right panel of Fig. 4.38). The anomalous precipitation pattern is strongest in this phase, with enhanced convection over the SPCZ and suppressed convection over much of the maritime continent, a pattern that projects strongly upon the ENSO signature. The pattern in Fig. 4.37c is even more striking in Fig. 4.39, which is based on daily data with the Nov-Mar means for each year included. The amplitudes of the anomalies are enhanced overall and the ENSO signature is much more clearly apparent.

Subtropical $u$ anomalies of the same sign develop simultaneously across all longitudes, but the pattern is not zonally uniform. The 150-hPa zonal wind fields (i.e., the zonal component of the wind vectors) in Fig. 4.37 are shown in colored shading in Fig. 4.40. The $u$ anomalies are strongest over the jet exit region in the Pacific sector. This region also corresponds to the northern flank of the Rossby wave couplet in Fig. 4.37. A
secondary maximum in $u$ over the Atlantic sector strengthens as the Rossby wave couplet propagates eastward, carrying with it a swath of equatorial easterly wind anomalies. We also note the high degree of equatorial symmetry in the zonal wind field.

When the same fields are regressed upon the time series of $[u]$ and $d[u]/dt$ at 25$^\circ$S during Nov-Mar (Nov-Mar means for each year removed), the anomalies are relatively weak, especially in the NH, as seen in Fig. 4.41. Nonetheless, the patterns as a whole are similar to those based on the time series at 25$^\circ$N (Fig. 4.37). Many of the same wave features, as are the zonal asymmetries and equatorial symmetry in the 150-hPa zonal wind field, shown in Fig. 4.42. These defining features vanish entirely when the regressions are based on the time series of $[u]$ and $d[u]/dt$ at 25$^\circ$S during May-Sep (not shown).

4.4.4 THE ROLE OF THE EDDY MOMENTUM FLUXES

In the phase space plot of Fig. 4.26b, the strongest equatorward $[u*v^*]$ precede the strongest equatorial $[u]$ by ~25$^\circ$, and they coincide with the reversal of the direction of $[\psi]$ in the inner tropics, consistent with the heuristic argument presented by Weickmann et al. (1997: their schematic Fig. 13) that the departure from thermal wind balance due to the eddy-induced westerly acceleration in the equatorial belt should give rise to an equatorward mass flux in the upper troposphere and sinking in the equatorial belt. The enhanced equatorward fluxes also contribute to the easterly accelerations and enhanced poleward flow in subtropical latitudes.

These features are more clearly apparent in the $[\psi]$, $d[\psi]/dt$, $[u]$ and $d[u]/dt$ fields regressed upon the standardized daily nonseasonal time series of the equatorial convergence of $[u*v^*]$ at the 150-hPa level (i.e., 10$^\circ$N minus 10$^\circ$S), shown in Fig. 4.43. At the time of maximum equatorial convergence of $[u*v^*]$, the Hadley circulation is enhanced in the outer tropics, but the $[\psi]$ anomalies are weak in the inner tropics because they are in the process of reversing sign, as reflected in the pattern for $d[\psi]/dt$. Moreover, both the $[u]$ and $d[u]/dt$ panels exhibit westerly anomalies over the equator, indicating that the equatorial $[u]$ anomalies are not only forced, but also reinforced, by equatorial
The convergence of eddy momentum fluxes into the equatorial belt in the upper troposphere, in turn, appears to be modulated by precipitation over the warm pool region, as implied by the phase relationship in Fig. 4.26b. In fact, when the variables are plotted in a phase space plot using the time series of \( P \) and \( dP/dt \) as the axes, shown in Fig. 4.26c, the proximity of the phases of \( [u,v] \), \( d[u]/dt \), \( [\psi] \) and MJO 1 indicates that they peak more or less simultaneously following an episode of enhanced precipitation over the warm pool region.

The EEMD spectrum of the time series of 150-hPa \( [u,v] \) at 10°N/S is much redder than the spectrum of 300-hPa \( [u,v] \) at 45°N/S, as seen in Table 4.1. The total eddy fluxes of westerly momentum at the tropical and extratropical latitudes are decomposed into components contributed by the transient eddies \( [u',v'] \) and by the modulation of stationary waves \( [U^*v^* + u^*V^*] \), where primes denote departures from the time mean and capital letters denote climatological daily means. Their relative contributions to the total flux are shown in Table 4.2 as the ratio of the r.m.s. amplitudes. It can be seen that \( [u,v] \) at 10°N/S, 150 hPa are mainly due to the modulation of the stationary waves, whereas the transient eddies are responsible for most of the \( [u,v] \) at 45°N/S, 300 hPa, thus explaining the different shapes of their spectra and magnitudes of their variance in Table 4.1. The intraseasonal variability of equatorial and subtropical \( u \) time series exhibit similar differences because they are presumably forced by tropical and extratropical fluxes, respectively; the time series of \( u \) at 25°N exhibits a whiter spectrum than that of equatorial \( u \) (Table 4.1). On the other hand, the spectra of \( u \) are much redder than those of the forcing, which suggests that feedback processes reden the response of \( u \) to the variations in \( [u,v] \).

### 4.4.5 Summary and Discussion

The dominant space-time patterns of \( u \) anomalies at tropical latitudes on the intraseasonal (~20-80 d) time scale have been obtained by

- performing MCA upon the \( u \) field paired with the \( d[u]/dt \) field (Fig. 4.9),
- performing EOF analysis on the angular momentum \( ([u]\cos\phi) \) field (Fig. 4.11),
• regressing the \([u]\) and \(d[u]/dt\) fields upon the PC 2 time series of \([\psi]\) (Fig. 4.4),
• regressing the \([u]\) field upon the time series of 150-hPa \([u]\) or \(d[u]/dt\) over the equator (Figs. 4.15 and 4.16) and at 25°N during Nov-Mar (Figs. 4.17 and 4.18), and
• regressing the \([u]\) field upon the EC 1 time series of low-pass filtered \([\psi]\) paired with the unfiltered \([u^*v^*]\) field (Fig. 4.23).

The paired patterns, consisting of a pattern with \([u]\) anomalies centered on or a few degrees south of the equator and the other with \([u]\) anomalies centered in the subtropics (~25°N/S), tend to appear in quadrature with one another in the intraseasonal time scale. The emergence of these paired patterns reflects the tendency for equatorial symmetry of the \([u]\) field and poleward propagation of \([u]\) anomalies. When the modest lag correlation between the equatorial and subtropical \([u]\) anomalies at the 150-hPa level \((r = -0.26)\) is taken into account, time-latitude sections for the equatorial and subtropical annular modes assume the form of a pulsing of \([u]\) in the respective latitude belts, with only a hint of poleward propagation (not shown). The correlation between 150-hPa \([u]\) at 25°N and 25°S is also quite modest \((r = 0.36)\); time-latitude sections of \([u]\) based on subtropical \([u]\) in one hemisphere exhibit relatively weak signatures in the other hemisphere. Intraseasonal variability of \([u]\) occurs in association with variations in zonally-varying \(u\) and \(\Phi\) that appear to be related to equatorial Kelvin and Rossby waves. The evolution of these zonally-varying patterns at the 150-hPa level, as revealed by the regression maps, time-latitude and time-longitude sections, sheds light upon the dynamics of the tropical annular variability.

Consideration of the momentum balance is helpful in diagnosing these variations in \([u]\). The strongest westerly accelerations over the equator are observed at the time when the equatorial westerly anomalies are strengthening as they spread eastward from the central Pacific (Fig. 4.31). At these times, the convergence of \([u^*v^*]\) in the equatorial upper troposphere, as inferred from Fig. 4.16, is anomalously strong. The anomalous couplet in \([u^*v^*]\) at ~12°N/S is mainly due to the superposition of the anomalous equatorial planetary waves and the zonally-varying climatological mean flow, as documented in Table 4.2. The equatorward flux of westerly momentum at 12°N/S is
enhanced when the anomalous flow and precipitation centers come into alignment with the structure of the climatological mean equatorial planetary waves and the associated heat source over the warm pool region (e.g., compare Fig. 4.31a with Fig. 2 of Dima and Wallace 2007).

The anomalies in the MMC are also in the proper sense to support the zonal accelerations over the equator (e.g., Figs. 4.4 and 4.16). The phase relationships shown in Fig. 4.26b indicate that the fluctuations in \([\psi]\) lead the fluctuations in \([u^*v^*]\) by approximately an eighth of a cycle, from which one might infer that it is \([\psi]\), rather than \([u^*v^*]\), that initiate the variations in equatorial \([u]\). However, it seems more likely that the fluxes are the prime mover and that their \(\sim 1/8\) cycle phase lag relative to \([\psi]\) is due to the change in the polarity of the MMC that occurs in response to the fluxes, as described in Fig. 4.43 and in previous studies of Kang and Lau (1994) and Weickmann et al. (1997).

A recent study of K. Grise and D. Thompson (*personal communication*) confirms the role of \([u^*v^*]\) by the equatorial planetary waves in modulating the tropical upper troposphere. The dominant pattern of variability that they define in terms of the eddy kinetic energy (EKE) field computed from daily NCEP-NCAR reanalysis data (1979-2009) is qualitatively similar to the patterns described in this study. For example, vertical cross-sections of the zonally-averaged EKE and zonal wind fields regressed upon the leading PC time series of the EKE show broad equatorial maximum centered at the 150-hPa level, reminiscent of the equatorial \([u]\) pattern of this study. Associated with these patterns is an anomalous eddy momentum flux convergence induced by a couplet of \([u^*v^*]\) centered at \(\sim 12^\circ\)N/S, as in Fig. 4.16. This pattern of variability appears to be modulated by convection over the western tropical Pacific and is shown to be related to ENSO and the MJO. Their findings add further credence to the results of this study.

The annular variability at subtropical latitudes appears to be a direct response to extratropical eddy forcing; a weakening of the poleward \([u^*v^*]\) induces a westerly acceleration at 25\(^\circ\)N. The range of latitudes in which the extratropical eddy forcing is influential extends poleward to \(\sim 50^\circ\)N, where \([u^*v^*]\) varies in association with fluctuations in the annular modes. Hence, our results support the conclusion of Thompson
and Lorenz (2004) that the extratropical annular modes induce annular variability in the tropics. That the fluxes are of roughly comparable strength in Figs. 4.17 and 4.18 indicates that they lead the variations in subtropical \([u]\) by about one-eighth of a cycle. Westerly anomalies at 25°N are attended by a weakening of the Hadley and Ferrel cells, which could also be a response to the weakening of the poleward eddy fluxes at extratropical latitudes.

Westerly anomalies at 25°N are also attended by anomalous poleward \([u^*v^*]\) at ~12°N/S, which serve to reinforce them while inducing easterly anomalies in the equatorial belt (Fig. 4.17). The anomalous flow at this time (Fig. 4.37c) is dominated by a Rossby wave signature over the central Pacific with positive \(\Phi\) anomalies at ~15°N/S and a swath of easterly anomalies over the equator. The anomalous Rossby wave wind signature partially cancels the convergence of \([u^*v^*]\) by the climatological mean stationary waves, resulting in the development of the anomalous easterlies over the equator. The cancellation (i.e., the anomalous poleward flux) is due to the diffluent anomalous flow over the central Pacific, where the climatological mean flow is westerly, and the confluent anomalous flow over the Indian Ocean sector, where the climatological mean flow is easterly. The Rossby wave signature is present only during the boreal cold season, when climatological mean westerlies are present over the Pacific sector. It strongly resembles the ENSO signature in the interannual variability. Once established over the Pacific sector, the equatorial easterlies tend to spread eastward across the Atlantic and Indian Ocean sectors, such that the peak in equatorial \([u]\) occurs a quarter of a cycle after the peak in \([u]\) at 25°N, consistent with the patterns in Figs. 4.17 and 4.18.

Intraseasonal variations in equatorial \([u]\) resemble those observed in association with the MJO. Positive equatorial \([u]\) anomalies develop in response to what appears to be an eastward propagating Kelvin wave front that develops following episodes of enhanced convection over the warm pool region, as documented in previous studies of Hendon and Salby (1994) and references in Zhang (2005). Whereas the Kelvin wave front defined by regressing equatorial \(u\) upon the MJO indices of Wheeler and Hendon (2004) can be traced all the way around the globe (Fig. 4.30a), its counterpart in the time-
longitude section formed by regressing equatorial $u$ upon the equatorial $[u]$ and $d[u]/dt$ time series (Fig. 4.30b) is evident only at longitudes remote from the region of active convection (i.e., from the date line to ~105°E). The $u$ signature is attended by an eastward propagating signature in $\Phi$ centered at ~20°N/S. The subtropical $\Phi$ anomalies appear to be out of phase with equatorial $u$ in Fig. 4.31, indicative of an equatorial Rossby wave couplet that spreads eastward in tandem with the westerly wind anomalies behind the Kelvin wave front. However, in Fig. 4.30, which provides a more reliable indication of the phase, the $\Phi$ anomalies are delayed relative to the zonal wind anomalies on the equator. The zone of westerly anomalies spreads both zonally and meridionally with the passage of time, subsequently weakens, and gives way to easterlies as the convection over the marine continent enters its suppressed phase.

Although the MJO clearly contributes to the tropical annular variability on the intraseasonal time scale, it accounts for only 25% of the variance of equatorial $[u]$ (Table 4.1), and much of the structure inherent in the regression maps, time-latitude and time-longitude sections can be captured in the analysis of residual data from which the MJO indices have been regressed out. The zonally-varying flow fields associated with a pulsing of precipitation over the warm pool region, shown in Fig. 4.44, are also characterized by equatorial Kelvin and Rossby waves propagating eastward all around the globe, and eastward propagation of the rainfall anomalies in the vicinity of the maritime continent.

Proceeding now to a more general discussion of the tropical annular variability, there are two features of these patterns need to be accounted for: (i) the tendency for even equatorial symmetry, with equatorial and subtropical maxima in the leading EOFs, and (ii) the tendency for poleward propagation of the anomalies. We showed in Section 4.3 that the tendency for even equatorial symmetry of the $[u]$ component is a robust feature of the tropical annular variability. We argued that it may be due to the greater effectiveness of the MMC in canceling the zonal accelerations induced by the equatorially-odd component of the eddy fluxes in the inner tropics. Another factor that might be responsible for, or at least contribute to, the symmetry is the prevalence of
equatorially-symmetric Kelvin and Rossby wave signatures that give rise to rectified, zonally-symmetric variations in \[ u \]. The same Rossby wave signatures favor equatorial and subtropical maxima in the leading modes of \[ u \] variability. Advection in the upper (poleward) branch of the climatological mean Hadley cell might play a role in the meridional spreading of the \[ u \] anomalies.

In the dynamical interpretations offered in this section, equatorial planetary waves provide a medium in which quasi-stationary Kelvin and Rossby waves can coexist, giving rise to a rectified, zonally-symmetric variability. The presence of the climatological mean equatorial planetary waves serves to leverage the intraseasonal variations in the momentum fluxes in the ~12°N-12°S latitude belt (i.e., the momentum flux term arising from the superposition of the intraseasonal variability upon the climatological-mean stationary waves gives rise to time-varying momentum fluxes toward and away from the equator that are larger than those resulting from the intraseasonal variations alone) that drive the \[ u \] variability in the equatorial belt. Wave-mean flow interactions of the kind described in this section would not exist on an aqua-planet model, nor should they be prominent during the summer months in the real atmosphere, when the westerly wave guide that supports the existence of Rossby waves over the Pacific sector is absent.

4.5 Model results

In most of the existing literature, tropical intraseasonal variability is portrayed as a zonally-propagating planetary wave phenomenon. The related annular variability considered in this study offers a complementary perspective that raises a number of fundamental questions about the tropical general circulation that could be addressed in numerical experiments with idealized models. For example:

- Why is the equatorially-even component of the nonseasonal \[ u \] variability more prominent than the equatorially-odd component?
- Why is the frequency spectrum of the equatorially-even component of the nonseasonal \[ u \] variability much redder than that of the equatorially-odd component?
• Why is the equatorially-odd component of the nonseasonal $[u]$ variability smaller than the variability associated with the climatological-mean annual cycle?
• Does tropical annular annular variability play a role in determining the structure and time scale of the MJO, or is it just a passive response to the MJO-induced eddy forcing?

While a thorough investigation of the above questions is beyond the scope of this study, we examined the fundamental characteristics of the tropical annular variability in a hierarchy of models to determine which of the observed features are well simulated in all of the models and which of them are not, and from that knowledge, gain some insights on the underlying causes of the variability. We analyzed daily data from the dry dynamical core model (“dry model”), moist dynamical core model (“moist model”), and two versions of the Atmospheric Model 2 (“aqua-planet AM2” and “full AM2”). Detailed descriptions of the models are found in the Appendix. In the analyses of the full AM2 data — the only model run with a seasonal cycle — the climatological-mean daily means were subtracted from the daily data to construct the nonseasonal time series, following the analysis technique outlined in Section 4.2.

4.5.1 CLIMATOLOGICAL MEAN STRUCTURES

Vertical cross-sections of the climatological-mean annual mean and r.m.s. amplitudes of $[u], [v], [\psi]$ and $[u^*v^*]$ of the four models are shown in Figs. 4.45-4.48, which can be compared to the observed fields based on ERA-Interim data (Fig. 1.1). We note that the zonal mean fields of the dry model, moist model and the aqua-planet AM2 model are equatorially symmetric, by construction.

The dry model simulates the extratropical features reasonably well, but the amplitudes in the tropics are generally much too weak (Fig. 4.45). The quiescent tropics that is characteristic of the dry model is evident in, for example, the absence of the equatorial eddy momentum fluxes in Fig. 4.45d. In contrast, the tropics of the moist model is much more active (Fig. 4.46), in better agreement with the observations. The enhancement of tropical activity due to the inclusion of moist processes appears to be
compensated by a weakened extratropical general circulation, as seen in the reductions in the strength of the Ferrel cells and the poleward eddy momentum fluxes at ~30°N/S.

The climatological mean fields of the aqua-planet version of the AM2 model (Fig. 4.47) indicate an overall increase in the amplitudes of the annual mean and temporal standard deviations. A stronger concentration of the rising branches of the Hadley cells in the equatorial belt compared to observations is accompanied by large peaks in the r.m.s. amplitudes of \( v \) and \( \psi \) over the equator. The variability in the \( u \) field is simulated reasonably well in the tropics, but the variability in the extratropics is much too large.

These biases in the aqua-planet AM2 model are largely eliminated in the simulation of the AM2 model with land-sea contrasts, topography and climatologically-varying SSTs (Fig. 4.48). The full AM2 model also exhibits realistic hemispheric asymmetry, for example, in the strength of the subtropical jets and the Hadley and Ferrel cells. Compared to observations, nonseasonal \( u \) variability in the tropics is slightly weaker, whereas the \( v \) variability tends to be stronger. The annual mean \( u^*v^* \) couplet in the equatorial belt is about three times as strong as in the ERA-Interim.

The climatological annual mean precipitation and the 157-hPa wind fields over the tropical domain of the full AM2 model are shown in Fig. 4.49. Heavy precipitation is observed over the ITCZ, SPCZ and the Indo-Pacific warm pool region. Equatorial winds at this level diverge out of the region of deep convection over the western Pacific, and westerlies prevail in subtropical latitudes of both hemispheres. These tropical features are generally in reasonable qualitative agreement with observations.

Since the climatological mean fields of the dynamical core models differ from observations in such a way that can be problematic for studying the tropical annular variability, we proceed with the investigation using only the simulations of the AM2 model. Results from the dry and moist models will be mentioned as necessary.

4.5.2 VARIABILITY OF THE MEAN MERIDIONAL CIRCULATIONS

The leading EOFs of the nonseasonal \( \psi \) field in the aqua-planet and full versions of the AM2 model are shown in Figs. 4.50 and 4.52, respectively. The dominant mode in
both versions is characterized by an equatorially-odd cell centered on the equator, reminiscent of the cell in Fig. 4.3 based on ERA-Interim data. Compared to observations, however, the cell in the full version is weaker, whereas the cell in the aqua-planet version is stronger and considerably narrower. This mode explains 40% and 24% of the total variance in the aqua-planet and full versions of the AM2 model, respectively (recall 23% in ERA-Interim).

The pattern of the second mode of the nonseasonal \( \psi \) field, shown in Figs. 4.51 and 4.53, exhibits equatorially-even, Hadley and Ferrel-like cells, similar to the pattern based on observations (Fig. 4.4). They also resemble the patterns of their respective climatological annual mean \( \psi \) fields (Figs. 4.47c and 4.48c). This mode explains 17% and 11% of the total variance in the aqua-planet and full versions of the AM2 model, respectively (recall 12% in ERA-Interim).

The dynamical core models exhibit qualitatively similar results (not shown). While the strength and meridional structure of the MMC cells in the leading EOFs of \( \psi \) differ amongst the models and from observations, the contrasting equatorial symmetry of EOFs 1 and 2 appears to be a robust characteristic of the nonseasonal variability in \( \psi \).

The \([u^*v^*], [u] \) and \( d[u]/dt \) fields regressed upon the standardized PC 1 and 2 time series of \( \psi \) for the aqua-planet AM2 model are shown in Figs. 4.50 and 4.51, respectively. These figures are analogous to Figs. 4.3 and 4.4 based on the ERA-Interim data. By convention, positive excursions in PC 1 of \( \psi \) are accompanied by anomalous southward cross-equatorial \( u^*v^* \), in the opposite direction of the equatorial \( v \) of the MMC in the tropical upper troposphere. As in observations, the tendency for opposing zonal wind tendencies of \( u^*v^* \) and the Coriolis force induced by \( \psi \) contributes to the smallness of the anomalies in \( u \) and \( d[u]/dt \). Vertical cross-sections of \( u \) and \( d[u]/dt \) regressed upon the standardized PC 2 time series of \( \psi \) (Fig. 4.51) exhibit tropical structures with much weaker amplitudes than in the observations (Fig. 4.4). The associated \( u^*v^* \) couplet in the equatorial belt is also relatively weak and narrow.

Regression cross-sections for the full version of the AM2 model, shown in Figs. 4.52 and 4.53, bear a closer resemblance to the observed patterns. Major improvements
from the results of the aqua-planet version are evident, particularly in the $[u]$ and $d[u]/dt$ fields. For example, statistically significant zonal wind tendencies are observed in the lower branch of the cell in Fig. 4.52, where the Coriolis force induced by the MMC is unopposed by $[u*v*]$. Moreover, positive excursions in PC 2 are accompanied by nearly symmetric, easterly anomalies in the subtropical upper troposphere that are maintained by the anomalous divergence of eddy momentum fluxes over the same region, consistent with observations. A couplet of $[u*v*]$ in the equatorial belt evidently forces westerly accelerations in the tropical upper troposphere. Although the amplitudes of the features are somewhat weaker in the full AM2 model than in ERA-Interim, this model simulates the tropical annular variability reasonably well.

4.5.3 VARIABILITY OF THE MERIDIONAL WIND COMPONENT

The three leading EOFs of the $[v]$ field in the full AM2 model are examined in Fig. 4.54. The first EOF exhibits an equatorially-odd pattern with maximum in the equatorial upper troposphere, consistent with observations (Fig. 4.5). The higher modes show alternating pancake-like structures centered on the equator that resemble the EOF 2 pattern in Fig. 4.5. The leading EOFs of $[v]$ are not well simulated in the other models with respect to the equatorial symmetry and the vertical scales of the equatorial features (not shown).

The frequency dependence of the aspect ratio of the MMC is explored in Fig. 4.55. Consistent with the results based on ERA-Interim (Fig. 4.6), the MMC associated with the low-frequency variability (periods $\geq 14$ days) in $[v]$ is narrower and deeper in the full AM2 model, in accordance to the downward control principle. Both the moist model and the aqua-planet versions of the AM2 model simulate this frequency dependence, whereas the dry model does not (not shown).

4.5.4 VARIABILITY OF THE ZONAL WIND COMPONENT

Figure 4.56 shows vertical cross-sections of $[u]$ regressed upon the leading PC time series of the angular momentum field in the 46°N-46°S and 74-975 hPa domain,
analogous to Fig. 4.11 but based on model data. The leading EOFs in the AM2 models are generally dominated by extratropical variability. In the aqua-planet version (Fig. 4.56a), the equatorially-even subtropical and equatorial extrema do not appear as separate modes as in ERA-Interim, but appear together in the first EOF mode, though the equatorial maximum is substantially weaker than in the observations. The second and third modes are dominated by extratropical variability in the NH and SH, respectively. The coupled patterns of the $[u]$ variability are better simulated in the full AM2 model (Fig. 4.56b), with the subtropical and equatorial patterns appearing in the first and third EOF modes, respectively.

Nonseasonal variability in $[u]$ exhibits a tendency for even equatorial symmetry in the tropical upper troposphere. The profiles of the correlation coefficients between the time series of $[u]$ anomalies of the northern and southern latitudes at each pressure level for the two versions of the AM2 model are shown in Fig. 4.57. Compared to Fig. 4.2 of the observations, the tendency for equatorial symmetry is weak in both versions. The profiles of the two versions are similar in the ~150-600 hPa layer, both showing peak correlations at ~150 hPa. Correlations fall off above this level, more rapidly in the full version than in the aqua-planet version. The land-sea contrasts, topography and the annual cycle of the SSTs in the full AM2 model all introduce odd equatorial symmetry, and consequently, reduce the amplitude of the equatorially-even component of the nonseasonal variability. Thus it is not surprising that the full version of the AM2 model tends to exhibit less equatorial symmetry than the aqua-planet version.

4.6 Summary and concluding remarks

In Section 4.3, the nonseasonal variability of zonally-averaged zonal wind $[u]$ based on 9 years of daily ERA-Interim data (Simmons et al. 2007) was shown to be dominated by the equatorially-even component, with most of the variance on the intraseasonal (~20-80 d) time scale. The intraseasonal variability consists of alternating episodes of equatorial easterly and westerly anomalies at the 150-hPa level that widen with time and propagate poleward. These nonseasonal fluctuations in $[u]$ anomalies
appear to be linked to modulations in the strength of the equatorially-even mean meridional circulations (MMC) in the tropics and the equatorward eddy fluxes of westerly momentum \([u \times v^*]\) at \(\pm 10^\circ\) in the upper troposphere.

In Section 4.4, we examined in further detail the structure, evolution and origin of the tropical annular variability of \([u]\) on the intraseasonal time scale. We showed that the dominant space-time patterns of \([u]\), consisting of an equatorial pattern with an extremum over the equator and a subtropical pattern with an extremum at \(\pm 25^\circ\), can be recovered by various analysis techniques. Variations in \([u \times v^*]\) by the equatorial planetary waves initiate and reinforce equatorial \([u]\) anomalies both directly and indirectly via the forcing of the MMC. Subtropical \([u]\) anomalies develop in response to the modulation of \([u \times v^*]\) by time-varying transient waves at extratropical latitudes, resulting in a much whiter spectrum of \([u]\) at \(25^\circ\) than the spectrum of \([u]\) over the equator.

In Section 4.5, we investigated the robustness of the observed features in a hierarchy of models. The equatorial symmetry properties of the nonseasonal variability are relatively well simulated in the aqua-planet and full versions of the AM2 model; the leading EOFs of the zonally-averaged streamfunction \([\psi]\) and meridional wind \([v]\) fields are dominated by the equatorially-odd component, whereas the leading EOF of \([u]\) and EOF 2 of \([\psi]\) tend to be dominated by the equatorially-even component. The frequency dependence of the aspect ratio of the MMC is also reasonably well simulated in both versions of the AM2 model. Many aspects of the tropical annular variability, however, can only be studied using the most sophisticated GCM. The fact that only the full AM2 model exhibits modestly realistic tropical annular variability (e.g., the paired patterns of \([u]\)) leads us to conclude that equatorial stationary waves, which are not present in the other three models, play an important role in the intraseasonal variability of the tropics. It is of interest to explore the role of these tropical waves in further detail using a more complete set of data of the full AM2 model or other GCMs of similar complexity.

While this study focuses on the variability of the zonally-symmetric component, the zonally-varying features presented in Section 4.4 shed light upon the structures related to the Madden-Julian Oscillation (MJO). Intraseasonal variations in \([u]\) are
associated with variations in zonally-varying zonal wind $u$ and geopotential height $\Phi$ anomalies that appear to be related to equatorial Kelvin and Rossby waves. Equatorial stationary waves provide a medium in which these waves can coexist, giving rise to a rectified $[u]$ variability. The MJO contributes to the variability over the equator, particularly with respect to the eastward propagating waves, but it accounts for only a modest fraction of the variance of $[u]$ over the equator. The MJO-related variability and the variability that occurs in association with fluctuations in equatorial $[u]$ both appear to be linearly related to variations in the strength of the convection over the Indo-Pacific warm pool region (Fig. 4.44), which modulates the strength of the equatorial planetary waves and the associated equatorward eddy flux of westerly momentum.

By construction, the MJO indices of Wheeler and Hendon (2004) tend to emphasize the part of the response to these fluctuations that propagates across the region of minimum OLR over the Indo-Pacific warm pool, whereas the annular variability analyzed in this study evidently places greater emphasis on the far field response. Rather than viewing the MJO and the MJO-residual annular variability as separate phenomena, it may be more appropriate to think of them as different and complementary ways of visualizing the planetary-scale response to variations in convection over the Indo-Pacific warm pool region. Thus, the singular importance of convection over the Indo-Pacific warm pool region in contributing to these statistics transcends the existence of the MJO.
Fig. 4.1: Daily time series of the 150-hPa level, zonally-averaged (a) zonal $[u]$ and (b) meridional $[v]$ wind components over the equator based on ERA-Interim data from 1 Jan 2000 to 31 Dec 2004.

Fig. 4.2: Vertical cross-section of the correlation coefficients between the daily nonseasonal time series of zonally-averaged zonal winds at the same latitude of the Northern and Southern Hemispheres (e.g., $[u]$ at 5°N and 5°S) at each pressure level.
Fig. 4.3: Vertical cross-sections of the zonally-averaged variables, as indicated, regressed upon the standardized PC 1 time series of the nonseasonal, zonally-averaged streamfunction $[\Psi]$ field. $d[u]/dt$ refers to the 2-day time rate of change of $[u]$. Contour intervals are $0.2 \times 10^{10}$ kg s$^{-1}$ for $[\Psi]$, $0.2$ m s$^{-1}$ for $[u]$, $1$ m$^2$ s$^{-2}$ for $[u^*v^*]$, and $0.04$ m s$^{-1}$ d$^{-1}$ for $d[u]/dt$. Dashed contours denote negative values. Shading represents correlations $\geq 0.2$ (significant above the 99% level).

Fig. 4.4: As in Fig. 4.3 but regressed upon the standardized PC 2 time series of the nonseasonal $[\Psi]$ field. $d[u]/dt$ refers to the 10-day time rate of change of $[u]$, with $0.02$ m s$^{-1}$ d$^{-1}$ contour intervals.
Fig. 4.5: Leading EOFs of the nonseasonal, zonally-averaged meridional wind \([v]\) field. EOFs 1 and 2 explain 7% and 5% of the total variance, respectively. Contour intervals are 0.1 m s\(^{-1}\). Dashed contours denote negative values.
Fig. 4.6: Vertical cross-sections of zonally-averaged (top) meridional wind $[v]$ and (bottom) streamfunction $[\Psi]$ fields regressed upon the leading intrinsic mode functions (IMFs) of high to low frequency based on Ensemble Empirical Mode Decomposition of the standardized PC 1 time series of the nonseasonal $[v]$ field: IMFs with a period of (a) $\sim$3-4 days and (b) $\geq$14 days. Contour intervals are 0.1 m s$^{-1}$ for $[v]$, and $0.2 \times 10^{10}$ kg s$^{-1}$ for $[\Psi]$. Dashed contours denote negative values. Shading represents correlations $\geq$ 0.3 (significant above the 99% level).

Fig. 4.7: Vertical cross-sections of the zonally-averaged streamfunction $[\Psi]$ field regressed upon the standardized, 30-day (a) high-pass and (b) low-pass Lanczos filtered PC 1 time series of the nonseasonal $[\Psi]$ field. Contour intervals are $0.2 \times 10^{10}$ kg s$^{-1}$. Dashed contours denote negative values.
Fig. 4.8: Daily time series of the 150-hPa level, zonally-averaged zonal winds at 12° latitudes: (a) equatorially-even component (i.e., 12°N + 12°S) and (b) equatorially-odd component (i.e., 12°N − 12°S).
Fig. 4.9: (top) Vertical cross-sections of the (a) zonally-averaged zonal wind \([u]\) and (b) 10-day time rate of change of \([u]\), \(d[u]/dt\), regressed upon the standardized EC 1 time series of \(d[u]/dt\) and \([u]\), respectively, derived from MCA of the nonseasonal \([u]\) and \(d[u]/dt\) fields. (bottom) As in top panel but for the second MCA mode. MCA 1 and 2 explain 26.1% and 25.8% of the total squared covariance, respectively. Contour intervals are 0.2 m s\(^{-1}\) for \([u]\), and 0.02 m s\(^{-1}\) d\(^{-1}\) for \(d[u]/dt\). Dashed contours denote negative values.

Fig. 4.10: Meridional profiles of the 150-hPa level, zonally-averaged zonal wind \([u]\) of the 50 days with the highest (black) and lowest (gray) equatorial \([u]\) during Dec through Mar from 2000 to 2008.
Fig. 4.11: Vertical cross-sections of the zonally-averaged zonal wind [$u$] regressed upon the standardized PC 1 and 2 time series of the nonseasonal, zonally-averaged angular momentum field ($[u]\cos\phi$, where $\phi$ is latitude; 45°N-45°S, 100-1000 hPa). EOFs 1 and 2 explain 19% and 13% of the total variance, respectively. Contour intervals are 0.5 m s$^{-1}$. Dashed contours denote negative values.

Fig. 4.12: As in Fig. 4.11 but regressed upon the standardized PC 1-3 time series of the nonseasonal, zonally-averaged angular momentum field over the global domain (90°N-90°S, 100-1000 hPa). EOFs 1-3 explain 16%, 13% and 10% of the total variance, respectively.
Fig. 4.13: Lag-correlation functions between (a) PC 1 and 2 time series of the nonseasonal, zonally-averaged angular momentum field (45°N-45°S, 100-1000 hPa; see Fig. 4.11) and (b) nonseasonal time series of zonally-averaged zonal wind $u$ at 25°N (solid) and 25°S (dashed) and $u$ over the equator, at lags ranging from −40 to 40 days. Positive lags indicate PC 2 leading PC 1 and equatorial $u$ leading subtropical $u$. Dotted line represents the 99% significance level.
Fig. 4.14: Vertical cross-sections of the zonally-averaged zonal wind $[u]$ regressed upon the linear combinations of PC 1 and 2 time series of the nonseasonal, zonally-averaged angular momentum field (45°N-45°S, 100-1000 hPa; see Fig. 4.11), with the relative weights determined by the sines and cosines of the phase angle of the cycle: (left) +PC 2 and successive one-eighths of a cycle; (right) +PC 1 and successive one-eighths of a cycle. Contour intervals are 0.5 m s$^{-1}$. Dashed contours denote negative values. Shading represents correlations $\geq 0.5$ (significant above the 99% level).
Fig. 4.15: Vertical cross-sections of the zonally-averaged variables, as indicated, regressed upon the standardized daily nonseasonal time series of 150-hPa $[u]$ over the equator. $d[u]/dt$ refers to the 10-day time rate of change of $[u]$. Contour intervals are $0.2 \times 10^{10}$ kg s$^{-1}$ for $[\psi]$, 0.2 m s$^{-1}$ for $[u]$, 1 m$^2$s$^{-2}$ for $[u*v*]$, and 0.02 m s$^{-1}$ d$^{-1}$ for $d[u]/dt$. Dashed contours denote negative values. Shading represents correlations $\geq 0.2$ (significant above the 99% level).

Fig. 4.16: As in Fig. 4.15 but regressed upon the standardized daily nonseasonal time series of 150-hPa $d[u]/dt$ over the equator.
Fig. 4.17: As in Fig. 4.15 but regressed upon the standardized daily nonseasonal time series of 150-hPa $[u]$ at 25°N during the boreal cold season Nov-Mar, from which the Nov-Mar means for each year have been subtracted out.

Fig. 4.18: As in Fig. 4.15 but regressed upon the standardized daily nonseasonal time series of 150-hPa $d[u]/dt$ at 25°N during the boreal cold season Nov-Mar, from which the Nov-Mar means for each year have been subtracted out.
Fig. 4.19: As in Fig. 4.15 but regressed upon the standardized daily nonseasonal time series of 150-hPa $[u]$ at 25°S during the austral cold season May-Sep.

Fig. 4.20: As in Fig. 4.15 but regressed upon the standardized daily nonseasonal time series of 150-hPa $[u]$ at 25°S during the boreal cold season Nov-Mar, from which the Nov-Mar means for each year have been subtracted out.
Fig. 4.21: Vertical cross-sections of the zonally-averaged variables, as indicated, regressed upon the standardized EC 1 time series of $\psi$, derived from MCA of the nonseasonal $u*v*$ and $\psi$ fields. Contour intervals are $0.2 \times 10^{10}$ kg s$^{-1}$ for $\psi$, 0.2 m s$^{-1}$ for $u$, $1$ m$^2$ s$^{-2}$ for $u*v*$, and 0.02 m s$^{-1}$ d$^{-1}$ for $d[u]/dt$. Dashed contours denote negative values. Shading represents correlations $\geq 0.2$ (significant above the 99% level).

Fig. 4.22: As in Fig. 4.21 but for the second MCA mode.
Fig. 4.23: Vertical cross-sections of the zonally-averaged variables, as indicated, regressed upon the standardized EC 1 time series of $\Psi$, derived from MCA of the nonseasonal, unfiltered $[u^*v^*]$ and 10-day low-pass Lanczos filtered $\Psi$ fields. Contour intervals are $0.2 \times 10^{10}$ kg s$^{-1}$ for $\Psi$, 0.2 m s$^{-1}$ for $[u]$, 1 m$^2$ s$^2$ for $[u^*v^*]$, and 0.02 m s$^{-1}$ d$^{-1}$ for $d[u]/dt$. Dashed contours denote negative values. Shading represents correlations $\geq 0.2$ (significant above the 99% level).
Fig. 4.24: Vertical cross-sections of the zonally-averaged variables, as indicated, regressed upon the linear combinations of MJO 1 and 2, with the relative weights determined by the sines and cosines of the phase angle of the MJO cycle, in phase with maximum 150-hPa \[ u \] over the equator. \[ d[u]/dt \] refers to the 10-day time rate of change of \[ u \]. Contour intervals are 0.2 x 10^{10} \text{kg s}^{-1} for \[ \Psi \], 0.2 \text{m s}^{-1} for \[ u \], 1 \text{m}^2 \text{s}^2 for \[ u^*v^* \], and 0.02 \text{m s}^{-1} \text{d}^{-1} for \[ d[u]/dt \]. Dashed contours denote negative values. Shading represents correlations \( \geq 0.2 \) (significant above the 99% level).

Fig. 4.25: As in Fig. 4.24 but regressed upon the linear combinations of MJO 1 and 2, in quadrature with maximum 150-hPa \[ u \] over the equator.
Fig. 4.26: Regression coefficients between MJO indices (MJO1, MJO2), standardized daily nonseasonal time series of the global atmospheric angular momentum (GAM), zonally-averaged zonal wind at the 150-hPa level over the equator ($u_0$), 10-day time rate of change of $u_0$ ($dudt_0$), equatorial convergence of eddy flux of westerly momentum at the 150-hPa level ($u^*_v^*$), PC 2 time series of streamfunction ($\Psi$), precipitation averaged over the warm pool region ($15^\circ$N-$15^\circ$S, $75^\circ$E-$165^\circ$E; $P$), and 10-day time rate of change of $P$ ($dP/dt$) plotted on the axes of (a) MJO1 and 2, (b) $u_0$ and $dudt_0$, and (c) $P$ and $dP/dt$. Time evolution follows a counter-clockwise direction.
Fig. 4.27: The 150-hPa level, zonally-averaged (left) zonal wind \([u]\), (middle) streamfunction \([\Psi]\), and (right) northward eddy flux of westerly momentum \([u^*v^*]\) fields regressed upon the linear combinations of (a) MJO 1 and 2, and (b) 150-hPa level \([u]\) and \(d[u]/dt\) over the equator, with the relative weights determined by the sines and cosines of the phase angle of the (a) MJO and (b) \([u]\) cycles. Contour intervals are 0.5 m s\(^{-1}\) for \([u]\), 0.1 \(\times\) \(10^{10}\) kg s\(^{-1}\) for \([\Psi]\), and 1 m\(^2\) s\(^{-2}\) for \([u^*v^*]\). Shading represents correlations \(\geq 0.2\) (significant above the 99% level).
Fig. 4.28: Composite maps with respect to the highest and lowest 200 days of MJO 1 and 2. The 150-hPa level geopotential height (contours; m) and wind (arrows; m s⁻¹) fields superimposed on the 300-hPa omega field (color; Pa s⁻¹), as an indicator of the rainfall distribution. Geopotential heights ≥ 14200 m are plotted with 10 m intervals. Wind vectors are plotted only in the 23°N-23°S domain.
Fig. 4.29: Precipitation (color; mm d\(^{-1}\)), 150-hPa level geopotential height (contours; 5 m) and 150-hPa level wind (arrows; m s\(^{-1}\)) fields regressed upon the linear combinations of MJO 1 and 2, with the relative weights determined by the sines and cosines of the phase angle of the MJO cycle: (a) \(-\text{MJO 1}\), (b) halfway between \(-\text{MJO 1}\) and \(-\text{MJO 2}\), (c) \(-\text{MJO 2}\), (d) halfway between \(-\text{MJO 2}\) and \(+\text{MJO 1}\).
Fig. 4.30: The 150-hPa level (left) zonal wind $u$ at 0°N/S and (right) geopotential height $\Phi$ at 26°N/S regressed upon the linear combinations of (a) MJO 1 and 2, (b) 150-hPa level $[u]$ and $d[u]/dt$ over the equator, and (c) the MJO-residual, 150-hPa level $[u]$ and $d[u]/dt$ over the equator, with the relative weights determined by the sines and cosines of the phase angle of the (a) MJO and (b,c) $[u]$ cycles. Contour intervals are 1 m s$^{-1}$ for $u$, and 5 m for $\Phi$. Colored shading represents equatorial precipitation ($\geq |1 \text{ mm d}^{-1}|$) regressed upon the same time series.
Fig. 4.31: Precipitation (color; mm d$^{-1}$), 150-hPa level geopotential height (contours; 5 m) and 150-hPa level wind (arrows; m s$^{-1}$) fields regressed upon the linear combinations of the standardized daily nonseasonal time series of 150-hPa level $u$ and $d[u]/dt$ over the equator, with the relative weights determined by the sines and cosines of the phase angle of the $u$ cycle: (a) $+d[u]/dt$, (b) halfway between $+d[u]/dt$ and $+[u]$, (c) $+[u]$, (d) halfway between $+[u]$ and $-d[u]/dt$. 
Fig. 4.32: (top) Vertical cross-sections of the (a) zonally-averaged zonal wind \([u]\) and (b) 10-day time rate of change of \([u]\), \(d[u]/dt\), regressed upon the standardized EC 1 time series of \(d[u]/dt\) and \([u]\), respectively, derived from MCA of the nonseasonal \([u]\) and \(d[u]/dt\) fields based on MJO-residual data. (bottom) As in top panel but for the second MCA mode. MCA 1 and 2 explain 27.1% and 23.6% of the total squared covariance, respectively. Contour intervals are 0.2 m s\(^{-1}\) for \([u]\) and 0.02 m s\(^{-1}\) d\(^{-1}\) for \(d[u]/dt\). Dashed contours denote negative values.
Fig. 4.33: Vertical cross-sections of the zonally-averaged zonal wind $[u]$ regressed upon the standardized PC 1-4 time series of the nonseasonal, zonally-averaged angular momentum field ($[u]\cos\phi$, where $\phi$ is latitude; 45°N-45°S, 100-1000 hPa) based on MJO-residual data. EOFs 1-4 explain 19%, 13%, 11% and 10% of the total variance, respectively. Contour intervals are $0.5 \text{ m s}^{-1}$. Dashed contours denote negative values.
Fig. 4.34: Vertical cross-sections of the zonally-averaged variables, as indicated, regressed upon the standardized daily nonseasonal time series of 150-hPa $u$ over the equator, based on MJO-residual data. $d[u]/dt$ refers to the 10-day time rate of change of $u$. Contour intervals are $0.2 \times 10^{10}$ kg s$^{-1}$ for $\Psi$, 0.2 m s$^{-1}$ for $u$, 1 m$^2$ s$^2$ for $u \times v^*$, and 0.02 m s$^{-1}$ d$^{-1}$ for $d[u]/dt$. Dashed contours denote negative values. Shading represents correlations $\geq 0.2$ (significant above the 99% level).

Fig. 4.35: As in Fig. 4.34 but regressed upon the standardized daily nonseasonal time series of 150-hPa $d[u]/dt$ over the equator.
Fig. 4.36: The 150-hPa level, zonally-averaged (left) zonal wind $[u]$, (middle) streamfunction $[\Psi]$, and (right) northward eddy flux of westerly momentum $[u \times v^*]$ fields regressed upon the linear combinations of 150-hPa level $[u]$ and $d[u]/dt$ (a) 25°N and (b) 25°S during the boreal cold season Nov-Mar (from which the Nov-Mar means for each year have been subtracted out), with the relative weights determined by the sines and cosines of the phase angle of the $[u]$ cycles. Contour intervals are 0.5 m s$^{-1}$ for $[u]$, $0.1 \times 10^{10}$ kg s$^{-1}$ for $[\Psi]$, and 1 m$^2$ s$^{-2}$ for $[u \times v^*]$. Shading represents correlations $\geq 0.2$ (significant above the 99% level).
Fig. 4.37: Precipitation (color; mm d$^{-1}$), 150-hPa level geopotential height (contours; 5 m) and 150-hPa level wind (arrows; m s$^{-1}$) fields regressed upon the linear combinations of the standardized daily nonseasonal time series of 150-hPa level $u$ and $d[u]/dt$ at 25°N during the boreal cold season Nov-Mar (from which the Nov-Mar means for each year have been subtracted out), with the relative weights determined by the sines and cosines of the phase angle of the $u$ cycle: (a) $+d[u]/dt$, (b) halfway between $+d[u]/dt$ and $+u$, (c) $+u$, (d) halfway between $+u$ and $-d[u]/dt$. 
Fig. 4.38: The 150-hPa level (left) zonal wind $u$ at 0°N/S and (right) geopotential height $\Phi$ at 26°N/S regressed upon the linear combinations of 150-hPa level $[u]$ and $d[u]/dt$ at 25°N during the boreal cold season Nov-Mar (from which the Nov-Mar means for each year have been subtracted out), with the relative weights determined by the sines and cosines of the phase angle of the $[u]$ cycle. Contour intervals are 1 m s$^{-1}$ for $u$, and 5 m for $\Phi$. Colored shading represents equatorial precipitation ($\geq 1$ mm d$^{-1}$) regressed upon the same time series.
Fig. 4.39: As in Fig. 4.37c but based on daily data with Nov-Mar means for each year not subtracted out.
Fig. 4.40: The 150-hPa level zonal wind field regressed upon the linear combinations of the standardized daily nonseasonal time series of 150-hPa level \( u \) and \( \frac{du}{dt} \) at 25°N during the boreal cold season Nov-Mar (from which the Nov-Mar means for each year have been subtracted out), with the relative weights determined by the sines and cosines of the phase angle of the \( u \) cycle: (a) \( +\frac{du}{dt} \), (b) halfway between \(+\frac{du}{dt}\) and \(+u\), (c) \(+u\), (d) halfway between \(+u\) and \(-\frac{du}{dt}\).
Fig. 4.41: As in Fig. 4.37 but the fields are regressed upon the linear combinations of the standardized daily nonseasonal time series of 150-hPa level $u$ and $d[u]/dt$ at 25°S during the boreal cold season Nov-Mar (from which the Nov-Mar means for each year have been subtracted out).
Fig. 4.42: As in Fig. 4.40 but regressed upon the linear combinations of the standardized daily nonseasonal time series of 150-hPa level \( u \) and \( \frac{du}{dt} \) at 25°S during the boreal cold season Nov-Mar (from which the Nov-Mar means for each year have been subtracted out).
Fig. 4.43: Vertical cross-sections of the zonally-averaged variables, as indicated, regressed upon the standardized daily nonseasonal time series of the equatorial convergence (10°S minus 10°N) of the eddy fluxes of westerly momentum \([u^*v^*]\) at the 150-hPa level. \(d[\Psi]/dt\) and \(d[u]/dt\) refer to the 10-day time rate of change of \([\Psi]\) and \([u]\), respectively. Contour intervals are \(0.1 \times 10^{10}\) kg s\(^{-1}\) for \([\Psi]\), 0.2 m s\(^{-1}\) for \([u]\), \(0.01 \times 10^{10}\) kg s\(^{-1}\) for \(d[\Psi]/dt\) and 0.02 m s\(^{-1}\) d\(^{-1}\) for \(d[u]/dt\). Dashed contours denote negative values. Shading represents correlations \(\geq 0.2\) (significant above the 99% level).
Fig. 4.44: Precipitation (color; mm d\(^{-1}\)), 150-hPa level geopotential height (contours; 5 m) and 150-hPa level wind (arrows; m s\(^{-1}\)) fields regressed upon the linear combinations of the standardized daily nonseasonal time series of precipitation (P) averaged over the warm pool region (15°N-15°S, 65°E-175°E) and 10-day time rate of change of P (dP/dt), with the relative weights determined by the sines and cosines of the phase angle of the P cycle: (a) +P, (b) halfway between +P and −dP/dt, (c) −dP/dt, (d) halfway between −dP/dt and −P.
Fig. 4.45: Climatological-mean (a) annual mean and (b) root-mean-squared amplitudes of the zonally-averaged variables, as indicated in the left panels, for the dry model. Contour intervals are 5 and 1 m s$^{-1}$ for $[u]$, 0.5 and 0.1 m s$^{-1}$ for $[v]$, $2 \times 10^{10}$ and $0.5 \times 10^{10}$ kg s$^{-1}$ for $[\psi]$, and 5 and 10 m$^2$ s$^{-2}$ for $[u^*v^*]$. Dashed contours denote negative values. Gray shading in the left panels indicate values $\geq 5$ m s$^{-1}$ for $[u]$, $\geq 0.5$ m s$^{-1}$ for $[v]$, $\geq 3 \times 10^{10}$ kg s$^{-1}$ for $[\psi]$, and $\geq 50$ m$^2$ s$^{-2}$ for $[u^*v^*]$. 
Fig. 4.46: As in Fig. 4.45 but for the moist model.
Fig. 4.47: As in Fig. 4.45 but for the aqua-planet version of the AM2 model.
Fig. 4.48: As in Fig. 4.45 but for the AM2 model with topography and climatologically-varying SSTs (“full AM2” model).
Fig. 4.49: Climatological-mean annual mean fields of precipitation (colored shading) and 157-hPa level winds (vectors, plotted equatorward of 21°N/S) for the full AM2 model.
Fig. 4.50: Vertical cross-sections of the zonally-averaged variables, as indicated, regressed upon the standardized PC 1 time series of the zonally-averaged streamfunction $[\Psi]$ field for the aqua-planet AM2 model. $d[u]/dt$ refers to the 2-day time rate of change of $[u]$. Contour intervals are $0.2 \times 10^{10}$ kg s$^{-1}$ for $[\Psi]$, 0.2 m s$^{-1}$ for $[u]$, 1 m$^2$ s$^{-2}$ for $[u*v*]$, and 0.04 m s$^{-1}$ d$^{-1}$ for $d[u]/dt$. Dashed contours denote negative values. Shading represents correlations $\geq 0.2$ (significant above the 99% level).

Fig. 4.51: As in Fig. 4.50 but regressed upon the standardized PC 2 time series of the $[\Psi]$ field. $d[u]/dt$ refers to the 10-day time rate of change of $[u]$ with 0.02 m s$^{-1}$ d$^{-1}$ contour intervals.
Fig. 4.52: As in Fig. 4.50 but for the full AM2 model based on nonseasonal data.

Fig. 4.53: As in Fig. 4.51 but for the full AM2 model based on nonseasonal data.
Fig. 4.54: Leading EOFs of the nonseasonal, zonally-averaged meridional wind [$v$] field for the full AM2 model. EOFs 1-3 explain 10%, 6%, 5% of the total variance. Contour intervals are 0.1 m s$^{-1}$. Dashed contours denote negative values.
Fig. 4.55: Vertical cross-sections of the zonally-averaged (top) meridional wind \([v]\) and (bottom) streamfunction \([\Psi]\) regressed upon the leading intrinsic mode functions (IMFs) of high to low frequency based on Ensemble Empirical Mode Decomposition of the standardized PC 1 time series of the nonseasonal \([v]\) field for the full AM2 model: IMFs with a period of (a) \(~3-4\) days and (b) \(\geq 14\) days. Contour intervals are 0.1 m s\(^{-1}\) for \([v]\) and 0.2 \(\times 10^{10}\) kg s\(^{-1}\) for \([\Psi]\). Dashed contours denote negative values. Shading represents correlations \(\geq 0.3\) (significant above the 99% level).
Fig. 4.56: Vertical cross-sections of the zonally-averaged zonal wind [$u$] field regressed upon the standardized PC 1-3 time series of the nonseasonal, zonally-averaged angular momentum field ($[u]\cos\phi$, where $\phi$ is latitude; 46°N-46°S, 74-975 hPa) for the (a) aqua-planet and (b) full AM2 models. EOFs 1-3 explain 18%, 17%, 15% (14%, 14%, 11%) of the total variance in the aqua-planet (full) AM2 model. Contour intervals are 0.5 m s$^{-1}$. Dashed contours denote negative values.
Fig. 4.57: Vertical cross-sections of the correlation coefficients between the daily nonseasonal time series of zonally-averaged zonal winds at the same latitude of the Northern and Southern Hemispheres (e.g., \([u]\) at 5\(^\circ\)N and 5\(^\circ\)S) at each pressure level: (a) aqua-planet and (b) full AM2 models.
Table 4.1: Total variance of the sum of IMF time series 1-11 and percent contribution of the IMFs to the total variance of PC 1 and 2 time series of zonally-averaged streamfunction field ([Ψ]_{PC1}, [Ψ]_{PC2}); zonally-averaged, 150-hPa meridional and zonal wind components over the equator ([v]_0, [u]_0); even and odd components of 150-hPa level [u] at 12°N/S ([u]_{12°-even}, [u]_{12°-odd}); MJO indices of Wheeler and Hendon (2004; MJO 1, MJO 2); MJO-residual [u]_0 ([u]_{0-MJOres}); 150-hPa [u] at 25°N during Nov-Mar ([u]_{25N}); zonally-averaged northward eddy flux of westerly momentum [u*v*] at 10°N/S and 150 hPa ([u*v*]_{10°,150hPa}); and [u*v*] at 45°N/S and 300 hPa ([u*v*]_{45°,300hPa}); MJO 1, MJO 2; MJO-residual [u]_0 ([u]_{0-MJOres}); 150-hPa [u] at 25°N during Nov-Mar ([u]_{25N}); zonally-averaged northward eddy flux of westerly momentum [u*v*] at 10°N/S and 150 hPa ([u*v*]_{10°,150hPa}); and [u*v*] at 45°N/S and 300 hPa ([u*v*]_{45°,300hPa}). The period range (in days, determined by the average time interval between the peaks in the time series) of the IMFs of the time series shown here are indicated in parentheses.

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<th>4 (28-52)</th>
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Table 4.2: Ratio of the root-mean-squared amplitudes of the components of the zonally-averaged northward eddy flux of westerly momentum $[u^*v^*]$ at the respective locations: fluxes due to the modulation of stationary eddies ($[U^*v^*'+ u^*V^*]$) and by transient eddies ($[u^*v^*']$).

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Chapter 5

Conclusions

In this dissertation, we have illuminated various aspects of the tropical general circulation. The study represents an effort to close the gap in our understanding of the atmospheric general circulation by touching upon a variety of issues that had been difficult to address in previous years, mainly due to data limitations.

In Chapter 2, we have shown that the zonally-averaged structures of the thermally-direct tropospheric Hadley circulation and the thermally-indirect stratospheric Brewer-Dobson circulation (BDC) are fundamentally different. The agreement between the patterns derived from multiple data sources – temperature, meridional wind and radiative heating rates from two reanalysis data sets and satellite-derived measurements of cloud, water vapor, ozone and carbon monoxide – illustrates the clear distinction between the tropospheric and stratospheric circulation cells, with regard to the meridional width of the tropical upwelling and the equatorial symmetry of the seasonal cycle structure. The relatively narrow tropical upwelling branch of the Hadley circulation is in striking contrast to the broad tropical upwelling in the BDC. The seasonal cycle, represented as the difference between Jan-Feb and Jul-Aug means, exhibits patterns with odd equatorial symmetry up to ~150 hPa and predominantly equatorially-even patterns above that level, consistent with the annual cycles of the Hadley circulation and BDC, respectively.

Data with increased vertical resolution in the upper troposphere and lower stratosphere are needed to more precisely identify the extent to which tropospheric and stratospheric influences are observed within the tropical tropopause layer (TTL), though it is debatable whether a clear separation of the TTL processes is possible. Future investigations of this type may lead to a better understanding of the recent observed trends in the TTL (e.g., the slight cooling trend in the TTL, the increase in stratospheric WV) and how the changes in the TTL might feed back upon the global climate system.
They may also help to settle the unresolved issue regarding the forcing of the tropical upwelling in the BDC, as discussed in Chapter 3.

Future work will be needed to reconcile the discrepancy between our empirical results, which indicate that high-latitude eddy heat fluxes associated with breakdowns in the polar night jet play the dominant role in forcing variations in tropical upwelling, and the results of our model study and those in a number of previously published studies cited in Chapter 3, which indicate that a low-latitude forcing is required to drive variability in tropical upwelling as large as the observed. A more realistic, non-uniform radiative damping time scale in the idealized axisymmetric dry model could, to some degree, modify the structure of the meridional profiles of lower-stratospheric temperature to better match those of the observations. The latitudinal structure of the stratification and the feedback of the dynamically-induced ozone variations on temperature also appear to be important factors that should be included in mechanistic models of the interannual variability and trends in lower-stratospheric temperature (Fueglistaler et al. 2010). However it appears unlikely that such modifications to our dry model will yield substantially different results from those presented in this study.

The response of the zonal wind and temperature fields to a prescribed easterly torque in the presence of a stratospheric polar vortex, shown in Fig. 5.1, offers some clues as to why the vertical motions in the forced mean meridional circulation (MMC) are concentrated at the edges of the forcing. In contrast to the complete breakdown of the polar night jet during sudden warming events in the real atmosphere, the wave drag in the model simply redistributes the zonal momentum and reshapes the jet; warmer temperatures and weaker westerlies are observed within the forcing region, whereas cooler temperatures and stronger westerlies are observed at the edges of the forcing. Thus it is apparent that a forcing (either in the form of an easterly torque or EP flux divergence) prescribed over a fixed region in the model domain does not accurately represent the forcing distribution associated with the sudden warmings.

Preliminary results of two independent studies of G. Chen and E. Gerber based on idealized dry model simulations show that waves generated over a prescribed mountain in
the Northern Hemisphere extratropics are capable of inducing a global-scale MMC response with broad tropical upwelling (*personal communication*). The breaking of planetary waves in the vicinity of the polar vortex (~40°-70°N) produces sudden warming events and accounts for the bulk of the global-scale residual circulation in Gerber’s model, consistent with observations. Additionally, Gerber finds smaller-scale waves (i.e., higher wavenumbers) at lower latitudes, which are presumably related to the high-latitude planetary waves, that appear to be playing a meaningful role in forcing the mean meridional circulation equatorward of ~40°N.

It is plausible that the small-scale features simulated in Gerber’s model experiments, but not well resolved in global reanalysis data sets, may significantly modulate the tropical upwelling in the BDC. It will be informative to diagnose the role of these small-scale features by examining how far equatorward these features extend and to what extent their extensions into lower latitude influence the strength of the tropical upwelling. Future work will be needed to clarify the phenological origin of these waves and their effect on tropical upwelling.

In Chapter 4, we have shown that the nonseasonal variability of the zonally-averaged zonal wind [$u$] in the equatorial upper troposphere is primarily influenced by variations in precipitation over the Indo-Pacific warm pool region. A pulsing of convection over the warm pool gives rise to equatorial Kelvin and Rossby waves that can be traced all around the globe, which conspire to produce a rectified zonally-symmetric zonal wind variability. Fluctuations in the warm pool precipitation explains much of the variance of equatorial [$u$], but contributes very little to the variations in [$u$] at subtropical latitudes.

Subtropical [$u$] variability imparts an observable response in equatorial zonal wind by inducing an equatorially-symmetric Rossby wave couplet that straddles the equator in the Pacific sector. The variability appears to be forced, in part, by the variations in the eddy momentum fluxes at extratropical latitudes. Our results are consistent with the lagged tropical [$u$] response to extratropical annular modes documented by Thompson and Lorenz (2004), but we have shown that this tropical-
extratropical linkage is more pervasive than implied by their study. Compositing analysis based on the El Niño Southern Oscillation (ENSO) phases may yield further insights on the Rossby wave pattern in the central Pacific, which projects strongly upon the ENSO pattern. Another factor that could potentially modulate \( u \) at subtropical latitudes is the variability of the climatological-mean westerly jet streams, particularly the wintertime jet stream over the North Pacific. Instability of the jet at the jet exit regions over the Pacific may generate intraseasonal fluctuations in \( u \) at subtropical latitudes with an equatorial Rossby wave pattern, thereby influencing the \( u \) over the equator.

We have presented preliminary results pertaining to the observed tropical annular variability, obtained from analyses of the daily data in a hierarchy of models. Because the Atmospheric Model 2 (AM2) model with topography and seasonally-varying sea surface temperatures is the only model capable of simulating the tropical \( u \) variability reasonably well, the AM2 model (or a model of similar complexity) can be used to study the role of the equatorial stationary waves in the tropical \( u \) variability. From the observations alone, it is not apparent how and why the equatorial waves widen over the course of the \( u \) evolution, which appears to give rise to the poleward propagation of \( u \) anomalies in the zonal mean. It will be informative to first test whether the equatorial wave features similar to those in the ERA-Interim are observed in the AM2 model with a seasonal cycle, but no ENSO cycle. The seasonal dependence of the results (i.e., the finding that the relevant features are observed only during the boreal cold season) can also be further explored.
Fig. 5.1: Vertical cross-sections of the time-mean, zonally-averaged (a) temperature (K) and (b) zonal wind (m s\(^{-1}\)) for axisymmetric simulations with a polar vortex in the Northern Hemisphere (Fig. A.3 with \(\gamma = 2\)) and constant tropical \(T_{eq}\) (Fig. A.4): (left) no forcing, and (right) 180-day periodic forcing of 0.3 m s\(^{-1}\) d\(^{-1}\) in the region 30-60\(^{\circ}\)N, 0-50 hPa.
References


Appendix

Model Descriptions

In the foregoing chapters, we have explored various aspects of the zonal-mean tropical atmosphere based on various observational data sets. While these observational findings are certainly valuable, our interpretation of them is subject to the limitations of the data sets, including their spatial and temporal resolution, quality of measurements, and the length of record. Reanalysis data with full global coverage and continuous record span are useful for studying the large-scale atmosphere, but they too suffer from the fact that the data are not absolutely dynamically consistent because of errors in the input observations. Numerical model simulations of the physical system provide an alternative method of studying the atmosphere. Aside from an important caveat that the model output data are not “real”, model data are dynamically consistent by design and thus allow quantitative predictions to be made and tested. When used in conjunction with observational analysis, models are particularly useful for understanding the underlying dynamics of a complex system such as the atmosphere.

Idealized models are relatively easy to understand and the results derived from them tend to be robust and easily reproducible. They offer a good testing ground for theories of the general circulation. Needless to say, however, idealized models are limited to the investigation of a specific set of problems whose applicability to the real world may be questionable. A useful approach is to experiment with multiple models of varying complexity, from the simplest idealized model to a full atmospheric general circulation model (GCM). Brief descriptions of the hierarchy of models used to investigate key elements of the observational findings are given in the following sections and summarized in Table A.1.

A.1 Dry dynamical core model

To demonstrate the use of their proposed benchmark calculations for evaluating and comparing the dynamical cores of atmospheric general circulation models (GCM),
Held and Suarez (1994) introduced a standard Eulerian spectral dynamical core model consisting of the primitive equations on a sphere with no topography. Radiation and convection are parameterized as Newtonian cooling to a prescribed radiative equilibrium temperature \(T_{eq}\) profile. The only other physical parameterization in this model is Rayleigh damping of low-level winds to represent boundary layer friction. Because the model lacks any moist physics and the complications that accompany it, we refer to this model of least complexity as the “dry dynamical core model (dry model)”.

The lack of treatment of moisture including precipitation and clouds is an obvious limitation of the dry model. Other weaknesses include the fact that the atmosphere is not heated from below as in the real atmosphere. Land-sea contrast cannot be taken into account due to the absence of surface fluxes. Diabatic processes tend to be quite weak in contrast to the active baroclinic eddies. For the purpose of our study of the tropical atmosphere, we also bear in mind that the tropics in this model is very quiet. Despite its limitations, the simple formulations make the dry model a useful tool for testing basic theories of the general circulation. The model is able to reproduce the general features of the atmosphere quite well and thus has been widely used to study many problems involving dry dynamics (e.g., Franzke 2002; Seager et al. 2003b; Kushner and Polvani 2004; Harnik and Chang 2004; Franzke et al. 2004; Kraucunas and Hartmann 2005).

Except where specifically stated, we ran the model at 2.8° latitude x 2.8° longitude resolution using the default settings of Held and Suarez (1994) as described in their article. The zonally-symmetric \(T_{eq}\) profile in 20 equally-spaced \(\sigma = p/p_0\), where \(p\) is any given pressure level and \(p_0\) is the surface pressure level) levels, shown in Fig. A.1a, is symmetric about the equator with an equator-to-pole temperature gradient of 60 K. \(T_{eq}\) decreases with height up to 200 K above which the atmosphere (i.e., stratosphere) becomes isothermal. The vertical gradient of potential temperature at the equator is specified as to yield a roughly moist adiabatic vertical structure of the tropical troposphere. Temperatures relax to this statically-stable \(T_{eq}\) profile at a time scale of \(k_T = (40 \text{ d})^{-1}\) in the free troposphere and stratosphere and at a much shorter time scale \(k_T = (4 \text{ d})^{-1}\) within the boundary layer \((\sigma_b = 0.7)\) to prevent the formation of a strong
inversion near the surface. Linear damping of winds is also strong within the boundary layer with maximum $k_u = (1 \ d)^{-1}$ at the surface and decreasing to zero at the top of the boundary layer. A complete description of the dry dynamical core model can be found in Held and Suarez (1994). The climatological features of the dry model presented in following sections are based on an analysis of 1600 days of data after spin up.

A.1.1 Adjustments to the Model Stratosphere

An isothermal atmosphere of 200 K in Fig. A.1 is obviously a misrepresentation of the real stratosphere, which is characterized by a strong polar vortex in the winter hemisphere. Furthermore, the equally-spaced $\sigma$-coordinate system does not sufficiently resolve the temperature structure $\leq \sim 200 \ hPa$, as shown in Fig. A.1b. Figure A.2 shows the number of vertical levels in a given pressure layer when the model is resolved by 20 and 40 equally-spaced $\sigma$ levels. Only 2 of the total 20 vertical levels are located in the stratosphere above the 100-hPa level (Fig. A.2a). The stratosphere is coarsely resolved with only 4 levels, even when the total number of model levels is doubled (Fig. A.2b). Thus, in order to use the dry model to the study of the Brewer-Dobson circulation, some adjustments to the Held and Suarez (1994) set-up are needed. We followed the steps in Polvani and Kushner (2002), who used a modified version of the Held and Suarez (1994) model to examine the tropospheric response to stratospheric perturbations.

First, the vertical resolution was increased in the stratosphere. Rather than dividing the atmosphere into equally-spaced $\sigma$ levels, which concentrates the levels near $p_0$ (Figs. A.1 and A.2), the locations of $n$ model levels ($s_i$, where $0 \leq i \leq n$) were determined by

$$s_i = \left(\frac{i}{n}\right)^x$$  \hspace{1cm} (A.1)

where $x = 5$. Polvani and Kushner (2002) found that the sensitivity of the magnitude and latitude of surface zonal wind maximum in the winter hemisphere to the model’s vertical resolution was generally small between $n = 40$ and 80, but can be quite substantial
between $n = 20$ and 40 under certain circumstances. Based on their results and because it is computationally less expensive to run a model with fewer vertical levels, we chose $n = 40$ model levels for our study of the Brewer-Dobson circulation. Compared to the vertical resolution of the Held and Suarez (1994) dry model whose levels are determined by Eqn. A.1 with $x = 1$, the stratosphere in this modified version of the dry model is highly resolved between 0 and 100-hPa levels, as shown in Fig. A.2b. 40 vertical levels distributed based on Eqn. A.1 were then located in the model using the Simmons and Burridge (1981) scheme, which resulted in 24 levels in the 0-100 hPa layer and 4 levels in the 101-200 hPa layer. As in the model of Polvani and Kushner (2002), we added a sponge layer in the top six levels ($p < 0.5$ hPa) so that the waves reaching the top of the model domain are linearly damped with a damping coefficient increasing from zero at $p = 0.5$ hPa to a maximum of $k_{sp} = (0.5 \, d)^{-1}$ at the top-most level.

Figure A.5a shows the vertical cross-section of the time-mean, zonally-averaged temperature based on this $T_{eq}$ profile with 40 vertical levels. The zonal-mean temperature profile within the troposphere resembles the $T_{eq}$ profile with slightly flatter temperatures at low latitudes. The temperature minimum at the tropical cold point is clearly observed. The zonally-averaged zonal wind field, shown in Fig. A.6a, consists of a westerly jet confined within the troposphere of both hemispheres. Weak easterlies dominate the tropics except in the ~3-20 hPa layer.

Next, we replaced the $T_{eq}$ profile shown in Fig. A.1 with several different versions of it. First of these is a profile introduced by Polvani and Kushner (2002) that takes into account the stratospheric polar vortex of the winter hemisphere, shown in Fig. A.3 for $\gamma = 4$ K km$^{-1}$ where $\gamma$ represents the value of the constant lapse rate within the vortex region that controls the strength of the vortex. The sensitivity of our results to the value of $\gamma$ is discussed in Section 3.5.3.4. The exact formulas are given in Polvani and Kushner (2002).

Another version of the $T_{eq}$ profile used in the study of the Brewer-Dobson Circulation consists of constant temperatures (i.e., no meridional gradient of $T_{eq}$) equatorward of 30°N/S, as shown in Fig. A.4. The profile is identical to the one of Held
and Suarez (1994) shown in Fig. A.1, except that $T_{eq}$ within the tropics (30°N-30°S) is a constant value equal to $T_{eq}$ at 30°N/S. This profile was used in order to suppress instabilities that were observed near the equator in response to perturbations in the extratropical stratosphere. Compared to Fig. A.5a, the time-mean, zonally-averaged temperature field based on this $T_{eq}$ profile in Fig. A.5b exhibits more uniform and colder temperatures in the stratosphere. The structure of the time-mean, zonally-averaged zonal wind, shown in Fig. A.6b, is strikingly different from that based on the original $T_{eq}$ profile (Fig. A.6a). The tropospheric jets extend well into the stratosphere.

We also used $T_{eq}$ profiles with varying static stability in the stratosphere. The static stability was varied by prescribing a constant lapse rate between the levels $0.001 < \sigma < 0.18$ (~1 hPa $< p < ~160$ hPa) at all latitudes. Figure A.7 shows vertical profiles of zonally-averaged $T_{eq}$ for the most stable ($dT_{eq}/dz = +3$ K km$^{-1}$), neutral ($dT_{eq}/dz = 0$) and the most unstable ($dT_{eq}/dz = -3$ K km$^{-1}$) cases.

We ran an axisymmetric version of the three-dimensional (3D) model. Zonal winds in an axisymmetric model increase substantially compared to its 3D counterpart, as seen in Fig. A.6c. Flattening the $T_{eq}$ profile within the tropics reduces the magnitude of the strong jets (Fig. A.6d) such that the zonal wind profiles of 3D and axisymmetric models are quite similar. The temperature structures in the stratosphere, shown in Figs. A.5c and d, are comparable to the structure of the 3D model with flattened tropical $T_{eq}$ profile (Fig. 5b).

A.1.2 FORCING SPECIFICATIONS

As a way of parameterizing the effect of wave breaking on the stratospheric wind field ($\partial u / \partial t$), a forcing was applied in a specified region of the stratosphere in one of two forms: (1) Rayleigh friction and (2) an easterly torque. The former involves a simple linear damping of the winds ($u$) with a given damping coefficient, $k_R$:

$$\frac{\partial u}{\partial t} = -k_R \times u \quad \text{(A.2)}$$
\( k_R \) of a given value (in \( \text{d}^{-1} \)) was applied in a region of westerly winds, such as the Northern Hemisphere extratropical stratosphere in a model with \( T_{eq} \) profile shown in Fig. A.3. Thus a forcing of this type induces anomalous easterlies that would, in principle, be balanced by a residual circulation with poleward flow in the forcing region. While this method alters the wind field in a physically meaningful and self-consistent manner, it does not allow one to control of the magnitude of the parameterized wave drag due to the dependence of \( \partial u/\partial t \) on the background flow. We therefore applied a forcing of the second form:

\[
\frac{\partial u}{\partial t} = -F_u
\]  

(A.3)

Wave breaking of this form is represented as an easterly torque with a prescribed magnitude, \( F_u \) (in \( \text{d}^{-1} \)). The second method proved to be more useful than the first method in making quantitative estimates of the effect of the forcing on the tropical upwelling in the Brewer-Dobson circulation.

To examine the effect of the time dependence of the forcing, we compared the responses to steady-state and time-varying forcing. In either of the cases, the results were analyzed after the model had spun up with a forcing and reached a statistically steady state. The difference between the simulation with a forcing and one without was computed and compared among the various types of forcing. A time-varying forcing is periodic:

\[
\frac{\partial u}{\partial t} = -F_u \ast \cos\left(1 + 2\pi t / t_F\right)
\]  

(A.4)

where \( t \) is the time step and \( t_F \) is the period of the forcing. An easterly forcing of this form increases in magnitude from 0 to \( 2F_u \) and relaxes back to zero over an interval of \( t_F \) days.

In addition to these choices with regard to the forcing type, the magnitude of the forcing (either as \( k_R \) or \( F_u \)) and the location of the forcing bounded by a given set of latitudes and pressure levels were specified.
A.2 Moist dynamical core model

The next level of complexity in the hierarchy of models includes the treatment of moisture. The intermediate complexity moist GCM (or the “moist dynamical core model (moist model)” of Frierson (2005) is an extension of the dry model that includes latent heat release associated with moisture flux from the surface of a mixed-layer slab ocean. A model with a mixed-layer ocean, which calculates a prognostic surface temperature to conserve energy in the time mean, requires surface radiative fluxes to complete the surface energy budget. Hence Newtonian cooling in the dry model is replaced by a more complicated “gray radiation scheme” in the moist model of Frierson (2005). The gray radiative transfer model predicts the upward and downward fluxes, which are functions of temperature only. As a result, all radiative feedbacks related to moisture (e.g., cloud, water vapor) are suppressed, allowing one to examine the dynamical effects of water in isolation. Clouds are not considered in this model.

Drag coefficients of the surface fluxes are calculated according to the simplified version of the Monin-Obukhov similarity theory, a method widely used in GCMs for computing coefficients in their lowest layers ~10-20 m above the surface. Briefly, the Monin-Obukhov similarity theory solves for the drag coefficient based on surface stress, surface buoyancy flux and density. Under neutral conditions (i.e., a buoyancy flux of zero), the drag coefficient for winds at a given height can be solved as a function of the surface roughness height (i.e., height at which winds become zero above a “rough” surface). Solving for the drag coefficient under stable or unstable conditions (i.e., when the buoyancy flux is nonzero) is more complicated, but the simplified version of the Monin-Obukhov theory used by Frierson (2005) treats the unstable cases as if they were neutral and computes the coefficient for stable cases with an additional variable of the bulk Richardson number evaluated at the lowest model level.

Compared to the dry model in which temperatures are relaxed to a statically stable $T_{eq}$ profile, near-surface temperatures in the moist model are strongly destabilized by surface heating. Thus a boundary layer scheme is necessary to prevent numerical instability at the surface. Vertical profiles of the diffusion coefficients for momentum, dry
static energy and moisture are matched to the surface fluxes calculated in accordance to
the simplified Monin-Obukhov theory and follow a K-profile up to the top of the
boundary layer. The boundary layer depth is not fixed and varies as a function of the
specified threshold of the bulk Richardson number.

The simplified Betts-Miller convection scheme is used in the moist model of
Frierson (2005). This scheme computes precipitation rates and changes in temperature
and humidity by relaxing vertical profiles of temperatures and humidity toward “post-
convective equilibrium profiles”: a virtual pseudoadiabat for temperature and a fixed
relative humidity profile relative to the calculated temperature profile. Details on the
adjustments made to ensure that energy is conserved in the process and the treatment of
shallow (non-precipitating) convection are found in Frierson (2005).

While these added complications of the moist model make the diagnosis of the
results more difficult than those of the dry model, the relatively simple consideration of
condensation in the moist model is useful for studying the effects of moisture on climate
(e.g., Frierson et al. 2006). Having an active surface with a closed energy budget is also a
step closer to the conditions of the real world. The model is run with 2.8° latitude x 2.8°
longitude resolution and 25 unequally-spaced σ levels, with increased resolution in the
stratosphere and the boundary layer (Frierson 2005; his Fig. 2.4). We analyzed 1080 days
of data after a spin up period of 360 days.

A.3 Atmospheric Model 2

The Atmospheric Model 2 (AM2) is a GCM developed at the Geophysical Fluid
Dynamics Laboratory (GFDL) whose goal is to realistically represent the dynamics,
thermodynamics and radiative characteristics of the climate system. It consists of a
gridpoint dynamical core and comprehensive physics of radiative transfer, clouds,
boundary layer and convection, as summarized in Table A.1. A full description of this
model is found in the documentation by The GFDL Global Atmospheric Model
Development Team (2004). We list here some of the key characteristics that distinguish
AM2 from the idealized models:
• The diurnal cycle is calculated every 3 hours using full radiative transfer (18 shortwave bands and 8 longwave bands) that includes prescribed ozone and aerosol climatologies.
• Clouds are parameterized with 3 prognostic variables (cloud liquid, cloud ice, cloud fraction).
• The relaxed Arakawa-Schubert convection scheme treats convection as a spectrum of entraining plumes.
• Boundary conditions consist of topography over land and climatologically-varying sea surface temperatures (SST) over the ocean.
• Orographic gravity wave drag is parameterized.

The model is run with 2° latitude x 2.5° longitude resolution and 24 vertical levels on a hybrid coordinate system. We analyzed 4018 days (11 years) of the AM2 simulation with the parameter setting as listed above. In addition, we examined 3650 days (10 years) of data from an aqua-planet version of AM2 where the topography and climatologically-varying SST field were replaced with a motionless, mixed-layer slab ocean identical to the one in the moist model. The aqua-planet version has no seasonal cycle (i.e., the obliquity is set to zero).
Fig. A.1: Vertical cross-sections of the radiative equilibrium temperature ($T_{eq}$) field of Held and Suarez (1994) dry dynamical core model in (a) $\sigma$ and (b) $p$ coordinates.
Fig. A.2: Number of levels within a pressure layer calculated via Eqn. (A.1) with \( x = 1 \) (circle) as in Held and Suarez (1994) and \( x = 5 \) (asterisks) as in Polvani and Kushner (2002): total number of model levels, \( n = (a) \) 20 and \( (b) \) 40.
Fig. A.3: Vertical cross-section of the radiative equilibrium temperature ($T_{eq}$) field as prescribed in Polvani and Kushner (2002) with polar vortex ($\gamma = 4$) in the Northern Hemisphere.
Fig. A.4: Vertical cross-section of the radiative equilibrium temperature ($T_{eq}$) in the 100-945 hPa domain with uniform temperatures within the tropics (30°N-30°S). Constant temperature of 200 K above the 100-hPa level.
Fig. A.5: Vertical cross-sections of the time-mean, zonally-averaged temperature for (a,b) 3D and (c,d) 2D simulations with $T_{eq}$ profile as shown in (a,c) Fig. A.1 and (b,d) Fig. A.4. Newtonian damping time scale ($kT$) is specified as (a,c) (40 d)$^{-1}$ and (b,d) (60 d)$^{-1}$. Solid contours ≤ 200 K.
Fig. A.6: As in Fig. A.5 but for zonally-averaged zonal wind. Solid contours $\geq 30$ m s$^{-1}$. 
Fig. A.7: Vertical profiles of the radiative equilibrium temperature over the equator with varying static stability ($dT_{eq}/dz = -3, -2, -1, 0, +1, +2, +3$ K km$^{-1}$) in the 1-200 hPa layer.
Table A.1: A summary of the specifications of the models used in this study.

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<tr>
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<th>Dry</th>
<th>Moist</th>
<th>AM2</th>
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<tr>
<td><strong>dynamical core</strong></td>
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<td>hydrostatic</td>
<td>hydrostatic</td>
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<tr>
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<td>spectral</td>
<td>gridpoint</td>
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<td><strong>horizontal resolution</strong></td>
<td>2.8° lat x 2.8° lon</td>
<td>2.8° lat x 2.8° lon</td>
<td>2° lat x 2.5° lon (staggered Arakawa B-grid)</td>
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<td><strong>vertical resolution</strong></td>
<td>σ-coordinate (20 levels, equally-spaced)</td>
<td>σ-coordinate (25 levels, unequally-spaced)</td>
<td>hybrid-coordinate (24 levels: 9 ≤1.5 km, 5 stratosphere)</td>
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<td>specific humidity</td>
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<tr>
<td></td>
<td></td>
<td></td>
<td>cloud variables (liquid, ice, fraction)</td>
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<td>gray radiation</td>
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<td></td>
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<td>N/A</td>
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<td></td>
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<td>• climatologically-varying SST</td>
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<td><strong>boundary layer</strong></td>
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VITA

Rei Ueyama was born in Kobe, Japan, to Kazuo and Fumiko Ueyama. She temporarily relocated to Mamaroneck, New York at the age of six with her parents and sister, Maki Ueyama. Upon graduation from Keio Academy of New York, she joined her family back in Tokyo, Japan to attend the College of Science and Engineering at Keio University. She transferred to Cornell University in Ithaca, New York during her sophomore year, and earned a Bachelor of Science degree in Science of Earth Systems with a minor in Atmospheric Sciences in January 2003. She stayed on to complete a Masters of Science degree in Oceanography in May 2004. In September of 2004, she joined the Department of Atmospheric Sciences at University of Washington in Seattle, Washington, and earned a Doctor of Philosophy in Atmospheric Sciences in December of 2010.