What is Driving Changes in the Tropospheric Circulation? New Insights from Simplified Models

Neil Tandon

Submitted in partial fulfillment of the requirements for the degree of Doctor of Philosophy in the Graduate School of Arts and Sciences

COLUMBIA UNIVERSITY

2013
This thesis seeks an improved understanding of what has been driving changes in the large scale tropospheric circulation. First, we consider the effects of stratospheric water vapor levels, which exhibit significant changes on both interannual and decadal timescales. It is shown that idealized thermal forcings mimicking increases in stratospheric water vapor produce poleward expansion of the Hadley cells (HCs) and poleward shifts of the midlatitude jets. Quantitatively, the circulation responses are comparable to those produced by increased well-mixed greenhouse gases. This suggests that stratospheric water vapor may be a significant contribution to past and projected changes in the tropospheric circulation.

The second part of this thesis focuses on the response to idealized thermal forcings in the troposphere. It is found that zonally uniform warming confined to a narrow region around the equator produces contraction of the HCs and equatorward shifts of the midlatitude jets. Forcings with wider meridional extent produce the
opposite effect: HC expansion and poleward shifts of the jets. If the forcing is confined to the midlatitudes, the amount of HC expansion is more than three times that of a forcing of comparable amplitude that is spread over the tropics. This finding may be relevant to recently observed trends of amplified warming in the midlatitudes. Furthermore, a simple diffusive model is constructed to explain the sensitivity of the circulation response to the meridional structure of the thermal forcing.

The final part of this thesis considers the possible influence of solar forcing on the tropospheric circulation. Of particular interest is the steady state response to a 0.1% increase in total solar irradiance (TSI), the approximate amplitude of the 11-year solar cycle. Using a comprehensive atmospheric general circulation model coupled to a mixed layer ocean, it is found that a 0.1% TSI increase produces a circulation response that has a high dependence on the background state. Specifically, a TSI perturbation applied to a present day climate produces an equatorward shift of the Southern Hemisphere (SH) midlatitude jet, while the same forcing applied to a warmer climate produces a poleward shift of the SH jet. Opposite-signed responses are also evident in regions of the sea surface temperature, sea level pressure, and precipitation fields. These divergent responses may help to explain why earlier studies reach highly disparate conclusions about the influence of solar variations on climate.
## Contents

List of Tables iii

List of Figures iv

Acknowledgments ix

Chapter 1 Introduction 1

Chapter 2 The Response of the Tropospheric Circulation to Water Vapor–Like Forcings in the Stratosphere 8
  2.1 Introduction ......................................................... 8
  2.2 Method .............................................................. 9
  2.3 Results ............................................................... 11
  2.4 Discussion ......................................................... 17
  2.5 Appendix: Model Description ................................. 19

Chapter 3 Understanding Hadley Cell Expansion vs. Contraction 23
  3.1 Introduction ......................................................... 23
  3.2 Experiments with an Idealized GCM .......................... 26
    3.2.1 Method ....................................................... 26
    3.2.2 Results ..................................................... 30
  3.3 A Diffusive Model of The Circulation Response ............. 39
List of Tables

2.1 Thermal forcing parameters used in Equation 2.8: $p_1$ and $p_2$ are the approximate lower and upper boundaries, respectively, of the forcing region; $p_t(\phi)$ is the calculated, zonally-averaged thermal tropopause of the control integration. In the last column, $\phi$ is the latitude measured in degrees. Additional labels are used in the text to indicate the nominal perturbation amplitude, $\delta T$ (e.g. “LS-NEG5”). . . . . 21

4.1 Configurations for the integrations performed in this study. Horizontal lines separate the integrations into groups for which the associated background state is given by the reference integration, indicated in bold. . . . . . . . . . . . . . . . . . . . . . . . . . . . . . . . . . . . 63
List of Figures

2.1 Zonal mean climatology of (a) our control integration using S04/SW06 parameterizations (see Appendix for details) and (b) the Held and Suarez (1994) configuration, shown for comparison. Black contours: zonal wind, with contour interval 5 m s$^{-1}$, negative contours dotted, and zero contour omitted. Thick black contour: the thermally-defined tropopause (World Meteorological Organization, 1957). Gray contours: potential temperature, with contour interval 10 K, and contours above 380 K omitted; the 380 K isentrope marks the approximate height of our climatological tropical tropopause.

2.2 Color shading shows the different thermal forcings used in this study, described in Section 2.2: (a) lower stratospheric (LS) forcing, with $\delta T = -5$ K, constructed to mimic Forster and Shine (2002); (b) extratropical lower stratospheric (ELS) forcing, with $\delta T = -5$ K; (c) tropical lower stratospheric (TLS) forcing, with $\delta T = -10$ K; (d) identical to (a) but shifted upwards. The shading interval is 1 K. Thin black contours: the radiative equilibrium temperature of the control integration (i.e. $\delta T = 0$), with contour interval 20 K. Thick black contour: the thermal tropopause of the control integration.

2.3 Color shading: the response to the specific forcings, as indicated at the top of each column. Thin black contours: the climatology of the control integration. Solid, thick black contour: the thermal tropopause of the control integration. Dotted, thick, black contour: the tropopause in the perturbed integration. Contour intervals are 10 K for temperature (top row); 5 m s$^{-1}$ for zonal wind (middle row) with contours below 5 m s$^{-1}$ omitted; and $20 \times 10^9$ kg s$^{-1}$ for the meridional mass streamfunction (bottom row) with negative contours dashed and zero contour omitted.
2.4 Changes in circulation metrics as functions of nominal perturbation amplitude, $\delta T$. (a) The change in jet latitude, defined as the latitude of maximum zonal wind at the lowest model level. This metric isolates changes in the extratropical eddy-driven component of the jet. (b) The change in Hadley Cell width, defined as $\phi_{\Psi_02} - \phi_{\Psi_01}$, where $\phi_{\Psi_0i}$ is the $i$th zero crossing of $\Psi$ at 500 hPa, starting nearest the equator. The right-hand axis multiplies the change in Hadley Cell width by two to measure the total tropical widening (cf. Seidel et al., 2008; Johanson and Fu, 2009). (c) The change in jet speed, defined as the maximum in zonal wind at the lowest model level. (d) The change in Hadley Cell strength, defined as the maximum of $|\Psi|$ between $\phi_{\Psi_01}$ and $\phi_{\Psi_02}$. Northern and southern hemisphere values are averaged together.

2.5 As in Figure 2.3 for the LSdp-NEG5 integration.

3.1 Thermal forcings applied in our idealized general circulation model (GCM) integrations. Color shading shows the lower-tropospheric thermal forcings with shading interval 0.1 K d$^{-1}$. Black contours show potential temperature of the control integration, with contour interval of 15 K and contours above 380 K omitted. Red curves are the perturbations of the convective equilibrium lapse rate, meant to mimic the lapse-rate feedback.

3.2 The steady-state responses to the thermal forcings indicated at the top of each column. Color shading shows the difference between the climatologies of the forced and control integrations for temperature (top row), zonal wind (middle row), and meridional mass streamfunction (bottom row). Thin black contours show the climatology of the control integration, with contour intervals of 10 K for temperature (top row); 5 m s$^{-1}$ for zonal wind (middle row) with negative contours dashed; and $20 \times 10^9$ kg s$^{-1}$ for the meridional mass streamfunction (bottom row) with negative contours dashed. Positive streamfunction values indicate clockwise motion and negative values indicate counterclockwise motion. The solid, thick black contour is the thermal tropopause of the control integration. The dashed, thick, black contour is the tropopause of the forced integration.
3.3 Changes in circulation metrics as functions of meridional width of the thermal forcing, $\phi_w$. Red circles refer to standard integrations with both lower tropospheric and lapse-rate forcings. Blue circles refer to integrations with only lower tropospheric forcing and no lapse-rate forcing. Green circles refer to integrations in which the lapse-rate perturbation is increased. Empty circles indicate results from the Phi35-20 integrations. (a) The shift of the Hadley Cell (HC) edge, defined using the standard $\Psi_{500}$ metric (see text). The right-hand axis multiplies the shift of the HC edge by two to measure the total tropical widening. (b) The shift of the midlatitude jet. Positive values on the y axis indicate poleward shifts. Vertical dotted lines mark the zero crossings for the standard integrations. Northern and Southern Hemisphere values have been averaged together. 35

3.4 As in Fig. 3.2 for (left column) the Phi35_LT integration, in which there is no lapse-rate perturbation; and (right column) the Phi35_UT integration, in which the lapse-rate perturbation is twice that of the standard Phi35 integration. 36

3.5 Changes in circulation metrics as functions of relative forcing amplitude, $\alpha$. The circulation metrics are defined in the caption of Fig. 3.3 and in the text. 39

3.6 Shift of the downward maximum of residual vertical velocity in the upper troposphere, $\bar{\omega}^*_{\text{max}}$, as a function of the meridional width of the thermal forcing, $\phi_w$. Red circles refer to standard GCM integrations, blue circles refer to LT integrations of the GCM, and green circles refer to UT integrations of the GCM. Black squares show results from the diffusive model described in Sec. 3.3. Empty markers indicate results from the Phi35-20 cases. 41

3.7 Results from the diffusive model described in Sec. 3.3 for the forcings indicated at the top of each column. Thick solid lines show output from the diffusive model. Thin black lines show output from the standard GCM integrations, shown for comparison. Thick dashed lines show the imposed thermal forcings in units of temperature ($\tilde{Q}_T$). The vertical dot-dashed lines in the bottom panels indicate the latitude of the HC edge ($\bar{\omega}^*_{\text{max}}$) from the control integration. Note, for clarity the vertical scale of panel (a) is different from the other panels. 46

3.8 As in Fig. 3.7 for the Phi35-20 case. 48
3.9 Relationships between HC widening and (a) change in total Phillips’ criticality, $\delta C$; (b) change in criticality due to bulk vertical shear, $\delta C_{sh}$; and (c) change in criticality due to bulk static stability, $\delta C_{st}$. Red markers indicate standard integrations, blue markers indicate LT integrations, and green markers indicate UT integrations. Results from the Northern and Southern Hemispheres have been averaged together. Note that negative values for $\delta C_{sh}$ indicate decreases in vertical shear, and negative values for $\delta C_{st}$ indicate increases in static stability.

4.1 Color shading shows (left) the difference between the climatologies of the 0.1%S and REF integrations and (right) the difference between the climatologies of the 0.1%S(2x) and REF(2x) integrations. (See Section 4.2 and Table 4.1 for details.) Gray contours show the climatology of (left) the REF integration and (right) the REF(2x) integration, with contour intervals of (top) 10 K for temperature and (bottom) 5 m s$^{-1}$ for zonal wind with negative contours dashed. The thick black contour is the background climate’s thermal tropopause (cf. World Meteorological Organization, 1957). Black dots mark responses that are statistically significant above the 90% confidence level, based on a two-tailed t-test at each latitude/pressure level. (See Sec. 4.2 for additional details on our significance tests.)

4.2 Color shading shows (left) the difference between the climatologies of the 0.1%S and REF integrations and (right) the difference between the climatologies of the 0.1%S(2x) and REF(2x) integrations for (top row) surface temperature, (middle row) sea level pressure, and (bottom row) precipitation. Black dots mark responses that are statistically significant above the 90% confidence level, based on a two-tailed t-test at each latitude/longitude. (See Sec. 4.2 for additional details on our significance tests.)

4.3 Timeseries of (a) latitude of the Southern Hemisphere (SH) midlatitude jet, (b) SH Hadley cell edge, (c) zonal sea surface temperature (SST) gradient in the equatorial Pacific, and (d) zonal sea level pressure (SLP) gradient in the equatorial Pacific for the (black) REF, (red) 0.1%S, (blue) REF(2x), and (green) 0.1%S(2x) integrations. A 10-year running average has been applied to all timeseries. Horizontal lines indicate climatological values. See the text for precise definitions of each metric.
4.4 Shifts of the SH midlatitude jet in response to the forcings indicated along the horizontal axis. Shifts are calculated by taking the jet latitude of the forced integration and subtracting the climatological jet latitude of its corresponding reference integration (see Table 4.1). Error bars indicate the standard deviation of the jet shifts calculated for multiple “ensemble members” obtained by splitting each timeseries into 20-year slices.
Acknowledgments

My deepest gratitude goes to my advisor, Lorenzo Polvani, who stimulated my interest in atmospheric science and guided and supported me through many twists and turns over the past five years. Many others have played pivotal roles in this journey as professors, mentors, and collaborators: Adam Sobel, Ron Miller, Ed Gerber, Mark Cane, Seok-woo Son, Sean Davis, Bill Randel, and Laura Pan. I am grateful to the reviewers and colleagues whose feedback and advice helped prepare parts of this thesis for publication: Karen Rosenlof, Shuguang Wang, Amy Butler, Thomas Reichler, Gang Chen, Tiffany Shaw, and Chris Bretherton. I thank Gus Correa and Larry Rosen for keeping the computers running and Mingfang Ting for serving on my defense committee. Finally, I thank my family and friends for loving me, supporting me, and keeping me grounded.
To my parents and my three sisters.
Chapter 1

Introduction

It is now well established that human activities are causing an increase in Earth’s global mean temperature (Solomon et al., 2007). With that matter settled, attention has been turning to the finer details of climate change: What regions will warm more than others? How will precipitation patterns change? Will droughts become more frequent? Will weather overall become more extreme? Understanding changes in the large scale atmospheric circulation is a key step towards answering these very important questions.

The large scale circulation in the troposphere consists of several prominent features. In the tropics, the Hadley Cells (HCs) transport air up near the equator and poleward to around 30° latitude, where the air descends and returns equatorward near the surface. Air in the rising branch of the HC is forced to cool, and most of its moisture condenses and precipitates. This means that air descending at the HC terminus is very dry, and most of the world’s deserts are located along this subtropical band. Thus, any shift in the location of the HC edge is very consequential for populations living in marginal climates. In particular, it can lead to devastating droughts in regions where the land was previously arable.

In the extratropics, the strong meridional temperature gradient gives rise to
eddies that become baroclinically unstable, producing a “storm track” where most midlatitude precipitation occurs. The persistent eddy stirring in this region generates a convergence of momentum that accelerates the mean flow. This produces the midlatitude jet, a pronounced maximum in the eastward flow of air. Any shift in the position of the midlatitude jet indicates a shift in the overall pattern of midlatitude precipitation, which also has major consequences for populations worldwide.

Together, the HC and the midlatitude jet are the key features of the tropospheric “zonal mean” circulation—that is, the circulation averaged over all longitudes. The edge of the HC is often considered the boundary between the tropics, where convection drives much of the weather, and the extratropics, where baroclinic eddies dominate.

The impact of increasing carbon dioxide on the zonal mean circulation has been the subject of much attention. Nearly all climate models project that, with increasing carbon dioxide, there will be poleward expansion of the HCs and poleward shifts of the midlatitude jets (e.g. Yin, 2005; Miller et al., 2006; Gastineau et al., 2008; Wu et al., 2011). These circulation changes indicate that there will also be poleward shifts of the subtropical dry zones and poleward shifts of the midlatitude storm tracks.

Looking beyond carbon dioxide, past studies have also examined the circulation response to other forcings. In particular, the effect of stratospheric ozone on the tropospheric circulation has attracted much interest. In recent decades, levels of stratospheric ozone have been declining due to human emission of chlorofluorocarbons and other ozone-depleting gases (e.g. World Meteorological Organization, 2011). The associated cooling in the stratosphere of the Southern Hemisphere (SH) has lead to a poleward shift of the SH storm track (Gillett and Thompson, 2003; Shindell and Schmidt, 2004; Arblaster and Meehl, 2006). As noted above, increased CO₂ also causes a poleward shift of the storm tracks. So the circulation response
to ozone depletion acts to amplify the SH circulation response to increased CO$_2$.

However, with the implementation of the Montreal Protocol, ozone levels are expected to recover during the 21st century (e.g. Eyring et al., 2007). Son et al. (2008) found that this recovery produces an equatorward shift of the SH midlatitude jet, a response that opposes the effect of increased CO$_2$. This takes on additional importance because most coupled models participating in the Fourth Assessment Report (AR4) for the Intergovernmental Panel on Climate Change (IPCC) include neither a well resolved stratosphere nor the photochemical processes that produce ozone. Ozone is typically prescribed in these models, and Son et al. (2008, 2009) found that whether or not they properly capture ozone recovery has a significant impact on the SH climate response in simulations of future climate change.

All this serves to illustrate that, to understand recent and projected changes in the tropospheric circulation, we have to look beyond the effect of just increasing CO$_2$, and consider other contributions. Some of these additional contributions serve to amplify or cancel the effects associated with increased CO$_2$.

In addition to documenting the responses to various forcings, there have been numerous efforts to understand the mechanisms driving the specific circulation responses. Kushner and Polvani (2004) showed that eddies play a key role in generating the tropospheric response to stratospheric perturbations. To demonstrate this, they set up an atmospheric general circulation model (AGCM) that consisted of just the zonally symmetric component of the circulation from a complete AGCM. The eddy terms from the full AGCM were then applied as an external forcing to the zonally symmetric AGCM. They found that this eddy forcing is required to produce a response in the troposphere. A thermal forcing in the stratosphere without this eddy contribution produces a response that is confined to the stratosphere. Simpson et al. (2009) gave further support to this when investigating the tropospheric response to warming in the stratosphere.
The crucial role of eddies has also been demonstrated in the circulation response to tropospheric forcings. Sun et al. (2013) showed that a meridionally broad thermal forcing in the troposphere produces a poleward shift of the jets and HC edges, resembling the response to increased CO$_2$. In the absence of eddy feedbacks, such a thermal forcing produces equatorward shifts of the jets and HC edges. Wu et al. (2012) came to a similar conclusion when investigating the transient adjustment to an instantaneous doubling of CO$_2$ in a more comprehensive GCM. Butler et al. (2011) focused on the response of eddy fluxes in relation to the slope of isentropic surfaces. They argue that the circulation response to tropical heating can be understood as the downgradient, diffusive response to an increased isentropic slope. Hartmann et al. (2000) found that the poleward shift of the jet streams is closely linked with the increase in high latitude wave refraction, while Kidston et al. (2010) and Rivière (2011) argued that the increase in the eddy length scale may be the crucial factor. Some have used baroclinicity scalings to explain the circulation response (Walker and Schneider, 2006; Lu et al., 2008), while others have focused on changes in the meridional temperature gradient (Lorenz and DeWeaver, 2007; Allen et al., 2012) or changes in tropopause height (Lorenz and DeWeaver, 2007). Thus, there has been much progress in our understanding of the mechanisms driving changes in the large scale circulation.

That said, a predictive theory for the steady-state circulation response to external forcings has remained elusive. To demonstrate this, consider the fact that the circulation response to an external forcing typically feeds back on to the meridional temperature gradient. So then it would be inaccurate to claim that changes in the meridional temperature gradient actually cause the circulation response. A truly predictive theory would have to explain both the circulation response and the change in the meridional temperature gradient.

Beyond this theoretical work, there has been much effort to document observed
changes in the large scale circulation and reconcile possible discrepancies with model simulations. What makes this a challenging task is that the wind fields that directly determine the locations of the HC edge and midlatitude jet are not directly observable over most of the globe. Thus, reanalysis products have been relied upon (e.g. Johanson and Fu, 2009; Davis and Rosenlof, 2012), but there are questions about the accuracy of such products, especially for the calculation of long term trends. Since the descending branch of the HC is a region of low cloud cover, some have attempted to infer widening of the HC from observations of outgoing longwave radiation (OLR) (Hu and Fu, 2007; Johanson and Fu, 2009). These studies show a significant widening trend in recent decades, a trend much greater than what coupled models produce (Johanson and Fu, 2009). But Davis and Rosenlof (2012) have shown that the trend from OLR data is highly sensitive to the particular OLR threshold used to define the edge of the tropics. Since there is little physical basis for the choice of a specific OLR threshold, this raises questions about whether the observed trend is robust, and whether there is a real discrepancy with models. Davis and Rosenlof (2012) have shown that different metrics for detecting the edge of the tropical belt produce very different tropical widening trends. Some of these trends agree with model simulations, and others do not.

Another recurring theme in the literature concerns the role of the tropical Pacific in driving long term changes in the tropospheric circulation. Given the strong influence of El Niño-Southern Oscillation (ENSO) on interannual timescales, it has been hypothesized that multidecadal trends in the tropical Pacific might strongly influence global climate change. This has motivated numerous studies examining long term changes in the Walker circulation and the zonal temperature gradient in the equatorial Pacific (e.g. Karnauskas et al., 2009; Tung and Zhou, 2010; Deser et al., 2010; Solomon and Newman, 2012). DiNezio et al. (2009) have shown that with increased CO₂, most coupled models exhibit an “El Niño–like” pattern of
enhanced warming in the tropical East Pacific. But Lu et al. (2008) have shown that the global response to increased CO$_2$ is very different from an El Niño. So questions remain about the global impact of long term changes in the tropical Pacific.

These earlier studies set the backdrop and provide the motivation for this thesis. Specifically, we seek to provide additional insights into what has been driving changes in the tropospheric circulation, and we approach this problem from several different angles. We begin in Chapter 2 by considering the response of the tropospheric circulation to thermal forcings that mimic increases in stratospheric water vapor (SWV). Our approach is idealized in that we impose analytically specified thermal forcings in a simplified dry GCM. We perform a series of such experiments, and these demonstrate that the circulation response to increased SWV is quantitatively on par with that produced by increased well mixed greenhouse gases. Furthermore, we manipulate the thermal forcings to determine important contributions to the overall circulation response. Specifically, we find that cooling in the extratropical stratosphere is the main driver of the overall circulation response, and that the response persists even if the forcing is elevated in altitude so that tropopause height is unperturbed.

In Chapter 3 we focus on the response to idealized thermal forcings in the troposphere. We find that zonally uniform thermal forcings confined to a narrow region around the equator produce contraction of the HCs and equatorward shifts of the midlatitude jets, resembling the circulation response under El Niño (Seager et al., 2003; Lu et al., 2008). In contrast, forcings with wider meridional extent produce HC expansion and poleward shifts of the jets, resembling the response to increased CO$_2$ (e.g. Miller et al., 2006; Lu et al., 2008). Warming concentrated in the midlatitudes produces a much stronger circulation response than a forcing of comparable amplitude that is spread over the tropics, a finding that may be relevant
to recently observed trends of amplified warming in the midlatitudes. We also construct a simple diffusive model that explains the transition from HC contraction to HC expansion.

Finally in Chapter 4, we consider the influence of small solar perturbations. Using a comprehensive AGCM coupled to a mixed layer ocean, we perform a series of experiments in which an increase in total solar irradiance (TSI) is imposed on different background states. We find that the climate response to this forcing is highly dependent on the background state. Specifically, a forcing imposed on a present day climate produces a response that is in many respects opposite to that of a forcing applied to a doubled CO₂ climate. This may partly explain why past studies have reached such disparate conclusions about the influence of the 11-year solar cycle on climate.

A common thread throughout this work is the desire to employ simplified models that reduce computational expense and thus allow for more exploration of the important factors driving the circulation responses. That said, we also make a conscious effort to perform sensitivity tests and compare with other studies to ensure that our key results do not depend greatly on the specifics of a particular model configuration. With this approach, we can reach more confident conclusions about what is driving changes in the tropospheric circulation.
Chapter 2

The Response of the Tropospheric Circulation to Water Vapor–Like Forcings in the Stratosphere

2.1 Introduction

Water vapor is the dominant greenhouse gas of Earth’s atmosphere, but important questions remain about its future changes and their impact on climate. This is particularly true of stratospheric water vapor (SWV), whose abundance depends on numerous factors—most importantly methane oxidation and tropical cold-point temperatures (e.g. Fueglistaler et al., 2009)—for which predictions are highly uncertain (e.g. Solomon et al., 2007, Ch. 10). Data over select regions have shown a $\sim 0.5$ ppmv decade$^{-1}$ increase in SWV for much of the 20th century (Rosenlof et al., 2001; Hurst et al., 2011). After a decrease of $\sim 0.4$ ppmv between 2001-2005 (Randel et al., 2006; Solomon et al., 2010), SWV levels have increased $\sim 0.5$ ppmv since 2006 (Hurst et al., 2011).

As for the effects of SWV changes, earlier studies have focused almost ex-
clusively on the radiatively-forced temperature response. Using fixed dynamical heating (FDH) and energy balance models, Forster and Shine (2002) have shown that an increase in SWV causes cooling in the stratosphere (enhanced in the extratropical lower stratosphere) and warming at the surface. This agrees with the FDH calculations of Maycock et al. (2011) and R. Portmann (personal communication), and with the general circulation model (GCM) simulations of Oinas et al. (2001) and Smith et al. (2001). That said, the circulation response to SWV change has received very little attention. One study (Joshi et al., 2006), using a comprehensive GCM, has shown that SWV increase is a significant contributor to recent strengthening of the North Atlantic Oscillation.

This leaves open several important questions which we consider in this study: First, how does SWV change impact the global circulation—both tropical and extratropical? Second, is the tropospheric circulation response preferentially driven by specific regions of stratospheric cooling? Third, how does the circulation response vary with the amount of SWV change? To address these questions, we employ an idealized, dry GCM in which we impose thermal forcings that mimic the effects of SWV. Using such a model affords greater control over the placement and magnitude of the thermal forcing. Thus we develop a clearer picture of not only the global circulation response to SWV, but also the precise contributors to that response.

2.2 Method

In this study, we use a model consisting of only dry dynamics and highly idealized parameterizations borrowed from Schneider (2004) and Schneider and Walker (2006), hereafter “S04” and “SW06” respectively. In contrast to the more popular Held and Suarez (1994) configuration (e.g. Polvani and Kushner, 2002; Haigh et al., 2005; Lorenz and DeWeaver, 2007; Butler et al., 2010), our choices produce a
tropical climatology that is closer to observations: The tropical tropopause reaches a potential temperature of 370-380 K, and static stability reflects a moist adiabatic lapse rate (Fig. 2.1a), compared to the 340-350 K tropopause and dry adiabatic lapse rate of Held and Suarez (1994) (Fig. 2.1b). For complete reproducibility, all model details are provided in the Appendix. Suffice it to say, our model produces a hemispherically symmetric mean circulation, resembling equinoctial conditions.

To study SWV changes, we impose a number of thermal forcings, consisting of perturbations of the model’s radiative equilibrium temperature. The amplitude of each perturbation is controlled primarily by a parameter, $\delta T$, and additional parameters control its latitudinal and vertical structure. In Figure 2.2, we provide visual examples, with complete details available in the Appendix. The “LS” (“lower stratosphere”) forcing (e.g. Fig. 2.2a) mimics the temperature response due to a uniform increase in SWV, with strong cooling in the extratropical lower stratosphere, and weaker cooling in the tropics and higher altitudes (Forster and
Shine, 2002; Maycock et al., 2011). As a point of reference, the LS-NEG5 case \((\delta T = -5 \text{ K})\) corresponds roughly to a 5 ppmv increase in SWV.

With the “ELS” forcing (e.g. Fig. 2.2b), we isolate the extratropical portion of the LS perturbation, and with the “TLS” forcing (e.g. Fig. 2.2c) we isolate the tropical portion. As an additional test, we apply the “LSdp” forcing (e.g. Fig. 2.2d), in which the LS perturbation is shifted in altitude. When considering specific perturbation amplitudes, we append the above shorthand appropriately; e.g. “LS-POS5” indicates a LS forcing integration with nominal amplitude \(\delta T = +5 \text{ K}\).

For each of the above forcing functions—and for each choice of \(\delta T\)—we start our model from rest and integrate for 10,000 days, which is sufficient to obtain a statistically stationary response. To compute all climatological fields, we discard the first 300 days as spin-up and time-average the rest. To obtain the “response” of the model, we subtract the climatology of the control integration, in which \(\delta T = 0\) (shown in Fig. 2.1a). Since there is no topography in this model, and all forcings are hemispherically symmetric, all responses should be hemispherically symmetric. Any asymmetry that remains is due to sampling error. Unless otherwise stated, all integrations are performed at spectral resolution T42 with 40 vertical levels. (See Appendix for details.)

2.3 Results

We first consider the effect of a thermal perturbation that mimics a 5 ppmv increase in SWV, corresponding to our LS-NEG5 forcing. The response is shown in Figure 2.3, left column. Because the stratosphere is close to radiative equilibrium, the temperature response (Fig. 2.3a) shows a strong resemblance to the thermal forcing (Fig. 2.2a). The cooling in the stratosphere produces a \(\sim 30 \text{ hPa}\) rise in polar tropopause height (thick dotted contour), with gradually vanishing tropopause changes towards lower latitudes.
Figure 2.2: Color shading shows the different thermal forcings used in this study, described in Section 2.2: (a) lower stratospheric (LS) forcing, with $\delta T = -5$ K, constructed to mimic Forster and Shine (2002); (b) extratropical lower stratospheric (ELS) forcing, with $\delta T = -5$ K; (c) tropical lower stratospheric (TLS) forcing, with $\delta T = -10$ K; (d) identical to (a) but shifted upwards. The shading interval is 1 K. Thin black contours: the radiative equilibrium temperature of the control integration (i.e. $\delta T = 0$), with contour interval 20 K. Thick black contour: the thermal tropopause of the control integration.
Figure 2.3: Color shading: the response to the specific forcings, as indicated at the top of each column. Thin black contours: the climatology of the control integration. Solid, thick black contour: the thermal tropopause of the control integration. Dotted, thick, black contour: the tropopause in the perturbed integration. Contour intervals are 10 K for temperature (top row); 5 m s\(^{-1}\) for zonal wind (middle row) with contours below 5 m s\(^{-1}\) omitted; and 20 \(\times\) 10\(^9\) kg s\(^{-1}\) for the meridional mass streamfunction (bottom row) with negative contours dashed and zero contour omitted.
Fig. 2.3b shows the zonal wind response to the LS-NEG5 forcing. There is deceleration of $\sim 1.5 \text{ m s}^{-1}$ on the equatorward side of the jet, accompanied by $\sim 1.5 \text{ m s}^{-1}$ acceleration on its poleward side, indicating a poleward shift in the jet. Fig. 2.3c shows the changes in the meridional mass streamfunction, $\Psi$. (See Peixoto and Oort, 1992, Sec. 7.4.3 for the definition.) In the northern hemisphere, there is a slight decrease in $\Psi$ within the Hadley Cell, a more substantial increase in $\Psi$ at the Hadley Cell edge, and a substantial decrease in $\Psi$ near the poleward edge of the Ferrel Cell. These features indicate a weakening and widening of the Hadley Cell and a poleward migration of the Ferrel Cell. The same applies in the southern hemisphere but with the sign of $\Psi$ reversed.

A natural question arises next: Which latitudes of SWV change contribute most to this response? To answer this, we show the ELS-NEG5 integration, in which cooling is confined to the extratropical lower stratosphere (Fig. 2.3d). In this case, the tropospheric response is nearly indistinguishable from LS-NEG5, showing a poleward shift of the jets (Fig. 2.3e), expansion and weakening of the Hadley Cells, and a poleward shift of the Ferrel Cells (Fig. 2.3f). This suggests that most of the tropospheric response in LS-NEG5 is attributable to cooling in the extratropical lower stratosphere.

To complement ELS-NEG5, we also present the TLS-NEG10 integration, in which cooling is confined to the tropical lower stratosphere (Fig. 2.3g). To elucidate the tropospheric response, which is weaker than with extratropical forcing, we here increase the perturbation amplitude to $\delta T = -10 \text{ K}$. Notice that with tropical forcing, the response is qualitatively opposite to that of LS-NEG5 and ELS-NEG5, with a slight equatorward shift of the jets (Fig. 2.3h), contraction of the Hadley Cells, and an equatorward shift of the Ferrel Cells (Fig. 2.3i). For lower amplitudes of tropical cooling (not shown), qualitatively similar features result, but the response is very weak and much longer integrations are required to reach equilib-
rium. These findings confirm that, in LS-NEG5 (Fig. 2.3a-c), cooling in the tropical stratosphere contributes negligibly to the tropospheric response.

We have extensively explored the space of forcing functions and amplitudes, and Figure 2.4 summarizes the results. Each panel shows several curves, each one representing a set of integrations with the same thermal forcing structure but different forcing amplitudes. For stratospheric cooling ($\delta T < 0$), both the LS (red circles) and ELS (blue circles) forcings produce poleward-shifted jets (Fig. 2.4a) and expanded Hadley Cells (Fig. 2.4b). Because of the hemispheric symmetry of our model, expansion of the Hadley Cells indicates a poleward shift of the Hadley Cell edges, which also indicates poleward migration of the Ferrel Cells. These results are also apparent from Fig. 2.3, discussed above. For $\delta T < 0$, the LS and ELS forcings also produce an increase in jet speed (Fig. 2.4c) and a weakening of the Hadley circulation (Fig. 2.4d). All these results, one should note, are robust to changes in the model’s vertical resolution: Whether it is doubled to 80 levels (red squares) or halved to 20 levels (red crosses), the results are essentially identical to the standard 40-level case, over the full range of $\delta T$.

The TLS integrations, with forcing confined to the tropics, are indicated by the green circles in Fig. 2.4. TLS shows a $\delta T$ dependence opposite to that of LS and ELS, as already illustrated in Fig. 2.3, and the response is comparatively small over the full range of $\delta T$. This further confirms that the tropical stratosphere contributes very little to our model’s response. This does not mean, however, that the LS results are simply the linear composite of ELS and TLS. For example, the shift of the jet in the LS-POS10 case ($-4.34 \pm 0.04^\circ$) is actually larger in magnitude than for ELS-POS10 ($-3.42 \pm 0.04^\circ$), while linearity would suggest a smaller shift.

The results of Fig. 2.4 are nonlinear in another sense: The response to stratospheric warming ($\delta T > 0$) is not simply the reverse of the one associated with stratospheric cooling ($\delta T < 0$). This nonlinearity is most extreme in the cases of
Figure 2.4: Changes in circulation metrics as functions of nominal perturbation amplitude, $\delta T$. (a) The change in jet latitude, defined as the latitude of maximum zonal wind at the lowest model level. This metric isolates changes in the extratropical eddy-driven component of the jet. (b) The change in Hadley Cell width, defined as $\phi_{\Psi_02} - \phi_{\Psi_01}$, where $\phi_{\Psi_0i}$ is the $i$th zero crossing of $\Psi$ at 500 hPa, starting nearest the equator. The right-hand axis multiplies the change in Hadley Cell width by two to measure the total tropical widening (cf. Seidel et al., 2008; Johanson and Fu, 2009). (c) The change in jet speed, defined as the maximum in zonal wind at the lowest model level. (d) The change in Hadley Cell strength, defined as the maximum of $|\Psi|$ between $\phi_{\Psi_01}$ and $\phi_{\Psi_02}$. Northern and southern hemisphere values are averaged together.
jet speed (Fig. 2.4c) and Hadley Cell strength (Fig. 2.4d), which even show some non-monotonicity.

### 2.4 Discussion

We first offer a few points of comparison with earlier work. Joshi et al. (2006) simulate a SWV change corresponding to roughly a $-1$ K perturbation of the northern extratropical stratosphere. The magnitude of their zonal wind response is somewhat larger than what we obtain in our LS-NEG2 integration (not shown). Note, however, that Joshi et al. (2006) consider winter averages rather than the equinoxes and that their model includes topography and a polar vortex, which may explain the quantitative difference. The differences are more drastic when we compare with earlier idealized studies. For example, the zonal wind responses in Haigh et al. (2005) and Lorenz and DeWeaver (2007) are greater than ours by factors of two to four, when considering comparable forcing functions. The reason for this model-dependence requires further study. Nonetheless, our results are in qualitative agreement with earlier work.

Second, as noted earlier, the thermal forcings we have imposed produce substantial changes in tropopause height. One might argue that the tropospheric circulation response we have shown is a simple consequence of tropopause height change, a finding that is well-documented in the literature (e.g. Williams, 2006; Lorenz and DeWeaver, 2007). To show that tropopause change is not here the primary cause of the tropospheric circulation response, we present an additional series of integrations, labeled LSdp, in which the stratospheric forcing (e.g. Fig. 2.2d) is identical to the LS forcing (e.g. Fig. 2.2a) but shifted higher so as not to affect tropopause height. The response of LSdp-NEG5, shown in Figure 2.5, is qualitatively almost identical to the one in Fig. 2.3a-c, with a slightly smaller amplitude, while tropopause height is basically unchanged. We have verified that this behavior
Figure 2.5: As in Figure 2.3 for the LSdp-NEG5 integration.

holds over the entire range of forcing amplitudes, as indicated by the red triangles in all panels of Fig. 2.4.

We also note that there are slight structural differences between our analytically constructed thermal forcing and the thermal response to SWV increase shown in Forster and Shine (2002). For instance, SWV-caused cooling decreases more gradually with height than in our LS forcing. One might wonder how sensitive our results are to the details of our forcing function. To address this question, we have conducted a series of integrations, labeled WS (“whole stratosphere”), in which we
impose uniform cooling of the entire stratosphere; i.e., we cool every point above the tropopause by the same amount $\delta T$, irrespective of height and latitude. The results (Fig. 2.4, black circles) show that, over the full range of $\delta T$, there is little difference between this simplest WS forcing and the original LS forcing (red circles) that was constructed to mimic Forster and Shine (2002). This further establishes the robustness of our results.

Whether one considers our $\delta T = -5$ K integrations, or more conservatively the $\delta T = -2$ K results (which scale almost linearly), one conclusion remains the same: Changes in SWV—in particular extratropical SWV—generate circulation responses that are of the same order as the modeled response to increased well-mixed greenhouse gases. This is clear, for instance, when comparing the tropical widening in Fig. 2.4b with Fig. 2 of Johanson and Fu (2009). Indeed, this response is small when compared to observed Hadley Cell expansion, but models in general have failed to reproduce this recent trend (Seidel et al., 2008; Johanson and Fu, 2009), so the dominant contributors to tropical widening remain unclear. As for the extratropical circulation, our results show responses that are significant in the context of both past changes and future projections: Compare the zonal wind response of Fig. 2.3b,e with Figs. 3 and 6 of Lorenz and DeWeaver (2007). Thus SWV is clearly a key component of extratropical circulation change, and it may prove significant to tropical circulation change as well.

2.5 Appendix: Model Description

We use the spectral dynamical core of the Geophysical Fluid Dynamics Laboratory (GFDL) Flexible Modeling System (FMS), which employs the spectral transform method (Hoskins and Simmons, 1975) in the horizontal and the finite-difference method (Simmons and Burridge, 1981) in the vertical. The horizontal truncation is T42. The vertical level interfaces, in sigma coordinates, are $\sigma_i = (i/L)^3$, $i =$
0, 1, 2, . . . , L, where L is an integer. For all integrations, \( L = 40 \) unless otherwise stated.

To mimic convective and radiative processes, we employ two linear relaxation terms in the temperature equation,

\[
\frac{\partial T}{\partial t} = \ldots - \frac{T - T_{CE}}{\tau_C} - \frac{T - (T_R + T^*)}{\tau_R},
\]

where \( T_{CE} \) and \( \tau_C \) are the convective equilibrium temperature and timescale, respectively; \( T_R \) and \( \tau_R \) are the radiative equilibrium temperature and timescale, respectively; and \( T^* \) is our external thermal forcing.

As in SW06, \( T_{CE} \) is given by

\[
T_{CE}(\lambda, \phi, p, t) = \begin{cases} 
T_m(\lambda, \phi, p, t) - E_C(\lambda, \phi, t) & p \leq p_0 \\
T(\lambda, \phi, p, t) & p < p_{LNB}(\lambda, \phi, t),
\end{cases}
\]

where

\[
E_C(\lambda, \phi, t) = \frac{1}{p_{LNB}(\lambda, \phi, t) - p_0} \int_{p_0}^{p_{LNB}(\lambda, \phi, t)} \left[ T_m(\lambda, \phi, p', t) - T(\lambda, \phi, p', t) \right] dp'.
\]

ensures conservation of enthalpy in (2.2); \( T_m \) is the moist adiabat,

\[
T_m(\lambda, \phi, p, t) = T_s(\lambda, \phi, t) \left( \frac{p}{p_0} \right)^{\Gamma_m/g}.
\]

\( T_s \) is the surface temperature at longitude-latitude-time \( (\lambda, \phi, t) \); \( \Gamma_m = 7 \text{ K km}^{-1} \); \( p_0 = 1000 \text{ hPa} \); and \( p_{LNB} \) is the level of neutral buoyancy for ascent from the surface along \( T_m \) (e.g. Bohren and Albrecht, 1998, Sec. 6.7). Equation 2.2 is applicable only when \( E_C \geq 0 \). If \( E_C < 0 \), then convection is inhibited; i.e. \( T_{CE} = T \) in the entire column. The timescale \( \tau_C \) is set to 4 hours.

As in S04, \( T_R \) is given by

\[
T_R(\phi, p) = T_{strat} \left[ 1 + d_0(\phi) \left( \frac{p}{p_0} \right)^{3.5} \right]^{1/4},
\]

(2.5)
Table 2.1: Thermal forcing parameters used in Equation 2.8: $p_1$ and $p_2$ are the approximate lower and upper boundaries, respectively, of the forcing region; $p_t(\phi)$ is the calculated, zonally-averaged thermal tropopause of the control integration. In the last column, $\phi$ is the latitude measured in degrees. Additional labels are used in the text to indicate the nominal perturbation amplitude, $\delta T$ (e.g. “LS-NEG5”).

<table>
<thead>
<tr>
<th>Forcing</th>
<th>$p_1$ [hPa]</th>
<th>$p_2$ [hPa]</th>
<th>$w(\phi)$</th>
</tr>
</thead>
<tbody>
<tr>
<td>LS</td>
<td>$p_t(\phi)$</td>
<td>40</td>
<td>$1 - 0.4e^{-(\phi/40)^2}$</td>
</tr>
<tr>
<td>ELS</td>
<td>$p_t(\phi)$</td>
<td>40</td>
<td>$1 - e^{-(\phi/45)^6}$</td>
</tr>
<tr>
<td>TLS</td>
<td>98.5</td>
<td>40</td>
<td>$e^{-(\phi/25)^6}$</td>
</tr>
<tr>
<td>LSdp</td>
<td>$p_t(\phi)e^{-3/7}$</td>
<td>20</td>
<td>$1 - 0.4e^{-(\phi/40)^2}$</td>
</tr>
<tr>
<td>WS</td>
<td>$p_t(\phi)$</td>
<td>-1000</td>
<td>1</td>
</tr>
</tbody>
</table>

where

$$d_0(\phi) = \left[ \frac{T_0 + \Delta_h \cos^2 \phi}{T_{strat}} \right]^4 - 1,$$

(2.6)

$T_{strat} = 200$ K, $T_0 = 260$ K, and $\Delta_h = 90$ K. (See contours in Fig. 2.2.) The timescale $\tau_R$ has a vertical and latitudinal dependence given by

$$\tau_R^{-1}(\phi, \sigma) = \tau_a^{-1} + (\tau_s^{-1} - \tau_a^{-1}) \max \left( 0, \frac{\sigma - \sigma_b}{1 - \sigma_b} \right) \cos^8 \phi,$$

(2.7)

where $\sigma = p/p_0$, $\tau_a = 50$ d, $\tau_s = 7$ d, and $\sigma_b = 0.7$ (S04). We have also tested a more realistic configuration in which $T_R$ increases and $\tau_R$ decreases in the middle-upper stratosphere. While this slightly affects the magnitudes of our results, there is no qualitative change.

The thermal forcing $T^*$ is the only term that we vary, and it takes the form

$$T^*(\phi, p) = \delta T \left[ \frac{1}{1 + e^{-s_p(p-p_1)}} - \frac{1}{1 + e^{-s_p(p-p_2)}} \right] w(\phi),$$

(2.8)

where $\delta T$ controls the forcing amplitude and $s_p = 0.2$ hPa$^{-1}$. The parameters $p_1, p_2$, and $w(\phi)$ control the vertical and latitudinal structure of the thermal forcing, and their values are given in Table 2.1 for each set of integrations. Note, the bracketed term in (2.8) is approximately one for $p_2 < p < p_1$ and zero elsewhere.

There is no topography in this model. For $\sigma > \sigma_b$, winds are linearly damped as in Held and Suarez (1994), but with a surface timescale of 0.5 days. We apply
a sponge layer top and $\nabla^6$ hyperviscosity identical to that in Polvani and Kushner (2002).
Chapter 3

Understanding Hadley Cell
Expansion vs. Contraction

3.1 Introduction

How does the large-scale atmospheric circulation respond to changing temperatures? This is an important question in climate change research, and it has motivated many past studies. These include numerous idealized modeling experiments examining the circulation’s response to thermal forcings in the stratosphere (e.g. Polvani and Kushner, 2002; Haigh et al., 2005; Gerber and Polvani, 2009; Tandon et al., 2011) as well as the troposphere (e.g. Son and Lee, 2005; Kang et al., 2009; Butler et al., 2010; Wang et al., 2012). The understanding of circulation changes over the long term is often informed by the study of short-term activity, such as stratospheric sudden warmings (e.g. Gerber et al., 2009) and volcanic eruptions (e.g. Soden et al., 2002).

In particular, the study of El Niño-Southern Oscillation (ENSO) has greatly aided our understanding of circulation change in the climate context. Using a general circulation model (GCM) with forced sea surface temperatures (SSTs), Seager
et al. (2003) examined the dynamics of the El Niño–driven circulation response in great detail. They found that the short-term response to El Niño SST anomalies resembles the steady-state response to a persistent SST increase in the deep tropics. This makes for a natural comparison between the El Niño circulation response and the response to the long-term increase of greenhouse gases, commonly termed the “global warming” response.

Under global warming, most coupled models produce enhanced warming of SSTs in the eastern tropical Pacific (e.g. DiNezio et al., 2009), a pattern resembling El Niño. This led to the hypothesis that the circulation response to global warming might resemble the circulation response to El Niño. Lu et al. (2008) tested this by performing a detailed analysis of output from coupled GCMs. They found that the circulation response due to global warming is in many respects qualitatively opposite to that of El Niño. Specifically, global warming produces expansion and weakening of the Hadley Cell (HC), while El Niño produces contraction and strengthening of the HC. Also, global warming produces a poleward shift of the midlatitude jets, while El Niño produces an equatorward shift. This contrast is intriguing because both El Niño and global warming produce substantial warming of the tropical troposphere (Lu et al., 2008). This means that seemingly subtle alterations to the structure of a thermal forcing can have a dramatic effect on the circulation response. It is this sensitivity that is the focus of this study.

The results of earlier studies point to a key factor behind this sensitivity. Chang (1995) and Son and Lee (2005), using idealized dry GCMs, showed that a thermal forcing applied to a narrow region around the equator in the lower troposphere produces an equatorward shift of the midlatitude jets. This contrasts with the findings of Butler et al. (2010) and Wang et al. (2012), who found that heating with wider meridional extent in the tropical upper troposphere produces a poleward shift of the jets. Furthermore, Chen et al. (2010) have shown that changes to
the meridional structure of the SST forcing in an atmosphere-only GCM produces a transition from an El Niño–like circulation response to a global warming–like response. Altogether, these earlier studies suggest that the contrast between the global warming and the El Niño circulation responses may be attributable to either the meridional extent or the vertical structure of the thermal forcing.

This provides the inspiration for the present study. Specifically, we take an idealized GCM and apply thermal forcings of varying meridional width centered at the equator (Sec. 3.2). We show that narrow thermal forcings produce El Niño–like HC contraction, while wider thermal forcings produce global warming–like HC expansion. We also show that changes in the vertical structure of the forcing have a relatively minor effect on the circulation response. The HC turns out to be particularly sensitive to warming in the midlatitudes, a finding which may be relevant in light of recent observations. In addition, we construct a simple diffusive model of the transformed Eulerian mean (TEM) circulation to explain the transition from HC contraction to HC expansion (Sec. 3.3).

Earlier idealized modeling studies have focused either on the El Niño circulation response alone (e.g. Robinson, 2002; Seager et al., 2003) or on the global warming response alone (e.g. Kidston et al., 2010; Levine and Schneider, 2011; Rivière, 2011). Thus, it has remained unclear how the mechanisms driving the El Niño– and global warming–like responses fit into the same physical framework. By studying both phenomena together, we can develop a more comprehensive understanding of what drives changes in the tropospheric circulation.
3.2 Experiments with an Idealized GCM

3.2.1 Method

Our idealized GCM is a dynamical core forced with highly simplified radiation and convection schemes. This GCM is nearly identical to that used in Chapter 2, and we provide complete details in the Appendix. In the GCM’s radiation scheme, temperatures are linearly relaxed to a prescribed equilibrium profile which mimics a gray atmosphere (Schneider, 2004; Schneider and Walker, 2006). When a column becomes statically unstable, the temperature in the column is relaxed to a moist adiabatic profile that conserves enthalpy (Schneider and Walker, 2006). This convection scheme compensates to an extent for the lack of explicit moisture in the model. The lapse rate of the convective equilibrium profile is a prescribed parameter, and we experiment with perturbing this parameter, as described below.

Compared to dry models that use the Held and Suarez (1994) forcings (e.g. Son and Lee, 2005; Butler et al., 2010, 2011; Wang et al., 2012), the model we use produces a climatology with more realistic stratification and tropopause height in the tropics (Tandon et al., 2011).

We run the GCM in a perpetual equinox configuration with hemispherically symmetric radiative forcing. All integrations are performed at spectral resolution T42 with 40 vertical levels. (See the Appendix for additional details.) We have verified that all of our key results are robust to doubling of either the horizontal or vertical resolution.

In each integration, we impose an additional thermal forcing consisting of 1) warming of lower tropospheric temperatures, mimicking an increase in longwave opacity, and 2) a decrease of the convective equilibrium lapse rate. This lapse-rate perturbation mimics the lapse-rate feedback in a moist atmosphere, which reduces warming near the surface and amplifies it aloft. The lower tropospheric thermal
forcing, $\tilde{Q}$, takes the form of a potential temperature tendency that is added to the heat equation. Specifically,

$$\tilde{Q}(\phi, p; \phi_w; \alpha) = \frac{\alpha \tilde{Q}_0}{\phi_w} e^{-(\phi/\phi_w)^2} \left(\frac{p}{p_0}\right)^{2.4},$$

(3.1)

where $\phi$ is latitude, $p$ is pressure, $\tilde{Q}_0 = 18$ K d$^{-1}$ x 1° latitude, and $p_0 = 1000$ hPa. The meridional e-folding width of the thermal forcing is controlled by the parameter $\phi_w$, and we refer to this simply as the “width” of the thermal forcing. The factor of $\tilde{Q}_0/\phi_w$ serves to keep the area integral of $\tilde{Q}$ constant as $\phi_w$ is varied. The value of $\tilde{Q}_0$ has been chosen so that, for all thermal forcings, the globally-averaged temperature increase at the lowest model level is 2-3 K. The factor $\alpha$ is used to scale the relative amplitude of the thermal forcing; we set $\alpha = 1$ in all cases unless stated otherwise.

In addition to this lower tropospheric forcing, we also perturb the lapse rate of the model’s convective equilibrium profile. This perturbation takes the form

$$\tilde{\Gamma}(\phi; \phi_w) = \tilde{\Gamma}_0 e^{-(\phi/\phi_w)^2},$$

(3.2)

where $\tilde{\Gamma}_0 = -0.2$ K km$^{-1}$ unless stated otherwise. Note that the parameter $\phi_w$ appears in both (3.1) and (3.2), so this single parameter controls the meridional extent of both the lower tropospheric forcing and the lapse-rate forcing.

We have selected thermal forcings with a range of $\phi_w$ values to examine the El Niño–like and global warming–like responses, as well as the transition between them. We will refer to these integrations using the following labels:

- **Phi5**, with $\phi_w = 5^\circ$, is a narrow El Niño–like perturbation with peak thermal forcing between $-5^\circ$ and $5^\circ$ latitude. This forcing is shown in Fig. 3.1a.

- **Phi35**, with $\phi_w = 35^\circ$, is a wider global warming–like thermal forcing (Fig. 3.1b).

- **Phi15** ($\phi_w = 15^\circ$), **Phi20** ($\phi_w = 20^\circ$), and **Phi25** ($\phi_w = 25^\circ$) are intermediate cases, meant to examine the transition from HC contraction to HC expansion as well as the linearity of the circulation responses.
• **Phi35-20** is a special case in which we confine the lower tropospheric forcing between 20° and 35° latitude in each hemisphere, while applying a lapse-rate perturbation between −35° and 35° latitude (Fig. 3.1c). In the notation of Eqs. (3.1-3.2), the lower tropospheric forcing is

$$\frac{\phi_{w2}\tilde{Q}(\phi, p; \phi_{w2}; \alpha) - \phi_{w1}\tilde{Q}(\phi, p; \phi_{w1}; \alpha)}{\phi_{w2} - \phi_{w1}}$$  (3.3)

and the lapse-rate perturbation is \(\tilde{\Gamma}(\phi; \phi_{w2})\), where \(\phi_{w1} = 20°\) and \(\phi_{w2} = 35°\). This is qualitatively the same as the Phi35 forcing, but with the tropical lower tropospheric portion removed.

• Forcings with the additional **LT** label (e.g. Phi5_LT, Phi35_LT, etc.) are identical to the standard forcings above, except the thermal forcing is applied only in the lower troposphere without any lapse-rate forcing (i.e. \(\tilde{\Gamma}_{0} = 0\)). This is meant to test the sensitivity of the circulation response to the change in the lapse rate.

• Forcings with the additional **UT** label (e.g. Phi5_UT, Phi35_UT, etc.) are identical to the standard forcings above, except the decrease in the lapse rate is doubled, i.e. \(\tilde{\Gamma}_{0} = -0.4 \text{ K km}^{-1}\). This is comparable to the lapse-rate change in the upper troposphere in coupled GCM simulations of global warming (Lorenz and DeWeaver, 2007, Fig. 2b).

For each thermal forcing, we start the model from rest and integrate for a total of 4000 days, which is sufficient to obtain a statistically stationary climatology. To compute all climatological fields, we discard the first 200 days as spin-up and time-average the rest. To obtain the “response” of the model, we subtract the climatology of a control integration in which no thermal forcing is applied (i.e. \(\tilde{Q} = 0\) and \(\tilde{\Gamma} = 0\)). Since there is no topography in this model and all forcings are hemispherically symmetric, the model responses should be hemispherically symmetric; any small asymmetry that remains is due to sampling error.
Figure 3.1: Thermal forcings applied in our idealized GCM integrations. Color shading shows the lower-tropospheric thermal forcings with shading interval 0.1 K d\(^{-1}\). Black contours show potential temperature of the control integration, with contour interval of 15 K and contours above 380 K omitted. Red curves are the perturbations of the convective equilibrium lapse rate, meant to mimic the lapse-rate feedback.
3.2.2 Results

Fig. 3.2 shows the model responses to the three thermal forcings shown in Fig. 3.1; these forcings have the same area integral and vary only in their meridional structure. Fig. 3.2, first column, shows the response to the Phi5 forcing, which is confined to a narrow band around the equator. The peak warming (Fig. 3.2a, shading) extends to the top of the troposphere because we have imposed a decrease of the convective equilibrium lapse rate in addition to the lower tropospheric thermal forcing. In the midlatitudes, there is a local minimum in warming. There is also a slight rise in global tropopause height (thick dashed contour), where the tropopause is defined using the standard lapse-rate criterion (World Meteorological Organization, 1957).

The Phi5 zonal wind response (Fig. 3.2b, shading) shows eastward acceleration on the equatorward flanks of the midlatitude jets, indicating equatorward shifts of the jets. Near the equator, there is strong westward acceleration. Fig. 3.2c shows the response of the meridional overturning streamfunction, $\Psi$. (See Peixoto and Oort, 1992, Sec 7.4.3 for the definition.) In the Northern Hemisphere (NH), there is anomalous clockwise motion in the middle and upper portions of the HC, indicating a strengthening and deepening of the HC. There is also a counterclockwise anomaly at the poleward edge of the HC, indicating equatorward contraction of the HC and anomalous ascent in the midlatitudes. This anomalous ascent coincides with the midlatitude minimum in the temperature response (Fig. 3.2a). At the equator, $\Psi$ decreases near the surface and increases at higher levels, indicating a decrease in vertical velocity near the surface and an increase aloft. Note that the response of $\Psi$ in the Southern Hemisphere (SH) has the opposite sign, but the physical interpretation is identical. So overall, the Phi5 response resembles the El Niño circulation response of comprehensive models (Seager et al., 2003; Lu et al., 2008). One discrepancy is that the El Niño temperature response in comprehensive models shows
Figure 3.2: The steady-state responses to the thermal forcings indicated at the top of each column. Color shading shows the difference between the climatologies of the forced and control integrations for temperature (top row), zonal wind (middle row), and meridional mass streamfunction (bottom row). Thin black contours show the climatology of the control integration, with contour intervals of 10 K for temperature (top row); 5 m s$^{-1}$ for zonal wind (middle row) with negative contours dashed; and 20×10$^9$ kg s$^{-1}$ for the meridional mass streamfunction (bottom row) with negative contours dashed. Positive streamfunction values indicate clockwise motion and negative values indicate counterclockwise motion. The solid, thick black contour is the thermal tropopause of the control integration. The dashed, thick, black contour is the tropopause of the forced integration.
cooling in the midlatitudes which is not reproduced in our model (Fig. 3.2a), but the circulation responses are in agreement. Another discrepancy is that comprehensive models produce much less westward acceleration at the equator, even though the eastward anomalies in the midlatitudes are of comparable magnitude (cf. Lu et al., 2008).

We next consider the response when the thermal forcing is widened meridionally. This is captured by the results of the Phi35 integration, shown in Fig. 3.2, second column. Due to the wider thermal forcing, the peak temperature response (Fig. 3.2d) is spread wider meridionally than for Phi5, and there is a clear contrast between warming in the tropical lower troposphere and the amplified warming aloft. There is also dynamically-induced cooling in the extratropical stratosphere, similar to that found in other idealized modeling studies (Butler et al., 2010, 2011; Wang et al., 2012). As in the Phi5 integration, there is a slight increase in global tropopause height. The zonal wind response (Fig. 3.2e) shows a clear dipole of westward-eastward acceleration flanking the jet, indicating a poleward shift of the jet. The meridional streamfunction (Fig. 3.2f) shows expansion of the HCs and poleward shifts of the Ferrel Cells, although the changes in $\Psi$ are substantially lower in magnitude than for Phi5. In short, the circulation response of Phi35 resembles the global warming response of comprehensive models (e.g. Yin, 2005; Miller et al., 2006; Gastineau et al., 2008; Wu et al., 2011), and it is in most respects qualitatively opposite to the El Niño–like response of Phi5.

Note, we are not claiming that the Phi5 and Phi35 forcings are actually equivalent to the heating produced by El Niño and increased well-mixed greenhouse gases. Of course, with an actual El Niño, there is no simple external forcing: the changes in diabatic heating are internally determined by feedbacks between the atmosphere and the ocean. Our focus here is on understanding the circulation responses to various external thermal forcings, as a key step towards understanding
circulation change in more realistic models and observations. In this regard, simple thermal forcings like Phi5 and Phi35 are sufficient to produce circulation responses resembling those produced under El Niño and global warming, respectively.

Also worth noting is that even though the Phi35 forcing is spatially confined, the temperature response shows substantial warming throughout the troposphere (Fig. 3.2d). (Indeed, this is true of all the thermal forcings considered in this study.) This contrasts with the temperature responses of Butler et al. (2010, 2011), which are more spatially confined. Unlike the model used in Butler et al. (2010, 2011), our model uses a statically unstable radiative equilibrium profile and parameterized convection, but precisely how these produce differences in the temperature responses requires further work.

The fact that the circulation responses of Phi5 and Phi35 are opposite in sign leads to another question: is the system linearly additive? That is, if we apply a thermal forcing like Phi35, but remove the portion near the equator, do we actually obtain more HC expansion compared to Phi35? We address this question more rigorously below, but as a first crude test, we consider the Phi35-20 forcing. This forcing is qualitatively the same as Phi35, except that the forcing amplitude approaches zero between $-20^\circ$ and $20^\circ$ latitude in the lower troposphere (Fig. 3.1c). The temperature response (Fig. 3.2g) shows peak warming in the subtropics and midlatitudes, along with enhanced warming in the tropical upper troposphere. The zonal wind response (Fig. 3.2h) is of substantially larger magnitude than in Phi35 (Fig. 3.2e), indicating a larger poleward shift of the jets. The zonal wind anomalies are also more vertically uniform than those of Phi35. The response of the meridional streamfunction (Fig. 3.2i) is also larger than that of Phi35 (Fig. 3.2f), indicating greater expansion and weakening of the HC. Thus overall, the circulation response of Phi35-20 qualitatively resembles the global warming–like response of Phi35, but quantitatively the Phi35-20 response is greatly amplified.
Beyond these illustrative examples, we have also performed a sweep of the parameter $\phi_w$, which controls the meridional width of the thermal forcing. Fig. 3.3, red circles, shows the associated shifts of the HC edge (Fig. 3.3a) and the midlatitude eddy-driven jet (Fig. 3.3b). The midlatitude jet is located by finding the latitude of maximum zonal wind at the lowest model level. We locate the HC edge using the standard $\Psi_{500}$ metric: that is, moving poleward from the subtropical maximum of $|\Psi|$, we find the first zero crossing of $\Psi$ at 500 hPa. Note that, because of the hemispheric symmetry of our model, a poleward shift of the HC edge implies a widening of the HC, and multiplying this widening by two gives the overall widening of the tropical belt (cf. Seidel et al., 2008; Johanson and Fu, 2009; Davis and Rosenlof, 2012).

Fig. 3.3 shows that there is a smooth transition from equatorward jet shift and HC contraction to poleward jet shift and HC expansion. Interestingly, the zero crossings (vertical dotted lines) are not the same for the two metrics, showing slight HC contraction still occurs even when there is no jet shift. At these zero crossings, there is still a circulation response, but the position of the anomalies with respect to the climatology is such that no shift occurs. For example, in the Phi15 case (not shown), there is eastward acceleration centered precisely over the jet, whereas for other values of $\phi_w$, the acceleration occurs more on the flanks of the jet. Fig. 3.3 also shows the large quantitative difference between the Phi35-20 integration and the other integrations. Comparing the empty red circles with the other points, one sees that Phi35-20 produces a factor of four increase in HC expansion (Fig. 3.3a) and a factor of two increase in jet shift (Fig. 3.3b).

We have found that the amount of HC expansion and jet shift has relatively little sensitivity to the change in the lapse rate. To demonstrate this, we have performed a series of integrations in which the thermal forcings have identical meridional structures to those in Fig. 3.1, but without any changes in the lapse rate. We mark
Figure 3.3: Changes in circulation metrics as functions of meridional width of the thermal forcing, $\phi_w$. Red circles refer to standard integrations with both lower tropospheric and lapse-rate forcings. Blue circles refer to integrations with only lower tropospheric forcing and no lapse-rate forcing. Green circles refer to integrations in which the lapse-rate perturbation is increased. Empty circles indicate results from the Phi35-20 integrations. (a) The shift of the HC edge, defined using the standard $\Psi_{500}$ metric (see text). The right-hand axis multiplies the shift of the HC edge by two to measure the total tropical widening. (b) The shift of the midlatitude jet. Positive values on the y axis indicate poleward shifts. Vertical dotted lines mark the zero crossings for the standard integrations. Northern and Southern Hemisphere values have been averaged together.
Figure 3.4: As in Fig. 3.2 for (left column) the Phi35_\text{LT} integration, in which there is no lapse-rate perturbation; and (right column) the Phi35_\text{UT} integration, in which the lapse-rate perturbation is twice that of the standard Phi35 integration. These integrations with the additional label “LT,” and the results are plotted in blue in Fig. 3.3. Removing the lapse-rate perturbation results in the peak warming being located in the lower troposphere rather than the upper troposphere. However, in terms of the shifts of the jet and the HC edge, there appears to be little difference between the LT integrations and the standard ones. The LT results show a slight negative offset from their standard integration counterparts, except for a slight positive offset for the jet shift in the Phi5_\text{LT} and Phi15_\text{LT} cases.

Fig. 3.4, left column, shows the response of the Phi35_\text{LT} integration in more de-
tail. Comparing the temperature response (Fig. 3.4a) with that of Phi35 (Fig. 3.2d), we see much less warming in the tropical upper troposphere and enhanced warming in the lower troposphere. Phi35 does show some westward acceleration in the tropical upper troposphere that is not apparent in Phi35_{LT} (compare Fig. 3.2e and Fig. 3.4b), but aside from that, the circulation responses are nearly indistinguishable. When we compare the other LT integrations to the standard integrations, the differences are all minor. The most noticeable differences are in the Phi5_{LT} integration (not shown): at the equator, there is no westward acceleration at upper levels, no deepening of the HC, and no vertical deceleration near the surface. (Compare this with Fig. 3.2b,c.) As noted above, our Phi5 integration produces much greater westward acceleration at the equator than in comprehensive model simulations of El Niño, so our results suggest that convection plays an important role in the equatorial circulation response. As for the Phi35-20_{LT} integration (not shown), the zonal wind response is slightly more barotropic than that of Phi35-20 (Fig. 3.2h).

We have also tested the effect of imposing a larger decrease of the lapse rate. These integrations are given the “UT” label and are plotted in green in Fig. 3.3. These integrations show a slight positive offset from the standard integrations. So for the global warming–like (large $\phi_w$) cases, decreasing the lapse rate does result in enhanced HC expansion, but this effect is small compared to the effect of changing $\phi_w$. Overall, the circulation responses of the UT integrations are qualitatively similar to those of the standard integrations, but there are notable quantitative differences. For example, in the Phi35_{UT} case (Fig. 3.4, right column), features that were barely noticeable in the Phi35 integration become more pronounced, like the local minimum in warming in the tropical lower troposphere (Fig. 3.4d), the westward acceleration around the equator (Fig. 3.4e), and the dipole streamfunction anomalies near the surface and near the tropopause at the equator (Fig. 3.4f). These
results, together with those of the LT integrations, suggest that the circulation responses are sensitive more to the horizontal structure of the thermal forcing than to its vertical structure. It is possible, however, that thermal forcings with more complicated vertical structure might produce different results.

We have also performed a set of integrations in which we sweep the relative amplitude of the thermal forcing by varying the factor $\alpha$, defined in Eq. (3.1). One might expect that the responses are linear in $\alpha$, in which case a doubling of the forcing amplitude should double the amount of HC expansion and jet shift. The results shown in Fig. 3.5 are approximately linear, except for the Phi5 integrations at high $\alpha$, which even show some non-monotonicity (Fig. 3.5b, triangles). The responses do not exhibit any jump discontinuity like that shown in Wang et al. (2012), even though the amplitudes of our thermal forcings are comparable. The Phi5 and Phi35 integrations show slight nonlinearity at low $\alpha$, but the circulation responses are very weak in these cases, so much longer integrations would be required to confirm a statistically robust nonlinearity. It is also clear that the Phi35-20 response is well-separated from that of Phi35: even if we reduce the amplitude of the Phi35-20 forcing by half ($\alpha = 0.5$), the response is still greater than the Phi35 response at its default amplitude.

The relatively large circulation response of Phi35-20, detailed above, suggests that there might be a linearly additive relationship between the responses to wide and narrow thermal forcings. To test this more rigorously, we have performed Phi35-20$_{LT}$ and Phi20$_{LT}$ integrations with their forcing amplitudes chosen so that their sum matches the exact amplitude of the Phi35$_{LT}$ forcing. This requires that we set $\alpha = 15/35$ for the Phi35-20$_{LT}$ forcing and $\alpha = 20/35$ for the Phi20$_{LT}$ forcing (see Eqs. 3.1 and 3.3). In this case, we find that Phi35-20$_{LT}$ produces $0.63 \pm 0.05^\circ$ HC expansion, compared to $0.54 \pm 0.06^\circ$ for Phi35$_{LT}$ and $-0.02 \pm 0.02^\circ$ for Phi20$_{LT}$. (Negative values indicate HC contraction.) So the Phi35-20$_{LT}$ response is larger
Figure 3.5: Changes in circulation metrics as functions of relative forcing amplitude, $\alpha$. The circulation metrics are defined in the caption of Fig. 3.3 and in the text.

than the difference of the Phi35$_{LT}$ and Phi20$_{LT}$ responses, but this nonlinearity is not statistically significant.

### 3.3 A Diffusive Model of The Circulation Response

#### 3.3.1 Approach

The key result from our GCM experiments is that the transition from HC contraction to HC expansion is determined primarily by the meridional width of the thermal forcing. We now seek a simplified explanation of this behavior. To begin, we remind ourselves that the HC edge coincides with a downward maximum of the zonal mean vertical velocity, $\bar{\omega}$. So if we wish to determine how the HC edge shifts in response to a particular thermal forcing, then we need to relate $\bar{\omega}$ to the net
diabatic heating, \( \overline{Q} \). Fortunately, these quantities are directly related through the temperature equation, but the temperature equation includes additional contributions, most important of which is the divergence of the meridional eddy heat flux, \( \overline{v'\theta'} \).

Thus the challenge is finding a way to represent the circulation that makes the problem tractable. To this end, we choose to parameterize the total circulation as diffusive, following an approach similar to that of Frierson et al. (2007a) and Kang et al. (2009). This parameterization accounts for transport due to both eddies and the mean flow by assuming that they together act to diffuse heat meridionally. Such an approach greatly simplifies the system, but in the process, it blurs the distinction between eddies and the mean flow. This makes it more appropriate that we work in terms of the transformed Eulerian mean (TEM; Edmon et al., 1980), which combines the Eulerian vertical velocity and eddy heat flux divergence into a single quantity representing the total heat transport. This quantity is called the residual vertical velocity, \( \overline{\omega^*} \), and it is defined as

\[
\overline{\omega^*} \equiv \overline{\omega} + \frac{1}{a \cos \phi} \frac{\partial}{\partial \phi} \left( \overline{v'\theta' \cos \phi} \right),
\]

(3.4)

where \( \overline{\theta_p} \) is the vertical stratification in pressure coordinates, \( \phi \) is latitude, and \( a \) is Earth's radius.

This raises a pivotal question: how do we locate the HC edge in the TEM system? The TEM meridional circulation consists of just one cell extending from the equator to the pole (Edmon et al., 1980), in contrast to the three-cell structure of the Eulerian mean circulation. However, we can still identify the HC from the TEM circulation. This is because, in the upper troposphere, eddy heat fluxes are small enough that there is a close correspondence between \( \overline{\omega^*} \) and \( \overline{\omega} \). As seen in Edmon et al. (1980), Fig. 6a, or Held and Schneider (1999), Fig. 3a, the upper half of the HC is clearly evident in the upper tropospheric portion of the TEM circulation, where the Eulerian mean flow dominates.
We have found that the HC edge can be accurately identified as the latitude where there is a downward maximum of $\bar{\omega}^*$ when averaged over 200-500 hPa; we call this quantity $\bar{\omega}_{\text{max}}^*$. By vertically averaging over the upper troposphere, we ensure that the maximum is robustly located. Most importantly for our purposes, this definition accurately captures changes in HC width due to thermal forcings. Fig. 3.6, circles, shows the shift of $\bar{\omega}_{\text{max}}^*$ from the GCM experiments of Sec. 3.2. Comparing Fig. 3.6 with Fig. 3.3, one sees that the $\bar{\omega}_{\text{max}}^*$ metric and the conventional $\Psi_{500}$ metric agree well with each other; the modest differences that do arise are not substantial enough to affect our key conclusions.

Defining the HC edge in terms of $\bar{\omega}^*$ is a key step because we can obtain a very simple relation between the change in $\bar{\omega}^*$ and the anomalous diabatic heating. This,
combined with our diffusive parameterization of the circulation, allows us to solve for the change in residual vertical velocity, and thus the shift of the HC edge ($\bar{\omega}'_{\text{max}}$). Not surprisingly, this diffusive model has important limitations, which we address below. Nonetheless, the model provides a very simple way of understanding the transition from HC contraction to HC expansion.

### 3.3.2 Mathematical formulation

Having outlined our approach, we now provide the formal details. Our domain is taken to be the arc spanning 0-90° latitude, representing a layer averaged zonally and vertically over the upper troposphere of one hemisphere. (We assume hemispheric symmetry.) In the TEM system, the temperature equation takes the form

$$\frac{\partial \bar{\theta}}{\partial t} + \bar{\theta} \bar{\omega}' = \bar{Q}, \tag{3.5}$$

where $\bar{\theta}$ is the zonal mean potential temperature and $t$ is time. We hereafter refer to $\bar{Q}$ as the “diabatic tendency,” and this term can be positive (diabatic heating) or negative (diabatic cooling). In contrast to the system considered by Held and Hou (1980), Eq. (3.5) neglects horizontal advection by the mean flow, but implicitly includes eddy heat flux divergence.

We assume steady-state conditions and parameterize the diabatic tendency as Newtonian cooling, so Eq. (3.5) becomes

$$\bar{\theta} \bar{\omega}' = -\frac{\bar{\theta} - \bar{\theta}_{eq}}{\tau}, \tag{3.6}$$

where $\bar{\theta}_{eq}$ is the equilibrium potential temperature and $\tau$ is the relaxation timescale. This means that temperature deviations from thermal equilibrium must be balanced by vertical advection. If we were to neglect eddy heat fluxes, Eq. (3.6) would reduce to a form equivalent to that obtained under the weak temperature gradient (WTG) approximation (e.g. Held and Hoskins, 1985; Sobel et al., 2001), as well as other
linear formulations of the tropical circulation (e.g. Schneider and Lindzen, 1976; Gill, 1980; Wang and Li, 1993). We consider this eddy-neglecting limit further below.

Thus our system has two unknowns: \( \bar{\omega}^* \) and \( \bar{\theta} \). To close the system, we parameterize the TEM circulation by assuming that vertical advection acts to diffuse potential temperature meridionally. Specifically,

\[
\bar{\theta}_p \bar{\omega}^* = -\frac{k}{a^2 \cos \phi} \frac{\partial}{\partial \phi} \left( \cos \phi \frac{\partial \bar{\theta}}{\partial \phi} \right),
\]

(3.7)

where \( k \) is the diffusivity, taken to be spatially uniform. We eliminate \( \bar{\omega}^* \) by equating (3.6) and (3.7), obtaining

\[
\bar{Q} - \bar{\theta}_\tau = -\frac{k}{a^2 \cos \phi} \frac{\partial}{\partial \phi} \left( \cos \phi \frac{\partial \bar{\theta}}{\partial \phi} \right),
\]

(3.8)

where \( \bar{Q} \) is the diabatic source term, defined as \( \bar{Q} = \bar{\theta}_{eq}/\tau \). This means that meridional diffusion acts to balance the diabatic tendency. This is analogous to the formulations of Frierson et al. (2007a) and Kang et al. (2009), in which the meridional diffusion of moist static energy acts to balance the net radiative heating.

We now perturb the system with a thermal forcing, \( \bar{Q} \). This in turn produces perturbations of temperature, \( \bar{\theta} \), and residual vertical velocity, \( \bar{\omega}^* \). For simplicity, we assume that the diffusivity and stratification remain fixed. We separate the perturbations from their associated background values, so that

\[
\langle \bar{Q} \rangle = \bar{Q} + \bar{Q}, \quad \langle \bar{\theta} \rangle = \bar{\theta} + \bar{\theta}, \quad \langle \bar{\omega}^* \rangle = \bar{\omega}^* + \bar{\omega}^* ,
\]

(3.9) (3.10) (3.11)

where angle brackets denote final values after the perturbation. Placing these into Eqs. (3.6) and (3.8), we can subtract the background state and obtain equations for
just the perturbation fields:

$$\tilde{Q} - \frac{\tilde{\theta}}{\tau} = -\frac{k}{a^2 \cos \phi} \frac{\partial}{\partial \phi} \left( \cos \phi \frac{\partial \tilde{\theta}}{\partial \phi} \right),$$  \hspace{1cm} (3.12)

$$\tilde{\omega}^* = \frac{1}{\tilde{\theta}_p} \left( \tilde{Q} - \frac{\tilde{\theta}}{\tau} \right).$$  \hspace{1cm} (3.13)

The quantity $\tilde{Q} - \tilde{\theta}/\tau$ represents the anomalous diabatic tendency. Thus in the case of stable stratification ($\tilde{\theta}_p < 0$), anomalous diabatic heating ($\tilde{Q} - \tilde{\theta}/\tau > 0$) is balanced by anomalous TEM ascent ($\tilde{\omega}^* < 0$). Eq. (3.12) is a one-dimensional boundary value problem in $\tilde{\theta}$. The boundary conditions are taken to be $\partial \tilde{\theta}/\partial \phi = 0$ at the equator (by hemispheric symmetry) and $\partial \tilde{\theta}/\partial \phi = 0$ at the pole (to maintain thermal wind balance with zero zonal wind). Once we solve (3.12) for $\tilde{\theta}$, then we can solve (3.13) for $\tilde{\omega}^*$.

Since we are primarily interested in the shift of the HC edge, we use this diffusive model to compute only perturbation fields. (This is not a model for the mean Hadley circulation.) The background state is obtained from output of our GCM control integration; this output is zonally and vertically averaged over 200-500 hPa, and values from both hemispheres are combined to double the sample size. We apply the same averaging scheme when comparing the GCM responses to the results of the diffusive model (see below). The parameters of the diffusive model are chosen as follows: We let $\tilde{\theta}_p = -4 \times 10^{-4}$ K Pa$^{-1}$, which matches the vertical stratification in the upper troposphere of the GCM control integration. Secondly, we find that the temperature response of the diffusive model adequately matches that of the GCM if we let $k = 1 \times 10^6$ m$^2$ s$^{-1}$ and $\tau = 35$ d. This value for $k$ is of the same order as that used in Frierson et al. (2007a) and Kang et al. (2009), and the value for $\tau$ is comparable to other estimates of the thermal equilibrium timescale in the troposphere (Held and Suarez, 1994; Robinson, 2002). The thermal forcings ($\tilde{Q}$) used in the diffusive model are equal to the thermal forcings used in the GCM integrations, vertically averaged over 100-1000 hPa. We average the thermal
forcings over the whole troposphere (rather than just the upper troposphere) to account for the fact that convection spreads the thermal forcing vertically.

3.3.3 Results

Fig. 3.7 shows numerical solutions of the diffusive model. The dashed curves in the top row show the thermal forcings, \( \tilde{Q} \), multiplied by \( \tau \). These represent what the temperature responses would be if there were no changes in the circulation. The e-folding widths of the thermal forcings range from 5° (Phi5) in the leftmost column to 25° (Phi25) on the right. The thick solid curves in the top panels show the calculated temperature responses. By construction, these show a diffusive character: the temperature responses are flattened compared to \( \tilde{Q}\tau \). Thus, there is a transition from anomalous diabatic heating (\( \tilde{Q}\tau > \bar{\theta} \)) in the region of peak thermal forcing to anomalous diabatic cooling (\( \tilde{Q}\tau < \bar{\theta} \)) elsewhere.

The bottom panels of Fig. 3.7 show the responses of the residual vertical velocity. As follows directly from Eq. (3.13), there is anomalous ascent in regions of anomalous diabatic heating (i.e. \( \tilde{\omega}^* < 0 \) for \( \tilde{Q}\tau > \bar{\theta} \)) and anomalous descent in regions of anomalous diabatic cooling (i.e. \( \tilde{\omega}^* > 0 \) for \( \tilde{Q}\tau < \bar{\theta} \)). Thus, there is anomalous descent on the poleward flank of the thermal forcing. The vertical dot-dashed lines in the bottom panels mark the edge of the HC (i.e. \( \tilde{\omega}_{\text{max}}^* \)) calculated from the GCM control integration. The results show that for the Phi5 case (Fig. 3.7d), there is a descending anomaly whose maximum is on the equatorward side of the HC edge, producing contraction of the HC. As the thermal forcing is widened, the peak of this descending anomaly moves to the poleward side of the HC edge (Fig. 3.7e,f), resulting in expansion of the HC. Thus, our simple diffusive model qualitatively reproduces the transition from HC contraction to HC expansion.

For comparison purposes, the thin black lines in Fig. 3.7 show the same fields obtained from the standard GCM integrations. For the temperature responses (top
Figure 3.7: Results from the diffusive model described in Sec. 3.3 for the forcings indicated at the top of each column. Thick solid lines show output from the diffusive model. Thin black lines show output from the standard GCM integrations, shown for comparison. Thick dashed lines show the imposed thermal forcings in units of temperature ($\tilde{Q}_\tau$). The vertical dot-dashed lines in the bottom panels indicate the latitude of the HC edge ($\tilde{\omega}_{\text{max}}$) from the control integration. Note, for clarity the vertical scale of panel (a) is different from the other panels.
row), the main discrepancy is that the GCM responses have less meridional gradient in the low- to midlatitudes when compared to the diffusive model. Better agreement may be achieved by spatially varying the diffusivity, but this would not affect any of the key conclusions drawn from the model. As for the residual vertical velocity (bottom row), the main discrepancy is that the GCM responses show ascending anomalies in the midlatitudes which are completely missing in the diffusive model. Calculating heat budget terms from the GCM (not shown), we find that these ascending anomalies are primarily associated with anomalies of the vertical eddy heat flux \( \bar{\omega}'\bar{\theta}' \), which is neglected in the TEM approximation. Edmon et al. (1980) have also noted the importance of vertical eddy heat fluxes in the midlatitudes. This discrepancy, however, occurs far enough poleward of the HC edge that it does not contribute significantly to the shift of the HC edge, except possibly in the Phi5 case.

Next, as a more quantitative test, we add \( \tilde{\omega}' \) from the diffusive model to the climatological \( \bar{\omega}' \) of the GCM and calculate the resulting shift of the HC edge \( \bar{\omega}_{\text{max}}' \). This is plotted as the black squares in Fig. 3.6. The diffusive model shows close quantitative correspondence with the output of the GCM (red circles), both in terms of the amplitude of HC expansion, as well as the transition from HC contraction to HC expansion. One point of disagreement is that the diffusive model produces about one degree less HC contraction than the GCM for the Phi5 integration. This may be due to the fact that, compared to the diffusive model, the GCM produces more descent just equatorward of the HC edge and more ascent just poleward of the HC edge (Fig. 3.7d). As noted above, the latter anomaly is associated with vertical eddy heat fluxes, which the diffusive model does not capture (Fig. 3.7d).

A bigger discrepancy in Fig. 3.6 is that the diffusive model does not reproduce the much-enhanced HC expansion seen in the Phi35-20 case. Instead, the diffusive
model produces slightly less HC expansion for Phi35-20 (empty black square) than it does for Phi35. Fig. 3.8 shows the output of the diffusive model for the Phi35-20 forcing. In this case, the model produces a broad ascending anomaly that peaks slightly equatorward of the HC edge (Fig. 3.8b, thick line). In contrast, the GCM shows a sharp, spatially confined ascending anomaly on the equatorward flank of the HC edge, and a similarly sharp descending anomaly on its poleward flank (Fig. 3.8b, thin solid line). Further examination of GCM output reveals that this dipole anomaly coincides with a similarly pronounced dipole of anomalous eddy momentum flux convergence/divergence (not shown).

Thus, this discrepancy appears to be due to our model’s inability to capture the effects of eddy momentum fluxes, which cannot be modeled as a simple diffusive process. Eddy momentum fluxes might also be partly responsible for driving the
anomalous vertical eddy heat fluxes associated with other model discrepancies noted above. It is worth noting that for the thermal forcings in Fig. 3.7, the peaks of the thermal forcings (and thus the peaks of the ascending anomalies) are situated at the equator, where the eddy momentum flux is negligible. Meanwhile, in the Phi35-20 case, the entire thermal forcing is situated in a region where normally there is substantial flux of eddy momentum. This may be a crucial aspect of the Phi35-20 forcing that leads to the discrepancy between the diffusive model and the GCM.

Earlier studies have applied the thermal wind balance principle to relate shifts of the midlatitude jet to changes in the meridional temperature gradient (Seager et al., 2003; Lorenz and DeWeaver, 2007; Allen et al., 2012). It is tempting to use our diffusive model to calculate the jet shift from the temperature response, but the model is not suitable for this purpose. This is because the diffusive model produces a temperature response whose meridional gradient lacks important structure. For example, in the Phi15 case (Fig. 3.7b), the diffusive model’s temperature response has its steepest gradient between 10-30° lat, whereas the GCM response is nearly flat through this region and has its steepest gradient farther poleward, between 35-45° lat. This difference is substantial enough that the diffusive model would produce shifts of the midlatitude jet that are highly inaccurate. This shortcoming of the diffusive model is not surprising, since eddy momentum fluxes are believed to play an important role in shifting the midlatitude jet (Seager et al., 2003; Wu et al., 2012; Chen et al., 2012), and our model, as noted above, is incapable of properly capturing them.

As an additional test, we have calculated the shift of the HC edge assuming there is no contribution from the meridional eddy heat flux. Such an assumption, as noted above, is common to linear models of the tropical circulation, and it means that there is no need to distinguish between the residual vertical velocity and the Eulerian vertical velocity (i.e. $\bar{\omega}^* = \bar{\omega}$). If we also assume the same scalings as used
for the TEM equations, then the change in Eulerian vertical velocity, $\tilde{\omega}$, is obtained directly from Eq. (3.13).

In this eddy-neglecting limit, we have used our diffusive model to calculate $\tilde{\omega}$ for each thermal forcing. Adding this change to the climatological $\bar{\omega}$ from the GCM control integration, we have also calculated the shift of the maximum of $\bar{\omega}$, which coincides with the HC edge. In this case (not shown), we obtain a transition from HC contraction to HC expansion at approximately the same value of $\phi_w$, but the actual magnitude of HC expansion is about an order of magnitude lower than that shown in Figs. 3.3a and 3.6. Therefore, to obtain a reasonable amplitude of HC expansion, we cannot assume that eddy heat fluxes are unchanged; changes in eddy heat fluxes appear to be a key contribution. This does not clarify whether the circulation response is actually driven by eddy heat fluxes, as suggested by Butler et al. (2011), rather than eddy momentum fluxes, as argued by others (Seager et al., 2003; Wu et al., 2012; Chen et al., 2012).

In any case, our diffusive model does demonstrate that the circulation response can be understood largely in terms of thermally-driven processes. That is, a positive thermal forcing produces anomalous TEM descent on its poleward flank. If this anomalous descent is located equatorward (poleward) of the HC edge, then the HC contracts (expands).

### 3.4 Discussion

#### 3.4.1 Changes in baroclinicity

Earlier studies have examined the degree to which HC width obeys the scalings suggested by baroclinic instability theory (e.g. Held, 2000; Walker and Schneider, 2006; Frierson et al., 2007b; Lu et al., 2008). Using the baroclinic criticality formulation of Phillips (1954), Lu et al. (2008) showed that a decrease in criticality
is associated with a poleward shift of the HC edge. Phillips' criticality depends on both bulk vertical shear and bulk static stability, but Lu et al. (2008) showed results suggesting that increased static stability is the dominant contributor to HC expansion in coupled GCMs. Lu et al. (2010) arrived at a similar conclusion when varying the SST forcing in an atmosphere-only GCM. These findings are seemingly at odds with our LT integrations, which produce significant HC expansion even when tropical static stability decreases (e.g. Fig. 3.4). We must emphasize, however, that the relevant changes in baroclinicity depend on static stability changes in the subtropics (i.e. on the equatorward flank of the jet), not the tropics.

Thus, to properly compare with earlier findings, we have calculated from our GCM output the change in Phillips' criticality using the same formulations as in Lu et al. (2008). Specifically, we compute the difference in criticality, $\delta C$, between each of our forced integrations and our control integration,

$$\delta C = \delta \left[ -\frac{f^2(u_{500} - u_{850})}{\beta g H(\theta_{500} - \theta_{850})/\Theta_0} \right],$$

where $u$ is the zonal wind, $g$ is the gravitational acceleration, $f$ is the Coriolis parameter, $\beta$ is the meridional gradient of the Coriolis parameter, $H$ is the height scale, $\Theta_0$ is a reference temperature, and the 500 and 850 subscripts indicate the pressure levels, in hPa, where $u$ and $\theta$ are evaluated. This expression is then expanded into contributions due to static stability,

$$\delta C_{st} \approx -\frac{f^2(u_{500} - u_{850})_{ctl} \delta(\theta_{500} - \theta_{850})}{\beta g H(\theta_{500} - \theta_{850})_{ctl}^2/\Theta_0},$$

and vertical shear,

$$\delta C_{sh} = \frac{f^2\delta(u_{500} - u_{850})}{\beta g H(\theta_{500} - \theta_{850})_{ctl}/\Theta_0},$$

where the $ctl$ subscript indicates quantities calculated from the control integration. These expressions indicate that the criticality can be reduced either by increasing static stability or by decreasing vertical shear. To compute these quantities from
Figure 3.9: Relationships between HC widening and (a) change in total Phillips’ criticality, $\delta C$; (b) change in criticality due to bulk vertical shear, $\delta C_{sh}$; and (c) change in criticality due to bulk static stability, $\delta C_{st}$. Red markers indicate standard integrations, blue markers indicate LT integrations, and green markers indicate UT integrations. Results from the Northern and Southern Hemispheres have been averaged together. Note that negative values for $\delta C_{sh}$ indicate decreases in vertical shear, and negative values for $\delta C_{st}$ indicate increases in static stability.

GCM output, we first meridionally average the zonal-mean wind and potential temperature fields over 21-46° latitude (which is the 25° band immediately equatorward of the midlatitude jet of the control integration, following Lu et al., 2008). Then we apply Eqs. (3.14-3.16) with $H = 5$ km, $\Theta_0 = 300$ K, and $f$ and $\beta$ computed at 33.5° (the midpoint of the latitude band).

We present the results of these calculations in Fig. 3.9. Specifically, Fig. 3.9a shows the change in HC width versus the change in total criticality, $\delta C$. This shows that, in agreement with earlier studies, decreases (increases) in criticality are generally associated with HC expansion (contraction). Fig. 3.9b plots HC widening versus $\delta C_{sh}$. This exhibits a pattern similar to that of Fig. 3.9a, although the data are shifted farther from the origin: several integrations show increases in $\delta C_{sh}$ associated with HC expansion. Fig. 3.9c shows HC widening versus $\delta C_{st}$, and the results here are widely scattered, with the LT integrations (blue markers) even showing a positive correlation between $\delta C_{st}$ and HC width.
Thus our results disagree with those of Lu et al. (2008, 2010): changes in vertical shear—not static stability—appear to be the dominant contributor to HC expansion in our model. This contrast may be due to the fact that our model is dry, and thus changes in static stability are not constrained in the same way as in moist models. Another possible explanation is that Lu et al. (2008, 2010) consider a more narrow range of forcings than we do, and that a different selection of forcings in comprehensive models might produce HC expansion with a more significant vertical shear contribution.

3.4.2 Jet position vs. Hadley Cell edge

Earlier studies (e.g. Fu et al., 2006; Seidel et al., 2008; Fu and Lin, 2011; Davis and Rosenlof, 2012) have used the position of the jet to examine the widening trend of the tropics. Our results suggest that using a metric based on jet latitude rather than HC edge can give a different impression of how the width of the tropical belt is changing. Fig. 3.3 shows that the shift of the HC edge and the shift of the jet can be quite different for the same thermal forcing. If one is more interested in the location of the dry zones, which is closely related to the location of the HC edge, then relying on a jet latitude metric would be somewhat misleading.

This difference between jet latitude and HC edge may relate to the fact that the subtropical jet and the midlatitude eddy-driven jet can separate from each other. The precise drivers of this jet separation remain unclear. Lu et al. (2008) took an initial step by showing that in coupled model simulations of global warming, the poleward shift of the SH midlatitude jet is about twice the shift of the HC edge. This result agrees with our global warming–like integrations (Fig. 3.3 for large $\phi_w$) but not with our El Niño–like integrations (Fig. 3.3 for small $\phi_w$). To further complicate matters, Kang and Polvani (2011) showed that in coupled models, there is no correlation between HC edge and jet latitude during winter in SH and during
all seasons in NH. Thus, many questions remain in this area.

3.4.3 Warming in the upper vs. lower troposphere

The results of Figs. 3.3 and 3.4 suggest that our lapse-rate perturbation has relatively little effect on the circulation response. This does not mean that warming in the upper troposphere is less important than warming in the lower troposphere. Note that for the Phi35 LT integration (Fig. 3.4), even though the thermal forcing is confined to the lower troposphere, there is still significant warming in the upper troposphere. We have also performed an integration in which the thermal forcing is more strictly confined to the upper troposphere between $-35^\circ$ and $35^\circ$ lat (not shown). We accomplish this by increasing the amplitude of the lapse-rate perturbation and reducing the amplitude of the thermal forcing in the lower troposphere. The associated temperature response is comparable to the upper tropospheric response of Phi35 (Fig. 3.2d), but there is much less warming in the lower troposphere. Despite this change in the vertical structure of the warming, the resulting HC expansion and poleward shift of the jet is nearly equal to that of Phi35. This gives further support to our earlier finding: there is relatively little sensitivity to the change in the lapse rate, and there is much greater sensitivity to the meridional structure of the thermal forcing.

There is, however, a caveat to this claim: a narrow thermal forcing confined to the tropical upper troposphere produces a response that is not completely El Niño–like. In this case, the HC contracts slightly, but the jets shift poleward. Thus, warming in the tropical lower troposphere appears to be essential for producing an El Niño–like circulation response. The reasons for this sensitivity are unclear, but we would argue that such a thermal forcing is highly unrealistic. Specifically, warming in the tropical upper troposphere would typically require some warming in the tropical lower troposphere as well, especially in the case of El Niño, where
there is substantial warming at the surface.

In the context of global warming, however, our results suggest that the lapse-rate feedback is not as consequential for the tropospheric circulation as earlier studies hypothesize (Butler et al., 2010, 2011; Wang et al., 2012). We obtain much the same circulation response whether peak warming occurs in the upper troposphere or the lower troposphere.

3.4.4 Implications for recent observations

While the results of our Phi35 integration resemble the global warming response of comprehensive models, it is not certain that this accurately represents the trends observed over recent decades. Satellite observing systems have been used to study the trend in vertically-averaged tropospheric temperatures, and these suggest that recent warming has been maximum over the Arctic and in the midlatitudes (Santer et al., 2003; Fu et al., 2006; Karl et al., 2006; Santer et al., 2012). In the vertical average, our Phi35 integration produces maximum warming in the tropics, and our Phi35-20 integration produces peak warming in the midlatitudes. So the warming pattern in satellite observations most closely resembles that of our Phi35-20 integration, which, as shown above, produces much more HC widening than the Phi35 integration. If the warming pattern in satellite observations is correct, this suggests that the HC might be widening at a rate much faster than in typical simulations of global warming.

There is, however, much uncertainty surrounding satellite observations, due to numerous changes in processing software, the appearance of cooling trends in some datasets, and some trends that appeared to contradict the lapse-rate feedback principle (for extensive discussions, see Karl et al., 2006; Santer et al., 2008; Thorne et al., 2011). While there are quantitative differences between datasets, they all do show enhanced warming in the NH midlatitudes, with some datasets also show-
ing mildly enhanced warming in the SH midlatitudes [e.g. see Karl et al. (2006), Fig. 3.5, and Santer et al. (2012), Fig. S5]. This does not prove that midlatitude amplification is a reality, but it does compel us to consider it as a serious possibility.

In simulations with historical forcings, some comprehensive models do produce midlatitude amplification, but most do not (Santer et al., 2012). The multimodel mean exhibits peak warming that is approximately flat equatorward of \( \sim 30^\circ \), while observations show a more pronounced local maximum in warming near 30° (Santer et al., 2012, Fig. S5). This lack of midlatitude warming might cause comprehensive models to underpredict rates of tropical widening. Johanson and Fu (2009) and Allen et al. (2012) have shown results suggesting that comprehensive GCMs have indeed underproduced historical tropical widening trends, but Davis and Rosenlof (2012) offer evidence that the observed widening trend may not be robust. So further monitoring and further analysis are needed to determine if there is a real discrepancy between models and observations. But our key point remains: midlatitude amplification is a very real possibility, and it might greatly enhance the rate of tropical widening.

What might be causing enhanced midlatitude warming in the first place? Allen et al. (2012) proposed that such warming may be due to tropospheric ozone or absorbing aerosols, which are more spatially confined than carbon dioxide. Another possibility is that changes in subtropical humidity and cloud cover are contributing to this pattern. These changes may in turn be related to changes in ocean temperatures. For example, Hoerling and Kumar (2003) have shown that, on multi-year timescales, cooling in the eastern tropical Pacific can lead to enhanced warming in the midlatitudes. It is left to future studies to pinpoint the possible drivers of midlatitude warming more conclusively.
3.5 Summary and Conclusion

Using an idealized GCM, we have shown that the contrast between the El Niño and global warming circulation responses depends on the meridional structure of the thermal forcing. A narrow positive forcing centered at the equator produces HC contraction and an equatorward shift of the jets, while a wider forcing has the opposite effect. Furthermore, warming concentrated in the midlatitudes produces much-amplified HC expansion and poleward jet shifts when compared to a thermal forcing that is spread over the tropics. These responses are primarily sensitive to changes in the meridional structure of the thermal forcing and are less sensitive to changes in the lapse rate. The exceptionally large circulation response to midlatitude warming points to the possibility that comprehensive GCMs might underpredict widening of the tropical belt.

We have also provided a simplified way of understanding these circulation responses. Specifically, we can parameterize the TEM circulation as the meridional diffusion of potential temperature. When a thermal forcing is applied, it results in anomalous diabatic cooling, and hence anomalous TEM descent, on the poleward flank of the thermal forcing. For a narrow (wide) thermal forcing, this anomalous descent occurs on the equatorward (poleward) side of the HC edge, producing an equatorward (poleward) shift of the HC edge.

One area ripe for future study concerns the possible causes of amplified warming in the midlatitudes. Possible contributors include absorbing aerosols (Allen et al., 2012) or long-term changes in tropical SSTs (Hoerling and Kumar, 2003). Experiments with full and intermediate-complexity GCMs will be key to testing various hypotheses. Finally, every effort should be made to determine the robustness of the midlatitude amplification patterns shown in satellite observations.
3.6 Appendix: GCM Description

Many aspects of the model we use are identical to those of Chapter 2, but we provide here the essential details. We use the spectral dynamical core of the GFDL Flexible Modeling System (FMS). The horizontal truncation is T42 for all results presented in the study, but we have also tested T85 and found no notable differences. The vertical level interfaces, in sigma coordinates, are \( \sigma_i = (i/L)^2 \), \( i = 0, 1, 2, \ldots, L \), where \( L \) is an integer. For all results presented in this study \( L = 40 \), but we have also tested \( L = 80 \) and found no notable differences.

We add terms to the temperature equation to capture convective and radiative processes, as well as our imposed thermal forcing. Specifically,

\[
\frac{\partial T}{\partial t} = \ldots - \frac{T - T_C}{\tau_C} - \frac{T - T_R}{\tau_R} + \tilde{Q} \left( \frac{p}{p_0} \right)^{R/c_p},
\]

where \( T_C \) and \( \tau_C \) are the convective equilibrium temperature and timescale, respectively; \( T_R \) and \( \tau_R \) are the radiative equilibrium temperature and timescale, respectively; \( \tilde{Q} \) is our external thermal forcing in terms of potential temperature, given by Eq. (3.1); \( R \) is the gas constant for dry air; and \( c_p \) is the specific heat of dry air. \( T_R \) and \( \tau_R \) are exactly as given in Chapter 2, mimicking the thermal structure of an atmosphere in gray radiative equilibrium.

\( T_C \) is given by

\[
T_C(\lambda, \phi, p, t) = \begin{cases} 
T_m(\lambda, \phi, p, t) - E_C(\lambda, \phi, t) & p_{LNB}(\lambda, \phi, t) \leq p \leq p_0 \\
T(\lambda, \phi, p, t) & p < p_{LNB}(\lambda, \phi, t),
\end{cases}
\]

where

\[
E_C(\lambda, \phi, t) = \frac{1}{p_{LNB}(\lambda, \phi, t) - p_0} \int_{p_0}^{p_{LNB}(\lambda, \phi, t)} [T_m(\lambda, \phi, p', t) - T(\lambda, \phi, p', t)] dp'
\]

ensures conservation of enthalpy in (3.18). Eq. (3.18) is applicable only when \( E_C > 0 \). If \( E_C \leq 0 \) then convection is inhibited, i.e. \( T_C = T \) in the entire column. \( T_m \) is
the moist adiabat,

\[
T_m(\lambda, \phi, p, t) = T_s(\lambda, \phi, t) \left( \frac{p}{p_0} \right)^{R(\Gamma_m + \tilde{\Gamma})/g} + \Delta_z \log \frac{p}{p_0},
\]

(3.20)

where \( T_s \) is the surface temperature at longitude-latitude-time \((\lambda, \phi, t)\); \( \Gamma_m = 6 \text{ K km}^{-1} \); \( \tilde{\Gamma} \) is the lapse rate perturbation given by Eq. (3.2); \( \Delta_z = 7 \text{ K} \); and \( p_{\text{LNB}} \) is the level of neutral buoyancy for ascent from the surface along \( T_m \). In contrast to Schneider and Walker (2006) and Chapter 2 of this thesis, Eq. (3.20) includes a second term which makes the lapse rate increase with altitude. This produces more realistic alignment between the upper- and lower-level wind maxima. The timescale \( \tau_C \) is set to 4 hours.

There is no topography in this model. For \( \sigma > 0.7 \), winds are linearly damped as in Held and Suarez (1994). We apply \( \nabla^6 \) hyperviscosity, and above 5 hPa, we apply a sponge layer top with the same functional form as in Polvani and Kushner (2002).
Chapter 4

The Climate Response to Small Perturbations in Total Solar Irradiance: The Importance of the Background State

4.1 Introduction

Changes in total solar irradiance (TSI) are a natural driver of climate change and variability. TSI varies on a range of timescales: there is the annual cycle of the distance between the Earth and the Sun; there is the 11-year cycle of solar activity; and there are less frequent changes, such as the exceptionally low solar activity during the Maunder Minimum. The 11-year solar cycle takes on particular importance because it may significantly contribute to climate variations on decadal timescales, and thus complicate our ability to attribute such changes to human activities and other sources. While the total radiative forcing applied by the solar cycle is small (only ~0.2 W m\(^{-2}\) for a 1 W m\(^{-2}\) increase in the solar constant), regional feedbacks
may act to produce climate variations that are appreciable.

Numerous studies have examined the effect of the 11-year solar cycle on all aspects of the climate system. (See Gray et al., 2010, for an extensive review.) Several studies have found variations in atmospheric temperature associated with the solar cycle, but they disagree on the spatial patterns of these variations (e.g. Scaife et al., 2000; Frame and Gray, 2009). Some have argued that the solar cycle drives changes in the annular modes of both hemispheres (e.g. Kodera, 2002; Kuroda and Kodera, 2005), while others have found no such signal (Moore et al., 2006).

The solar cycle has also been hypothesized to influence the phase of ENSO (e.g. Mann et al., 2005; Meehl and Arblaster, 2009), although other studies claim that no simple relationship exists (Roy and Haigh, 2012; Haam and Tung, 2012).

Thus more work is needed to elucidate the influence of solar forcing on climate. To this end, we have performed a set of integrations in which we cleanly isolate the effect of small changes in TSI, without including any confounding factors such as volcanic forcings or changes in orbital parameters. Furthermore, we perform long integrations to ensure that our results are statistically significant. The central finding of this study is that the response to small TSI perturbations is highly dependent on the background climate state. This may help to clarify why previous studies reach conflicting conclusions: a solar forcing applied to two different background states can produce responses that are, in many respects, qualitatively opposite to each other.

4.2 Method

We perform all of our integrations using the Community Atmosphere Model (CAM) version 3.0, from the National Center for Atmospheric Research (NCAR). CAM3 is an atmospheric general circulation model that includes comprehensive schemes for dynamics, radiation, convection, and clouds. The model has spectral resolution
T42 in the horizontal, with 26 vertical levels. 13 of these levels are at pressures less than 200 hPa, and the model top is at 2.917 hPa. There is no interactive chemistry in this model, and monthly mean aerosol and ozone climatologies are prescribed. For a full description of CAM3, the reader is referred to Collins et al. (2004).

For most of our integrations, we use a configuration in which CAM3 is coupled to a mixed layer ocean that includes a thermodynamic sea ice model. The depth of this mixed layer is \( \sim 50 \) m over most of the globe, but deeper values (\( \sim 160 \) m) are prescribed in the North Atlantic and the Southern Ocean. Heat transports within this mixed layer are simulated with a prescribed “Q flux” that is tuned to produce an SST climatology close to observations, assuming the forcing from the atmosphere is not changed. Such a configuration does not allow for changes in the ocean circulation, which as noted in Shindell et al. (2003), may influence the results in regions like the North Atlantic and the Southern Ocean, where deep water is formed. Clement et al. (2011) have shown, however, that such a slab ocean configuration does produce a Southern Oscillation (SO), although this SO has a different spatial structure and interannual variability from that of coupled models and observations.

For several integrations, we use a different “AMIP style” configuration in which SSTs and sea ice are prescribed. In both the mixed layer and AMIP style configurations, CAM3 includes the annual cycle of solar irradiance, but does not by default include any other form of solar variability. Orbital parameters are fixed at 1950 values.

Each of our integrations differs only in the choice of solar constant, CO\(_2\) mixing ratio, and the lower boundary condition. These details are provided in Table 4.1. The REF integration has present-day values for the solar constant and CO\(_2\), while in the REF(2x) integration the level of CO\(_2\) is doubled. Associated with each of these reference integrations are forced integrations in which we increase the solar constant
Table 4.1: Configurations for the integrations performed in this study. Horizontal lines separate the integrations into groups for which the associated background state is given by the reference integration, indicated in bold.

<table>
<thead>
<tr>
<th>Integration</th>
<th>Solar constant [W m(^{-2})]</th>
<th>CO(_2) mixing ratio [ppm]</th>
<th>Lower boundary condition</th>
</tr>
</thead>
<tbody>
<tr>
<td>REF</td>
<td>1367</td>
<td>355</td>
<td>mixed layer ocean</td>
</tr>
<tr>
<td>0.1%S</td>
<td>1368</td>
<td>355</td>
<td>mixed layer ocean</td>
</tr>
<tr>
<td>2%S</td>
<td>1394</td>
<td>355</td>
<td>mixed layer ocean</td>
</tr>
<tr>
<td>8%CO2</td>
<td>1367</td>
<td>383</td>
<td>mixed layer ocean</td>
</tr>
<tr>
<td>REF(2x)</td>
<td>1367</td>
<td>710</td>
<td>mixed layer ocean</td>
</tr>
<tr>
<td>0.1%S(2x)</td>
<td>1368</td>
<td>710</td>
<td>mixed layer ocean</td>
</tr>
<tr>
<td>2%S(2x)</td>
<td>1394</td>
<td>710</td>
<td>mixed layer ocean</td>
</tr>
<tr>
<td>8%CO2(2x)</td>
<td>1367</td>
<td>767</td>
<td>mixed layer ocean</td>
</tr>
<tr>
<td>REF(s)</td>
<td>1367</td>
<td>355</td>
<td>SSTs from REF integration</td>
</tr>
<tr>
<td>0.1%S(s)</td>
<td>1367</td>
<td>355</td>
<td>SSTs from 0.1%S integration</td>
</tr>
<tr>
<td>REF(δs)</td>
<td>1367</td>
<td>355</td>
<td>SSTs from REF, plus 2.18 K</td>
</tr>
<tr>
<td>0.1%S(δs)</td>
<td>1367</td>
<td>355</td>
<td>SSTs from 0.1%S, plus 2.18 K</td>
</tr>
</tbody>
</table>

by 1 W m\(^{-2}\), which is a 0.1% increase. We label these forced integrations 0.1%S and 0.1%S(2x), and below we will examine the “0.1%S – REF” and “0.1%S(2x) – REF(2x)” responses, obtained by taking the differences between the climatologies of these integrations. These four integrations will be the primary focus of this study.

We have performed additional tests that address specific questions raised by these integrations, and we will motivate these below in the appropriate context.

Each integration lasts 200 years, with the exception of the 2%S and 2%S(2x) integrations which last 100 years. Given the strong internal variability in these models, we have performed long integrations to ensure that the responses to small external forcings are statistically significant. As noted by Wu et al. (2012), this model takes approximately 20 years to equilibrate to an external forcing, so we discard the first 20 years of each integration and use the rest for our analysis.

To calculate statistical significance of the responses, we perform two-tailed t-tests on the timeseries of the forced and reference integrations. These t-tests are performed at each latitude/longitude for surface fields and at each latitude/pressure
level for zonally averaged fields. For surface temperature (ST) and air temperature, we have found that the monthly timeseries are autocorrelated, with high correlation at one month lag. However, these fields are not autocorrelated at lag 2, so to ensure our samples are independent, we take bi-monthly averages of the ST and air temperature timeseries before performing significance tests. (While this bi-monthly averaging does reduce the temporal noise in each timeseries, this is outweighed by the fact that the degrees of freedom are greatly reduced. Thus, this procedure does not produce an artificial increase in statistical significance.) For all other fields examined in this study, the monthly timeseries are not autocorrelated, so we perform significance tests on the entire monthly timeseries.

4.3 Results

First we consider the zonal mean response to a 0.1% increase in the solar constant on the present day climate, captured by taking the difference between the time averaged fields of the 0.1%S and REF integrations (Fig. 4.1, left column). The temperature response (Fig. 4.1a) shows warming throughout the tropical troposphere, enhanced in the tropical upper troposphere. In SH there is negligible warming in the midlatitudes. The zonal wind response (Fig. 4.1b) shows that there is eastward acceleration on the equatorward flank of the midlatitude jet, accompanied by deceleration on its poleward flank. This indicates an equatorward shift of the SH midlatitude jet. The dots in both panels indicate that these features are all statistically significant at the 90% confidence level, based on a two-tailed t-test. (See Sec. 4.2 for additional details on our significance tests.)

Next, we consider the response to a 0.1% increase in the solar constant applied to a doubled CO$_2$ climate. This is captured by differencing the 0.1%S(2x) and REF(2x) integrations (Fig. 4.1, right column). The temperature response (Fig. 4.1c) shows warming throughout the tropical troposphere, similar to that of the 0.1%S
Figure 4.1: Color shading shows (left) the difference between the climatologies of the 0.1%S and REF integrations and (right) the difference between the climatologies of the 0.1%S(2x) and REF(2x) integrations. (See Section 4.2 and Table 4.1 for details.) Gray contours show the climatology of (left) the REF integration and (right) the REF(2x) integration, with contour intervals of (top) 10 K for temperature and (bottom) 5 m s$^{-1}$ for zonal wind with negative contours dashed. The thick black contour is the background climate’s thermal tropopause (cf. World Meteorological Organization, 1957). Black dots mark responses that are statistically significant above the 90% confidence level, based on a two-tailed t-test at each latitude/pressure level. (See Sec. 4.2 for additional details on our significance tests.)
response (Fig. 4.1a). But in the SH extratropics the two responses are opposite to each other, with 0.1%S(2x) producing maximum warming in the midlatitudes, where 0.1%S produces minimum warming. The zonal wind response of 0.1%S(2x) (Fig. 4.1d) shows that the SH jet shifts poleward, a response that is opposite to that in Fig. 4.1b. Although the enhanced SH midlatitude warming in Fig. 4.1c does not pass our statistical significance test, the SH zonal wind anomalies in Fig. 4.1d, which are statistically significant, suggest that such a temperature response is reasonable. We emphasize that the responses in both columns of Fig. 4.1 are due to perturbations that are exactly the same, except that they are applied to two different background states. This suggests that the response to a small solar perturbation is highly dependent on the background state.

Fig. 4.2 shows the surface climate response to 0.1% TSI perturbations. Again, it is clear that the responses are highly dependent on the background state. The ST response in the 0.1%S case (Fig. 4.2a) shows cooling anomalies over the North Sea and South Atlantic, while the 0.1%S(2x) response (Fig. 4.2b) shows warming in these regions. Such contrasts are also apparent in the South Pacific. The globally averaged ST increase in the 0.1%S case is 0.10 K, compared to 0.06 K in the 0.1%S(2x) case, suggesting that the strengths of climate feedbacks are also affected by the background state.

The sea level pressure (SLP) responses in the two cases (Fig. 4.2, second row) also stand in sharp contrast to each other. The 0.1%S (Fig. 4.2c) and 0.1%S(2x) (Fig. 4.2d) responses exhibit annular mode–like patterns in SH that are opposite to each other, reflecting the opposite-signed responses in the zonal wind field (Fig. 4.1b,d). The responses in NH also exhibit marked contrasts, but with much more zonal asymmetry. Most notably, 0.1%S produces a dipole anomaly over the North Pacific and Canada (Fig. 4.2c) that is opposite in sign to that of 0.1%S(2x) (Fig. 4.2d).
Figure 4.2: Color shading shows (left) the difference between the climatologies of the 0.1%S and REF integrations and (right) the difference between the climatologies of the 0.1%S(2x) and REF(2x) integrations for (top row) surface temperature, (middle row) sea level pressure, and (bottom row) precipitation. Black dots mark responses that are statistically significant above the 90% confidence level, based on a two-tailed t-test at each latitude/longitude. (See Sec. 4.2 for additional details on our significance tests.)
Numerous contrasts also arise in the precipitation responses (Fig. 4.2, bottom row). Specifically, 0.1%S produces precipitation increases over the central Pacific and off southern Australia, with decreases over the North Sea and along the Antarctica coast. 0.1%S(2x) produces opposite-signed responses in these locations.

We have found that these opposite-signed responses are also apparent in the timeseries of the SH eddy-driven jet latitude (Fig. 4.3a). Here, the jet is located by finding the latitude of maximum zonal wind at the lowest model level. For clear presentation, we have applied a ten-year running average to all timeseries. The 0.1%S timeseries (red) is mostly above the REF timeseries (black) indicating an equatorward shift. Meanwhile, the 0.1%S(2x) (green) timeseries is mostly below the REF(2x) (blue) timeseries, indicating a poleward shift. We reach similar findings when considering the SH Hadley cell edge timeseries (Fig. 4.3b), although the separation between the forced and reference timeseries is much less than for SH jet latitude. Here, we have located the HC edge by finding the first zero crossing of the meridional mass streamfunction poleward of its subtropical maximum at 500 hPa.

4.4 Discussion

4.4.1 Comparison with past studies

There are numerous ways in which our results compare and contrast with earlier studies. Frame and Gray (2009) and Zhou and Tung (2013) regress the solar cycle onto atmospheric temperature from reanalysis. They find that peaks in the solar cycle correlate with peak warming in the midlatitudes of both hemispheres. Our 0.1%S(2x) integration produces midlatitude warming in SH (Fig. 4.1c), but otherwise, we do not obtain such patterns.

Roy and Haigh (2010, 2012) examine the influence of the solar cycle on SST and SLP from observational data. Roy and Haigh (2010) find SST signals in the South-
Figure 4.3: Timeseries of (a) latitude of the SH midlatitude jet, (b) SH Hadley cell edge, (c) zonal SST gradient in the equatorial Pacific, and (d) zonal SLP gradient in the equatorial Pacific for the (black) REF, (red) 0.1%S, (blue) REF(2x), and (green) 0.1%S(2x) integrations. A 10-year running average has been applied to all timeseries. Horizontal lines indicate climatological values. See the text for precise definitions of each metric.
ern Ocean and North Sea resembling those of our 0.1%S integration (Fig. 4.2a), although these signals are not statistically significant in their analysis. Roy and Haigh (2012) find a statistically significant increase in SLP over the North Pacific, resembling that of our 0.1%S(2x) integration (Fig. 4.2d). They also find a decrease in SLP over Antarctica, but it is not statistically significant. So the overall comparison with Roy and Haigh (2010, 2012) is far from clean. Some features of their results resemble our 0.1%S response, while others resemble 0.1%S(2x). We draw similar conclusions when comparing our results to those of van Loon et al. (2007), although they examine just the North and Central Pacific.

Similar complications arise when attempting a comparison with other studies. Zhou and Tung (2010) and Camp and Tung (2007), using a “composite-mean difference” approach, derive solar signals in observed and reconstructed SST that are very different from the patterns we obtain. When Zhou and Tung (2010) and Meehl et al. (2009) use the alternative approach of averaging just solar peak years and subtracting the climatology, they obtain SST patterns that have some resemblance to our 0.1%S(2x) response, including warming over the North Pacific and North Atlantic. But Zhou and Tung (2010) and Meehl et al. (2009) show cooling in the eastern tropical Pacific that is not apparent in our results. We obtain similar findings when comparing our results to those of Bal et al. (2011), who use a fully coupled model with an idealized sinusoidal forcing function. As for precipitation, the results in Meehl et al. (2009) show patterns very different from our results.

One might suspect that these discrepancies arise because our model has unrealistic ENSO variability and does not allow for changes in the ocean circulation. However, Shindell et al. (2006), using a fully coupled model, impose a step-function increase in TSI and obtain SST and precipitation responses much more similar to our 0.1%S integration. Thus, it appears that the choice of forcing function may play an important role in the response. Another crucial difference may be that the
signals obtained by us and Shindell et al. (2006) are steady-state responses, whereas the signals in Meehl et al. (2009) and Bal et al. (2011) are transient responses, obtained by examining just years with peak solar activity. Some discrepancies might also arise because we consider the annual average response, while other studies focus on the seasonally averaged response.

Liu et al. (2013) examine the solar signal obtained by comparing the climates of the Medieval Warm Period and the Little Ice Age in a millenium run of a fully coupled model. They find an SST signal that agrees mostly with that of our 0.1%S(2x) integration (Fig. 4.2b), although they do not obtain cooling in the northeast Pacific. Some differences might be expected since the solar variations in the millenium runs also include some changes in orbital parameters (Schmidt et al., 2011). Liu et al. (2013) also find precipitation patterns that agree with our 0.1%S(2x) integration (Fig. 4.2f), including a decrease in the central Pacific, flanked by increases to the north and south, along with increases over the Indian Ocean, Australia, and South America. This agreement is encouraging, though somewhat intriguing since the background state in Liu et al. (2013) is far from a doubled CO₂ climate.

Shindell et al. (2001, 2003) perform experiments with a slab ocean model more comparable to ours, and they obtain SST and SLP signals in NH that agree well with our 0.1%S integration. Rind et al. (2008) also perform simulations with a slab ocean model, but the comparison with our results is less clean, perhaps because they consider the difference between solar maximum and solar minimum years, rather than the steady state response. Several studies (Haigh, 1999; Shindell et al., 1999; Matthes et al., 2006) have also analyzed the response to solar forcing from the stratospheric perspective by using models in which SSTs are held fixed. These experiments do not compare cleanly with our results because they do not include the full tropospheric response.

Thus the overall conclusion is that our findings appear to agree with modeling
studies that examine the response to long term changes in TSI. There is much less agreement with studies that examine peak years in the 11-year solar cycle. Despite these discrepancies, our finding that the climate responses depend on the background state is supported by earlier studies. Specifically, Meehl et al. (2003) showed that the regional climate response due just to solar forcing can be quite different from a solar signal that is obtained as a residual from simulations that also include greenhouse gas forcing. Furthermore, Roy and Haigh (2012) and Zhou and Tung (2010) found that calculating solar cycle–induced anomalies with respect to different climatologies can produce very different results. What our results demonstrate is that this dependence on the background state is not merely a result of a data record that is short or irregular: it is physically real and meaningful.

4.4.2 The role of the tropical Pacific

A number of studies have looked for a possible influence of the solar cycle on ENSO, and have arrived at different conclusions (Mann et al., 2005; Meehl and Arblaster, 2009; Roy and Haigh, 2012; Haam and Tung, 2012). Emilie-Geay et al. (2007) have gone further and suggested that the response to solar forcing is actually communicated through its influence on ENSO. The results of our integrations are not clearly El Niño–like or La Niña–like (cf. Seager et al., 2003; Lu et al., 2008), but it is possible that longer-term changes in the tropics play a key role in determining the climate response. Roy and Haigh (2012) hypothesize that certain features of the background state, like the strength of the Walker circulation and the shallow meridional overturning circulation may influence the response to solar forcing.

We have performed some additional analysis to see if the tropics may be driving the responses to solar forcing. Fig. 4.3c shows the timeseries of the zonal SST gradient, calculated by averaging the SST over the western equatorial Pacific and subtracting the SST averaged over the eastern equatorial Pacific, adopting the same
averaging bins used in Karnauskas et al. (2009). The REF(2x) (blue) integration shows a significantly weaker SST gradient than the REF (black) integration, in agreement with the response of coupled models to increased CO$_2$ (DiNezio et al., 2009). However, there is no significant change in the SST gradient in response to solar forcing: the 0.1%S integration (red) shows no clear separation from the REF integration, nor does 0.1%S(2x) (green) show separation from REF(2x).

We reach similar conclusions when considering the zonal SLP gradient (Fig. 4.3d), which we have also calculated in the same way as Karnauskas et al. (2009): The REF and REF(2x) results are well separated, showing a weakening of the Walker circulation with increased CO$_2$, but the response to solar forcing is much less significant. 0.1%S does exhibit an appreciably weaker SLP gradient than in REF, but 0.1%S(2x) shows no substantial difference from REF(2x). These results support the hypothesis of Roy and Haigh (2012) that the background state in the equatorial Pacific may play a role in the response to solar forcing. But it appears unlikely that the response to solar forcing is actually driven by changes in the tropical Pacific. It is worth noting that the coupled models examined in Shindell et al. (2006) and Liu et al. (2013) also do not show substantial changes in zonal SST gradients, so our use of a slab ocean model does not appear to bias this result.

4.4.3 Nonlinearity of the responses

Earlier studies have attempted to investigate the response to solar forcing by increasing the amplitude of the forcing (Wetherald and Manabe, 1975; Hansen and Takahashi, 1984; Haigh, 1999; Cai and Tung, 2012). These studies typically consider the response to a 2% increase in the solar constant, which produces a surface temperature response comparable to that due to doubling CO$_2$. This amplified forcing approach has an advantage in that a statistically significant signal is obtained with less computing time. However, we have performed additional experiments that
Figure 4.4: Shifts of the SH midlatitude jet in response to the forcings indicated along the horizontal axis. Shifts are calculated by taking the jet latitude of the forced integration and subtracting the climatological jet latitude of its corresponding reference integration (see Table 4.1). Error bars indicate the standard deviation of the jet shifts calculated for multiple “ensemble members” obtained by splitting each timeseries into 20-year slices.

demonstrate that the responses to small TSI perturbations do not scale linearly with the responses to large perturbations.

Fig. 4.4 shows the shifts of the SH midlatitude jet for the various integrations performed in this study. For each forced integration, we locate the climatological latitude of the SH jet, and then we subtract the climatological latitude of the appropriate reference integration (see Table 4.1). To provide an error estimate, we separate each timeseries into nine 20-year slices, and obtain the standard deviation of the jet shifts calculated from these nine slices. This is comparable to calculating the intraensemble standard deviation for an ensemble of model integrations.

As shown in Fig. 4.1, solar forcing produces a clear shift of the SH jet in the zonal average, so the SH jet shift provides a convenient metric with which to determine whether the responses to different forcings have the same sign. Fig. 4.4
shows, as was apparent in Fig. 4.1, that the SH jet shifts in opposite directions in response to the 0.1%S and 0.1%S(2x) forcings. We have also performed the 2%S integration, in which a 2% increase in the solar constant is applied to a present day background state, and the 2%S(2x) integration, in which the same forcing is applied to a doubled CO₂ background state. Fig. 4.4 shows that both 2%S and 2%S(2x) produce poleward shifts of the SH jet, in contrast to the opposite-signed shifts of 0.1%S and 0.1%S(2x). Thus the response to solar forcing is nonlinear: small perturbations in TSI may produce a climate response that is very different from the response to large forcings.

One question this raises is whether such nonlinearity is unique to solar forcing, or does it also apply to other forcings, like changes in CO₂? To test this, we have performed experiments in which 8% CO₂ perturbations are applied to present-day and doubled-CO₂ background states. We label these integrations 8%CO₂ and 8%CO₂(2x). (Note that the longwave optical depth is approximately logarithmic in the CO₂ mixing ratio. So to produce temperature responses that are comparable, the fractional increases in CO₂ mixing ratio should be the same for the 8%CO₂ and 8%CO₂(2x) perturbations.) Fig. 4.4 shows that, indeed, 8%CO₂ and 8%CO₂(2x) produce opposite-signed responses, although the equatorward shift for 8%CO₂ is not statistically significant. That said, the 8%CO₂ and 8%CO₂(2x) responses are well separated, indicating that the background state also plays a significant role in the response to small CO₂ perturbations. Thus our results have implications beyond just consideration of the 11-year solar cycle. Specifically, the signature of anthropogenic climate change on decadal timescales may be very different from its signature on centennial timescales.
4.4.4 The role of SSTs

As mentioned above, some earlier studies have examined the response to solar forcing by focusing on just the stratospherically driven response (Haigh, 1999; Balachandran et al., 1999; Shindell et al., 1999; Matthes et al., 2006) or just the air-sea coupled response (Meehl et al., 2008; Meehl and Arblaster, 2009). One question this raises is to what extent are our results driven solely by changes in SST?

To address this, we have performed additional experiments in which the solar constant is held fixed and we prescribe the SSTs taken from the monthly averaged output of our REF and 0.1%S integrations. We label these new “AMIP-style” integrations REF(s) and 0.1%S(s). Fig. 4.4 shows that the jet shift of 0.1%S(s) is nearly the same as that of the 0.1%S integration, suggesting that the climate response is driven primarily through changes in SST.

We have performed an additional integration in which we perturb the solar constant while holding SSTs fixed. The resulting temperature response is mostly confined to the stratosphere, with little response in the troposphere. In this case (not shown), we do not recover climate responses that resemble the responses of 0.1%S and 0.1%S(2x). This further establishes that changes in SST are central to generating the climate responses. We should note, however, that our model may be biased in this regard, since it does not have a well resolved stratosphere, and it does not include interactive photochemistry. Also worth noting is that, because of nonlocal radiative effects, temperature changes in the stratosphere can lead to changes in SST. So the fact that SST changes drive the responses does not necessarily mean that the stratosphere plays a trivial role.

Another question this raises is what aspects of the SST background state lead to the opposite-signed responses of 0.1%S and 0.1%S(2x)? More specifically, is the SST pattern of the background state the key factor, or is the global mean SST more important? To consider this we have performed the REF(δs) and 0.1%S(δs)
integrations in which we prescribe the same SSTs used in the REF(s) and 0.1%S(s) integrations, but with 2.18 K added at each gridpoint. 2.18 K is the global mean change in temperature between the REF and REF(2x) integrations. Fig. 4.4 shows that 0.1%S(δs) produces almost no SH jet shift. This suggests that the global mean temperature of the background state plays some role in the response, but the SST pattern of the background state is also very important. This also means that, even without changes in CO₂ levels, biases in a model’s climate can substantially affect its response to small forcings. This may also explain why, as mentioned above, the results of Liu et al. (2013) appear to agree well with our 0.1%S(2x) integration: even though the global mean temperatures of the background states are quite different, the SST patterns of the background states may have much in common.

4.5 Conclusion

The key result of this study is that the response to small perturbations in TSI depends on the background state. A 0.1% increase in the solar constant applied to a present day background state produces a response that is in many respects opposite to that of an identical forcing applied to a doubled CO₂ climate. These responses are communicated almost entirely through changes in SST, and the pattern of the SST background state (not just its global-mean temperature) plays a crucial role in determining the climate response. There is little evidence to suggest that the climate responses are tropically driven, although the background state of the tropical Pacific may play an important role.

We have also shown that the response to solar forcing is nonlinear: when we impose a 2% increase in the solar constant, the results are not very sensitive to the background state. There is also evidence that CO₂ forcing exhibits a similar behavior: the response to small perturbations is highly dependent on the background state. This suggests that caution should be exercised when attributing decadal
trends: the signature of a forcing on decadal timescales may be very different from its signature on centennial timescales.

Our results may partly explain why past studies reach such disparate conclusions about the climate response to solar forcing. For example, the coupled model results of Shindell et al. (2006) agree well with our 0.1%S integration, while the results of Liu et al. (2013) agree well with our 0.1%S(2x) integration. Thus, differences in the background climate may explain why Shindell et al. (2006) and Liu et al. (2013) obtain different results.

To further ensure that our results are robust, we are extending our integrations beyond the 200 years analyzed here. We are also planning to perform integrations with GISS ModelE to see if other models produce similar results.
Chapter 5

Conclusion

In this thesis, we have used a variety of modeling experiments to gain insights into what has been driving changes in the large scale tropospheric circulation. Chapter 2 showed that changes in stratospheric water vapor produce HC expansion and poleward jet shifts that are comparable to what increased well mixed greenhouse gases produce. Furthermore, this circulation response is mostly driven by changes in the extratropical stratosphere, and is not entirely driven by changes in tropopause height.

In Chapter 3, we made sense of why the circulation response to El Niño is very much opposite to that of global warming. Specifically, a narrow thermal forcing centered at the equator produces El Niño–like HC contraction while a forcing with wider meridional extent produces global warming–like HC expansion. We devised a simple diffusive model to explain the transition from HC contraction to HC expansion.

In Chapter 4, we focused on the effect of perturbations in TSI with the amplitude of the 11-year solar cycle. We found that the climate response to such a forcing is highly dependent on the background state. A small TSI forcing applied to a present day climate produces a response that is in many respects qualitatively
opposite to that of an identical forcing applied to a doubled CO₂ climate. We showed that this background dependence is also apparent in the response to small CO₂ perturbations, suggesting that anthropogenic climate change may produce a climate response on decadal timescales that is very different from the one produced on centennial timescales.

This research prompts a number of questions that motivate additional work. Specifically, how do shifts of the midlatitude jet relate to shifts of the HC edge and subtropical jet? Kang and Polvani (2011) showed that the relationship can be quite complicated, with a systematic relationship apparent only in the SH during certain seasons. Taking a closer look at our results, we find that the structure and sign of external forcings play important roles in how the subtropical and midlatitude jets track with each other. Specifically, in response to stratospheric cooling, the shift of the HC edge is about 1/10 of the jet shift (Fig. 2.4a,b), whereas in response to tropospheric warming, the shift of HC edge can be 1/2 to 2/3 of the jet shift (Fig. 3.5). So future work may investigate the role of the stratosphere in driving the separation between the subtropical and midlatitude jet.

While Chapter 3 provides a simple explanation of HC expansion and contraction, our understanding of the shift of the midlatitude jet is much less clear. Recent studies have established that eddy momentum fluxes play a key role in shifting the midlatitude jet (e.g. Wu et al., 2012; Chen et al., 2012), and Barnes and Thompson (2013) have shown that much of the response is reproducible in a simple barotropic model. That said, more work is needed to develop a predictive theory for how the jet shifts.

Our results in Chapter 4 showed that the response to small perturbations in CO₂ depends on the background state. We can pursue this further to investigate the responses to successive small increases in CO₂. This may reveal more evidence that the circulation response to increased CO₂ on decadal timescales is nonmonotonic
and nonlinear. Another basic question raised by Chapter 4 is why the background state plays such an important role in the response to small perturbations? One way to investigate this further would be to see if a simpler model is capable of producing this behavior. For example, one could set up an aquaplanet model with an idealized Walker circulation by imposing the form for the implied ocean heat flux (cf. Merlis and Schneider, 2011). We can then vary the strength of this Walker circulation and investigate how this affects the response to small external forcings.
Bibliography


Forster, P. M. D. and K. P. Shine, 2002: Assessing the climate impact of


