

On the Circulation Response to Climatic Forcing



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A thesis submitted for the degree of

Doctor of Philosophy

Hilary Term, 2019

Abstract

Climate change is expected to cause changes in the general atmospheric circulation. Particularly important to the global population is an understanding of how weather and climate will change. North Atlantic jet stream changes will drive weather and climate changes, primarily over North-Western Europe, but also over North America, Asia and other parts of the globe. This makes it vital to understand how climate change will affect the jets to determine how the weather and climate will respond. However, the exact nature of the climate change forcings affecting the jet are uncertain.

Here, sensitivity studies using a hierarchy of models to determine possible circulation changes due to the thermal forcing of the atmosphere are performed. It is found that the location of forcing, both its latitude-longitude position and its height, has a profound effect on how the circulation responds. A simple mechanistic understanding for shifts and strengthening of the jet through changes in the mean meridional temperature gradients is presented, along with mechanisms for how Hadley circulation changes can affect the jet, and for how tropically forced waves affect the Northern Hemisphere circulation. Important differences between the responses in winter and summer are observed, which can lead to seasonally dependent extreme temperature and precipitation responses.

Key forcing locations that force the atmospheric circulation effectively are highlighted. In these locations, the uncertainties in how the forcing due to climate change will manifest must be reduced so that the circulation response can be better quantified. An important uncertainty is how the atmosphere heats up compared to how sea surface temperatures change and connected to this, how the land-sea temperature contrast will respond. As these uncertainties are reduced, the results from these sensitivity studies will provide a guide to understanding how the circulation will change and the associated impacts of the changes.

Acknowledgements

A thesis is much more than meets the eye. This work is the culmination of a significant period of my life and study at Oxford, and is a product of so many people who have contributed in many many ways.

I can go no further without expressing my utmost gratitude to my two supervisors, without whom, the cliché really is true, this work would not have been possible. I am indebted to the constant thought and guidance from Tim Woollings, whose insight, enthusiasm and encyclopaedic knowledge have made an immeasurable contribution to this work. He has provided continued support and mentorship throughout the years, a truly fantastic supervisor. I also owe a huge amount to Myles Allen; he first sparked my interest in pursuing climate science research with his lecture course. His passion and inspiration for the science have played a paramount part in this thesis, and I am grateful for all his time (especially when distracting him from the IPCC special report!) My deepest thanks also go to Cheikh Mbengue, for his invaluable thoughts, comments and insights on my work; his patience in helping run ‘his’ GCM; and for always having the time to answer my silly questions. Thanks also go to my thesis examiners, Paulo Ceppi and Peter Read, for their time critiquing this thesis, and the stimulating discussion in my viva.

There are a whole host of colleagues whom I would like to thank. To Richard Millar for his indispensable help and expertise on the carbon budget; to Sarah Sparrow for teaching me the ropes of climateprediction.net and tireless troubleshooting; to Chris O’Reilly for many stimulating ideas and discussions (about work and otherwise!); to Chris Forest for his contribution from afar; and to all those I have collaborated with: Dann Mitchell, David Karoly, Benoit Guilloid, Hideo Shiogama and Urs Beyerle. Thanks also to the atmospheric dynamics group in AOPP: my office mates Ronald Li and Matt Patterson who shared the DPhil life; and Marie Drouard and Peter Watson for their thoughts and ideas. I owe thanks to Gillian Lobo, Sophie Marjanac and ClientEarth, not only for financial support, but for helping me see the real world importance of the science we do. Thanks are also due to NERC for my funding, and to the climateprediction.net volunteers for running over 50,000 model years of simulations for this thesis.

Outside of my work, I am lucky to have been surrounded by such an amazing group of friends. To all my housemates in 46 and Number 9, and to all my many friends from OUAC, especially when winning la Decibel, it’s been a blast. Special thanks go to Izzy, who is always there for me and never fails to brighten my day. My final and greatest thanks are to my Mum and Dad: Jean and John. They have given me nothing but love and support from day one; without them I would not be where I am today.

List of Acronyms

AMIP	Atmospheric Model Intercomparison Project
AMO	Atlantic Multidecadal Oscillation
AMOC	Atlantic Meridional Overturning Circulation
CAM	Community Atmospheric Model
CFMIP	The Cloud Feedback Model Intercomparison Project
CGT	Circumglobal Teleconnection
CMIP	Coupled Model Intercomparison Project
DJF	December-January-February
EA	East Atlantic
ECMWF	European Centre for Medium-Range Weather Forecasts
ENSO	El Niño–Southern Oscillation
EOF	Empirical Orthogonal Function
EP	Eliassen-Palm
ERA	ECMWF Re-Analysis
GCM	General Circulation Model
GFDL	Geophysical Fluid Dynamics Laboratory
GHG	Greenhouse Gas
GTO	Global Teleconnection Operator
HadAM3P	Hadley Centre Atmospheric Model Version 3P
HAPPI	Half a degree Additional warming, Prognosis and Projected Impacts
HC	Hadley Cell
HCO ₂	High Carbon Dioxide
IPCC	Intergovernmental Panel on Climate Change
ITCZ	Intertropical Convergence Zone
JJA	June-July-August
LCO ₂	Low Carbon Dioxide
MIROC	Model for Interdisciplinary Research on Climate
MJO	Madden-Julian Oscillation
NAO	North Atlantic Oscillation
NCEP	National Centers for Environmental Prediction
NH	Northern Hemisphere
NOAA	National Oceanic and Atmospheric Administration
PNA	Pacific/North American
RCP	Representative Concentration Pathway
RPM	Random Perturbation Method
RWB	Rossby Wave Breaking
RWS	Rossby Wave Source
SLP	Sea Level Pressure
SST	Sea Surface Temperature
WBG	Wet-bulb Globe Temperature

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Chapter 1

Introduction

Climate change is one of the greatest challenges facing the planet in the 21st century. Increases in greenhouse gas concentrations affect the radiation transfer in the atmosphere, leading to changes in atmospheric thermal forcing. A key facet of climate change is understanding the changes in atmospheric circulation which are driven by the thermal forcing changes.

Over North-Western Europe, much of the weather is dictated by the behaviour of the North Atlantic jet stream. The jet is intimately connected to the storm tracks, which organise damaging midlatitude cyclone activity, making it vital to understand how climate change will affect the jets to determine how the weather and climate will respond. Further afield, changes in circulation over the North Atlantic can have wide ranging impacts over North America, North Africa and Asia. The greatest impact on society from these circulation changes comes not from the slowly changing trend in weather patterns, but from changes in extreme events such as heatwaves, flooding and drought. Understanding how the circulation will change is crucial if we are to know how we must act to adapt to climate change.

1.1 The problem

There are several connected problems which we face when seeking to understand how the circulation will change. Firstly, there is a larger than desirable uncertainty in the exact nature of forcing due to climate change. Namely, the quantification of the magnitude of the forcings, and also the exact spatial location and extent of the forcings. This impacts our understanding of how the circulation will respond because there are spatially varying responses to a particular forcing and also because there are competing forcings within the climate system. The second problem is the difficulty of identifying the specific drivers of processes causing change in the circulation and the mechanisms by which they act. Specifically, we can observe circulation changes in reanalysis and coupled climate models, but without performing idealized experiments it is very hard to determine causality and explicitly attribute circulation changes to a specific forcing. This is compounded by the uncertainty in the exact forcings, so climate model simulations in different models which simulate these forcings differently can produce a wide range of circulation changes. A part of this problem is the issue of disentangling the forcing from the response. As the climate system has many constituent parts with complex interactions and feedbacks, it is hard to separate forcings from responses. This means an understanding of these interactions and feedbacks is vital for studying the circulation response to climate change.

There are a plethora of modelling studies, both using idealized and more complex general circulation models, which take the second problem and try to determine which are the important drivers and to what extent they force the circulation. Understanding the exact mechanisms by which the different forcings affect the circulation allows us to disentangle the effects of a myriad of potentially important forcings. There are several common approaches that studies use to tackle the problem. One approach is to impose an idealized forcing into a model and study the response of the circulation and look at the mechanism by which the response arises. Another approach is to account for

various different forcings within a model using regression. However, these approaches are limited in their ability to fully assess the range of possible responses due to the nature of the uncertainty in the forcing itself.

This thesis focuses on the second problem, investigating the use of sensitivity studies as a tool to understand how the atmosphere responds to forcing. A sensitivity study is a study where the forcing is systematically varied in increments, across a set of parameters either in amplitude or in location, to build a full picture of potential circulation responses to forcing. The key advantages to sensitivity studies are that they can isolate specific locations where forcings are important, and that they assess the full range of possible responses. Using sensitivity studies we can identify the key forcing locations where we need to better constrain the uncertainty in the forcings, thus highlighting where we need to direct our efforts towards in tackling the first problem. We can also easily examine the mechanisms by which a specific forcing with specific parameters affects the circulation. The spatial structure of sensitivities may also inform on the response mechanisms. A corollary of using sensitivity methods to investigate the atmospheric response to forcing is that it lets us examine the circulation response not only to climate change, but also to other atmospheric forcings arising through natural variability of the coupled climate system.

1.2 Thesis aims and outline

Here we will use a hierarchy of models to conduct a range of sensitivity studies on the atmospheric response to forcing. This will allow us to study the responses in an idealized scenario, where it is easier to disentangle cause-and-effect, before verifying and expanding our findings in a more complex model. We will also look at the sensitivity to different types of forcing, and quantify their effect on circulation and the potential impacts on society.

The key aims of this thesis are:

1. To quantify the sensitivity of the atmospheric circulation to forcing.
2. To investigate the seasonal dependence of the sensitivities.
3. To determine the dynamical mechanisms by which the atmospheric circulation responds to forcing.
4. To explore potential impacts of the range of circulation changes on temperature and precipitation extremes.

The thesis is structured as follows:

- In Chapter 2 we provide an in-depth review of the important literature. First we explore the climatology of the atmospheric circulation, focusing on the North Atlantic region. We summarise the breadth of studies which investigate changes in the North Atlantic circulation, looking at the different methods and metrics used to evaluate our current understanding of circulation changes and identify areas where further understanding is required. We also outline some of the theory behind the atmospheric circulation, which is important to develop our understanding of potential circulation response mechanisms.
- In Chapter 3 we describe the first of the models used in this thesis, the Geophysical Fluid Dynamics Laboratory Flexible Modeling System. After explaining the model and sensitivity study setup, we present results showing the sensitivity of the eddy-driven jet to localized diabatic heating. We explore the seasonality of the sensitivity and the linearity of the responses to forcing magnitude before discussing how our results compare with the literature and what new knowledge they add. The majority of this work has been published in *Journal of Climate* (Baker et al., 2017).
- In Chapter 4 we take the same sensitivity experiments and use them to investigate the sensitivity of the Hadley circulation to forcing. We then use this understanding

to explain the cross-hemispheric response of the eddy-driven jet to forcing which is mediated by the Hadley circulation. Again, we explore the seasonality of the responses which arises here due to the contrast between the winter and summer Hadley circulations. The majority of this work has been published in *Geophysical Research Letters* (Baker et al., 2018a).

- In Chapter 5 we present the results of several spinup case studies performed using the idealized model. These studies allow us to explain the mechanisms by which the circulation responses are set up by observing the transient response of the model as the forcing is applied, through to when the model reaches a statistically steady state.
- In Chapter 6 we investigate the sensitivity of the North Atlantic circulation to sea surface temperatures in a more complex, atmosphere-only, general circulation model, HadAM3P. We use a linear statistical-dynamic method, employing a Green's function approach to quantify the circulation sensitivity. The sensitivity maps highlight key regions where SSTs can force the circulation, and the seasonality associated with these forcings. The method is used to reconstruct and forecast North Atlantic atmospheric circulation indices with some skill. We also use a regression based approach to investigate the mechanisms by which the circulation responds to the forcing. This work is under review in *Journal of Climate* (Baker et al., 2019a).
- In Chapter 7 we turn to understanding the response of the circulation to different types of forcing using HadAM3P. We look at the effects of sea surface temperatures versus direct radiative forcing in forcing the Northern Hemisphere circulation. We compare the results from HadAM3P to two other general circulation models, and then use a simple barotropic model to identify the key mechanisms and locations for the circulation responses we observe. We discuss how further quantification of the uncertainty in land-sea temperature contrast response will

allow us to draw more informed conclusions about the exact circulation changes in the future. The majority of this work has been published in *Climate Dynamics* (Baker et al., 2019b).

- In Chapter 8 we build on the previous chapter, examining the implications of the sensitivity of the atmospheric response to direct radiative forcing. In particular, we focus on the changes in extreme temperature and precipitation likelihoods due to direct radiative forcing effects on circulation. We then explore how, due to the circulation response to direct radiative forcing, the uncertainty in the climate sensitivity has important implications in the light of current policy targets aiming to limit global mean warming to 1.5°C. The majority of this work has been published in *Nature Climate Change* (Baker et al., 2018b).
- Finally, in Chapter 9 we draw our results together and present our conclusions, highlighting several areas of further work that would be useful to test our results further.

Chapter 2

Background

We begin with a succinct review of theories of the Hadley circulation, which leads into a description of the eddy framework required to understand the dynamics at the edge of the tropics and into the midlatitudes. From here, we will see how two different types of jet can be formed in the midlatitudes and their link to the eddies and regions of baroclinicity. We will then review the range of different theories that seek to explain changes in the eddy-driven jet. A large part of this thesis focuses on the North Atlantic circulation, so we will then review literature pertinent to the North Atlantic eddy-driven jet, the modes of North Atlantic variability, and studies using a range of model complexities looking at the sensitivity and response of the jet and modes of variability to forcing.

2.1 Theories of Hadley circulation

2.1.1 Early theories

The Hadley cell is a prominent circulation feature characterised by ascending air near the equator and descending air in the subtropics. It is fundamental in regulating the Earth's energy and momentum budgets. Hadley proposed an axially symmetric convection cell, a steady, thermally direct, circulation about the Earth's rotation axis, driven by the pole-to-equator temperature contrast (Hadley, 1735). Observations of midlatitude

tum conserving flow (Fig. 2.1). Under these assumptions, the Hadley cell extends to a finite latitude, beyond which radiative-convective equilibrium holds, with extent:

$$\phi_{HH} = \left(\frac{gH_t}{\Omega^2 a^2} \frac{\Delta_h \theta_R}{\theta_0} \right)^{1/2}, \quad (2.1)$$

and strength:

$$|\Psi_{HH}| \sim \rho_0 \frac{a^2 H_t}{\tau} \left(\frac{gH_t}{\Omega^2 a^2} \frac{\Delta_h}{T_0} \right)^{3/2} \frac{\Delta_h}{\Delta_v}. \quad (2.2)$$

Here, g is the gravitational acceleration; H_t the height of the upper branch of the Hadley circulation (i.e. approximately the tropical tropopause height); Δ_h the pole-to-equator temperature difference; θ_R the equator-to-pole radiative-equilibrium temperature difference; θ_0 a constant reference value of potential temperature; ρ_0 a reference density; τ the Newtonian relaxation timescale; Δ_v the gross stability; and T_0 a reference temperature. With typically observed Earth values, $\phi_{HH} \approx 22^\circ$ and $\Psi_{HH} \sim 10 \times 10^9 \text{ kg s}^{-1}$. This latitudinal extent is slightly smaller than observed on Earth, but the strength estimate is about an order of magnitude too small. The fact that these results are wrong is not too surprising because angular momentum is not conserved and the tropical atmosphere is not energetically closed (Walker and Schneider, 2006). Due to the non-linearity of the angular momentum conserving constraint, the circulation strength responds nonlinearly to movements of the maximum heating off the equator (Lindzen and Hou, 1988), see Figs. 2.2a,b,c. Whilst coincidentally improving the strength estimate of the model, there is no evidence for this nonlinear amplification of the Hadley cell strength on Earth (Dima and Wallace, 2003), suggesting that the axisymmetric theory of meridional circulation is not a good theory of the Earth's meridional circulation.

2.1.2 Baroclinic instability

Axisymmetric flows were originally only used as providing a basic state for baroclinic instability (Schneider, 1977). Zonal wind speeds consistent with an angular momentum

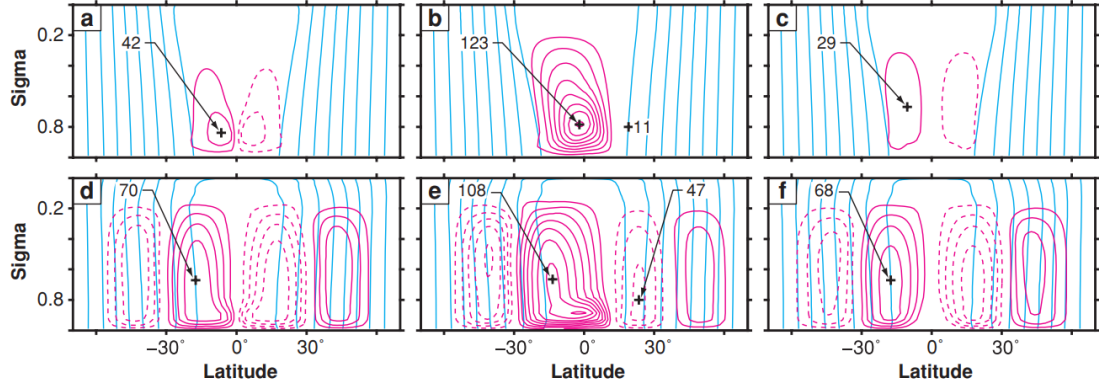


Figure 2.2: Eulerian mass flux streamfunctions (magenta) and angular momentum (light blue) in axisymmetric (top row) and macroturbulent (bottom row) simulations. (a, d) Maximum radiative equilibrium temperature on equator ($\theta_0 = 0^\circ$). (b, e) Maximum radiative equilibrium temperature north of equator ($\theta_0 = 6^\circ$). (c, f) Average of circulations for $\theta_0 = 6^\circ$ and $\theta_0 = -6^\circ$. Contour intervals are $15 \times 10^9 \text{ kg s}^{-1}$ for streamfunction and $0.1 \Omega a^2$ for angular momentum. Streamfunction extrema above $\sigma = 0.85$ are identified. Figure reproduced from Schneider (2006).

conserving mean flow would be large enough in the subtropics to be linearly unstable to baroclinic instability (Phillips, 1954). This led to the idea that the axisymmetric Hadley circulation may terminate at the latitude at which vertical shears become baroclinically unstable (Held, 2000). Taking the stability criterion in a two-layer quasigeostrophic model that the vertically averaged temperature gradient must not exceed the critical temperature gradient for baroclinic instability, assuming angular momentum conserving wind, and making the small angle approximation, leads to a Hadley cell extent given by:

$$\phi_{BC} = \left(\frac{b}{2} \frac{g H_t}{\Omega^2 a^2} \frac{\Delta_v \theta}{\theta_0} \right)^{1/4}, \quad (2.3)$$

where b is a constant (Korty and Schneider, 2008) and $\Delta_v \theta$ is a measure of the static stability. With typically observed Earth values, $\phi_{BC} \approx 22^\circ - 27^\circ$. Kang and Lu (2012) generalize this model to all seasons by relaxing the angular momentum conservation criterion as angular momentum is not perfectly conserved in the upper branch of the Hadley cell, especially in summer. By introducing a Rossby number which characterises the extent to which the wind deviates from the angular momentum conserving state, they

obtain estimates for the Hadley cell edge in summer and winter.

Using the two-layer model is heuristic as there is no critical temperature for baroclinic instability in a continuously stratified atmosphere (Schneider, 2006). Further theories of Hadley cell extent in a more realistic atmosphere utilise the concept of supercriticality (Walker and Schneider, 2006; Schneider and Walker, 2006; Schneider and O’Gorman, 2008; O’Gorman, 2011; Levine and Schneider, 2015). We will not discuss these theories here, but the basic concept separates the atmosphere into regions where baroclinic wave activity strongly affects the mean flow thus setting the tropopause height (i.e. the extratropics), and regions where radiative-equilibrium sets the tropopause (i.e. the tropics).

One might think the two Hadley circulation scaling theories (Held-Hou and the baroclinic theory) could be combined, ϕ_{HH} in the smaller temperature contrast regime, and ϕ_{BC} in the larger temperature contrast regime. On Earth however, with $\phi_{HH} \approx \phi_{BC}$, it is difficult to discern which mechanism may be operating in the atmosphere. Recent work suggests that the baroclinic instability theory may explain the poleward Hadley circulation terminus shift in comprehensive GCM simulations (Lu et al., 2007; Frierson et al., 2007; Kang and Lu, 2012). But a comparison with idealized GCM simulations shows significant discrepancies (Zurita-Gotor and Lindzen, 2007) consistent with other findings that the Hadley cell extent predicted by this model is poor (Walker and Schneider, 2006; Korty and Schneider, 2008).

In the presence of baroclinic instability, nonaxisymmetric perturbations grow, leading to macroturbulent statistically steady states (Schneider, 2006), see Figs. 2.2d,e,f. Eddies transport angular momentum into the midlatitudes, with surface westerlies occurring in zones of upper-level angular momentum flux convergence. Meridional temperature gradients are reduced by the poleward transport of heat by the eddies.

The influence of eddies on the Hadley circulation strength can be characterised by the local Rossby number $Ro = -\bar{\zeta}/f$, where $\bar{\zeta}$ is the temporal and zonal mean relative vorticity. The Rossby number measures the importance of eddy versus thermal driving

of the circulation strength (Bordoni and Schneider, 2008). The zonal momentum balance in the upper branches of the circulation is approximately:

$$(f + \overline{\zeta}) \bar{v} = f (1 - \text{Ro}) \bar{v} \approx S, \quad (2.4)$$

where S is the eddy momentum flux divergence, and the overbar indicates a temporal and zonal mean along isobars (Schneider and Bordoni, 2008). As $\text{Ro} \rightarrow 0$, the circulation is controlled by eddy momentum fluxes and only responds to variations in thermal driving indirectly through eddy momentum flux changes. As $\text{Ro} \rightarrow 1$, the circulation approaches the angular momentum conserving limit and its strength directly responds to thermal driving. In this case, the eddy momentum flux divergence approaches zero, and the mean meridional circulation decouples from the eddy momentum flux divergence (Schneider and Bordoni, 2008). The local Rossby number in the upper branches of Earth's Hadley cells varies from $\text{Ro} \lesssim 0.2$ in the summer cell to $\text{Ro} \gtrsim 0.5$ in the cross-equatorial winter cell, with intermediate values in the equinox cells (Walker and Schneider, 2006). In the dry idealized GCM of Schneider and Walker (2006), driven by the relaxation of temperatures toward radiative equilibrium, changes in the Hadley circulation strength are mediated by changes in the eddy momentum flux divergence (Walker and Schneider, 2006). Heat fluxes can be incorporated into the nearly inviscid theory by modifying the thermal driving (Schneider, 1984), but the dependence on eddy momentum fluxes causes departures of the atmosphere from the nearly inviscid limit. Eddies act to strengthen the equinoctial (compare Figs. 2.2a and d) and summer Hadley cells, but weaken the winter Hadley cell (compare Figs. 2.2b and e) which has a higher Rossby number (Schneider, 2006). Unlike the axisymmetric theory, in macroturbulent simulations the mass flux is on the same order of magnitude as Earth's observed Hadley cells (Schneider, 1984; Becker et al., 1997; Walker and Schneider, 2005).

This eddy framework also brings greater understanding to the Hadley cell extent question. The circulation extent can be considered as the equatorward boundary of the

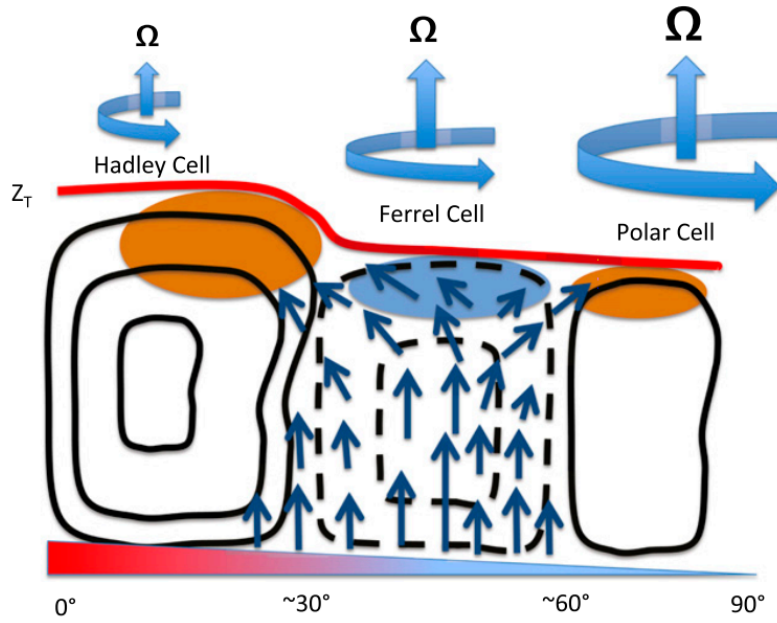


Figure 2.3: Cartoon of the effect of macroturbulence on the extent of the Hadley circulation. When the thermal contrast becomes large enough, baroclinic waves set the height of the tropopause (red line). A change in the sign of the potential vorticity gradient between the surface layer (as shown by temperature decreasing poleward at the surface) and the interior of the troposphere (as shown by planetary vorticity increasing poleward) triggers baroclinic instability near the surface. Eddies then propagate upward and equatorward into the subtropical upper troposphere, where they dissipate (blue arrows). The terminus of the Hadley circulation corresponds to a boundary between regions of eddy dissipation (subtropics) and eddy production (middle latitudes). In the upper troposphere, this is characterized by a transition between a region with eddy momentum flux divergence (orange patches) and poleward mean meridional winds (Hadley cell; solid black contours) to a region of eddy momentum flux convergence (blue patch) and equatorward mean meridional winds (Ferrel cell; dashed black contours). Figure reproduced from Levine and Schneider (2015).

region where baroclinic eddies become deep enough to reach the upper troposphere (Schneider and Walker, 2006). In the upper troposphere, wave activity propagates horizontally instead of vertically, leading to angular momentum convergence into the latitude band where it was generated (Edmon et al., 1980; Held and Hoskins, 1985). The Hadley cell terminus lies at the boundary between eddy production in the midlatitudes and dissipation in the subtropics. In the upper troposphere, this is the boundary between eddy momentum flux convergence in the midlatitudes and eddy momentum flux divergence in the subtropics (Fig 2.3).

2.2 The jets

At the poleward extent of the Hadley cell is the subtropical jet, located in the upper troposphere. This forms as a result of the poleward angular momentum transport by the thermally direct Hadley circulation (Held and Hou, 1980). Conservation of angular momentum of air moving poleward near the tropopause creates the zonal eastward flowing subtropical jet. An eddy-driven jet forms further poleward in the midlatitudes as a result of transient eddy momentum and heat flux convergence by atmospheric waves that develop in regions of enhanced baroclinicity (Held, 1975; Rhines, 1975; Panetta and Held, 1988; Panetta, 1993). This can be thought about in terms of potential vorticity. Following Vallis (2006), the quasigeostrophic potential vorticity, q , is given by:

$$q = f_0 + \beta y + \nabla^2 \psi + \frac{f_0^2}{\rho_0} \frac{\partial}{\partial z} \left(\frac{\rho_0}{N^2} \frac{\partial \psi}{\partial z} \right), \quad (2.5)$$

where f_0 is the Coriolis parameter at latitude ϕ_0 , β is the meridional gradient of the Coriolis parameter, $\nabla^2 \psi$ is the relative vorticity (ψ is the two-dimensional streamfunction), N is the buoyancy frequency and ρ_0 is the fluid density. Taking the zonally averaged meridional gradient gives:

$$\frac{\partial \bar{q}}{\partial y} = \beta - \frac{\partial^2 \bar{u}}{\partial y^2} + \frac{f_0^2}{\rho_0} \frac{\partial}{\partial z} \left(\frac{\rho_0}{N^2} \frac{\partial \bar{u}}{\partial z} \right), \quad (2.6)$$

where the overbars indicate zonal averaging and u is the zonal wind. From this equation, it is clear that there exists a strong potential vorticity gradient in the midlatitudes, where the meridional temperature gradient is large. The eddies occur at the polar front where polar air and warmer tropical air are separated by a steep meridional temperature gradient, creating a region of enhanced baroclinicity. Rhines theory tells us that, on a two dimensional sphere with turbulence, energy cascades from the small scales up to scales where Rossby wave propagation occurs (Rhines, 1975, 1979). These Rossby waves, po-

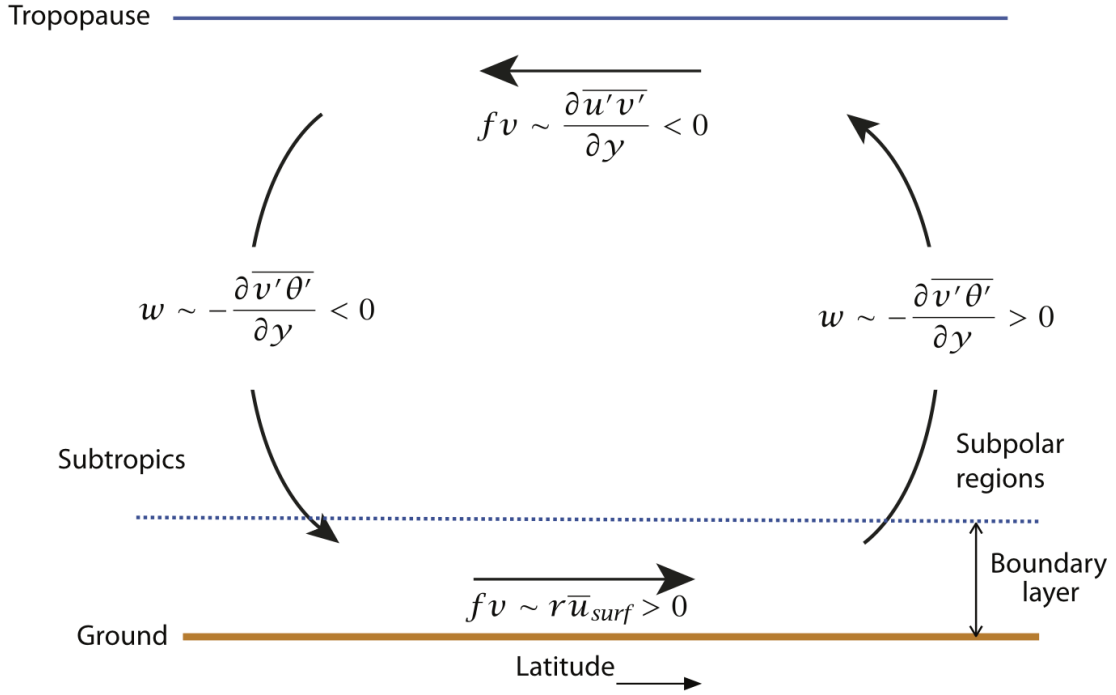


Figure 2.4: The eddy-driven Ferrel Cell, from an Eulerian point of view. Above the planetary boundary layer the mean flow is largely in balance with the eddy heat and momentum fluxes, as shown. The lower branch of the Ferrel Cell is largely confined to the boundary layer, where it is in a frictional-geostrophic balance. Figure reproduced from Vallis (2006).

tential vorticity-conserving motions, propagate north and south from the front and break and dissipate on either side of it. From the group velocity of these Rossby waves, it can be shown that the momentum flux associated with the Rossby waves is toward the source region (Vallis, 2006). This converges momentum into the source region (Held, 1975), sharpening the midlatitude potential vorticity gradient and creating a fast moving current of air, the eddy-driven jet. This eddy-driven jet is self maintaining due to the constant creation, radiation, and dissipation of Rossby waves, giving a self-sharpening potential vorticity gradient.

The generation of large-scale eddies in the midlatitudes propagating into the subtropics leads to meridional momentum flux polewards and downwards, giving the observed surface westerlies via the Coriolis effect (see Fig. 2.4). Eddies therefore affect the jet through direct eddy fluxes, and the associated meridional circulation. These ef-

fects can be combined in the transformed Eulerian-mean Eliassen-Palm flux framework, described and used in Chapter 5. The eddy-driven transport, known as the Ferrel Cell, also weakens the subtropical jet, explaining why it is weaker than predicted by angular momentum conservation (Hartmann, 2007). However, the Ferrel Cell, arising from the Eulerian zonal averaging, is not the full picture, as the direct eddy fluxes cause westerly acceleration (as opposed to the easterly acceleration of the equatorward flow by the Coriolis force in the upper branch of the Ferrel Cell). There is a strong cancellation between the eddy and mean flow terms, and this motivates describing the flow using the Transformed Eulerian Mean framework. This framework allows the underlying equations to be described in terms of a residual streamfunction. The framework also combines all the eddy terms into the Eliassen-Palm flux term of the momentum equation, where previously they appeared in both the momentum and temperature equations, which are coupled in thermal wind balance. This framework is outlined mathematically and applied in Chapter 5. The net effect of the eddies is to barotropise the zonal flow, weakening the westerlies aloft and strengthening them near the surface. The storm tracks, regions in the atmosphere where cyclonic activity predominantly occurs, lie in the baroclinic regions and interact with the jet; cyclones feed back onto the eddy-driven jet through eddy-mean flow interactions, and are also guided by the flow (Hoskins et al., 1983).

Both types of jet co-exist in the atmosphere, forming a single combined jet, or two distinct jets. Fig. 2.5 illustrates this, with a split jet in winter over the North Atlantic, and a merged jet over the Pacific. Lee and Kim (2003) study the relationship between the subtropical and eddy-driven jets. They show that baroclinic wave growth occurs in regions where there exists a strong subtropical jet, but where a moderate subtropical jet occurs, the most favourable region of baroclinic wave growth lies poleward of the subtropical jet, causing a separated eddy-driven jet to form. Son and Lee (2005) investigate the jet structures in a primitive equation model. Their experiments demonstrate that the interplay of tropical heating and high-latitude cooling dictates whether the eddy-driven jet is separated from the subtropical jet. Weak tropical heating and high

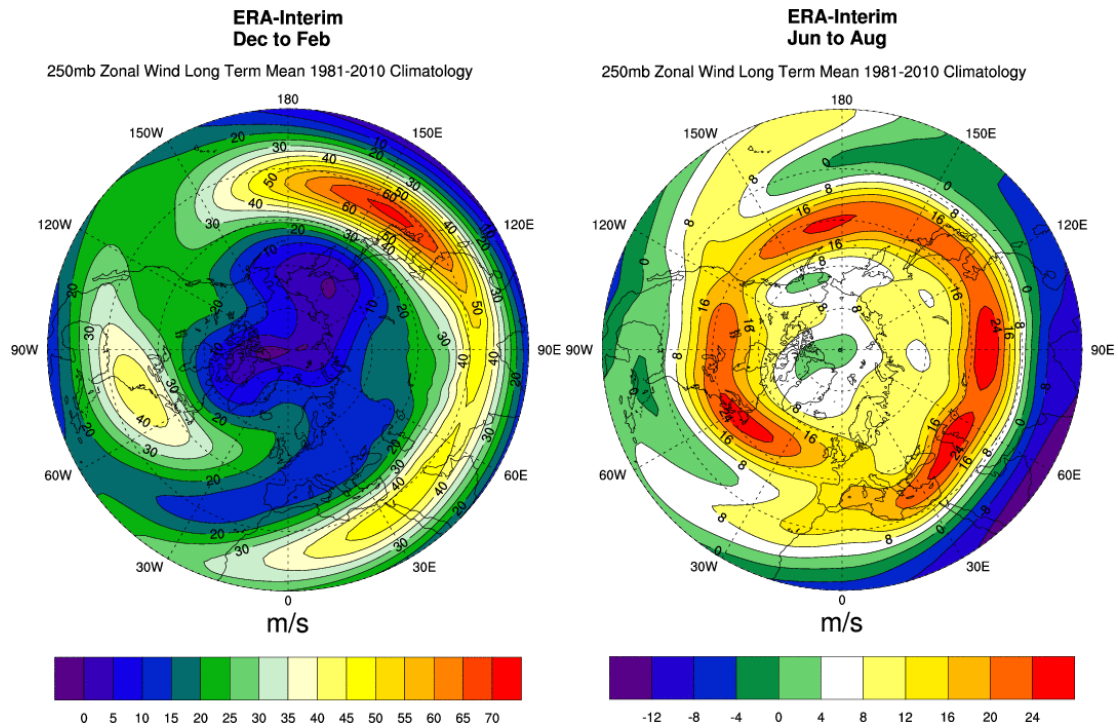


Figure 2.5: Winter (left) and summer (right) Northern Hemisphere 250mb zonal wind climatology (1981-2010) from ERA-40.

latitude cooling causes the eddy-driven jet to separate, forming a double-jet state. Over the North Atlantic, the two jets are predominantly split in mid-winter, but can merge into a single jet in early or late winter (Eichelberger and Hartmann, 2007). In summer, the overriding state is a much weaker, poleward shifted jet due to weakened meridional temperature gradients (Hall et al., 2015). In contrast to the North Atlantic, the Pacific jet is overwhelmingly found in the single jet state, both eddy and thermally driven. Li and Wettstein (2012) demonstrate this by decomposing the processes driving the jet variability into a thermally driven and an eddy driven component. They show that the primary mode of variability in the Pacific jet is a pulsing (an increase in strength and length of the jet), compared to a latitudinal shift of the jet over the North Atlantic.

2.2.1 Variability

A simple Gaussian model of a jet can be used to explore the different types of variability in a jet. Wittman et al. (2005) show three different modes of variability: pulsing,

a change in the jet strength; wobbling, a shift in the position of the jet; and bulging, a change in width of the jet distribution. In reality, this bulging can, and often does, only occur on one flank. Whether the primary mode of variability in the jet is a shift in latitude or a pulsing of the jet depends largely on the type of jet (Thompson et al., 2003). Basic differences arise between the thermally driven subtropical jet and eddy-driven jet as the subtropical jet has an externally imposed potential vorticity gradient, whereas the eddy-driven jet has eddy-mean flow coupling which self-maintains a potential vorticity gradient. The externally imposed potential vorticity gradient means that thermally driven jets have strong zonal flow and vertical wind shear, but are latitudinally confined. Rossby waves propagating along the gradient are trapped meridionally and so eddies that grow along the jet have limited movement meridionally, preventing large meridional meanders or shifts in zonal flow. In contrast, eddy-driven jets are driven by eddy momentum fluxes. Without the externally imposed potential vorticity gradient, the eddy driven jet is more mobile and can shift in latitude more easily. A potential mechanism for this variability, suggested by Lee and Feldstein (1996), is that meanderings occur when the meridional eddy size is insufficiently large to fully remove the negative potential vorticity gradient in the lower layer. The eddies then move towards regions where the lower level potential vorticity gradient is large to stabilize the flow, causing a shift in jet latitude (Thompson et al., 2003).

The differences in potential vorticity gradients suggest subtropical jets have less variability than eddy-driven jets, which exhibit most variability when the size of the baroclinic zone is larger than the eddies. As such, which mode of variability dominates depends on the strength of the subtropical jet and the size of the baroclinic zone. A strong jet and/or small baroclinic zone favours pulsing, whilst a weaker jet with a larger baroclinic zone is more likely to exhibit jet shifts. Barnes and Hartmann (2011) investigate the effect on eddy-driven jet variability depending on mean jet position. They find that as the mean jet position moves poleward, the leading mode of variability transitions from a shift to a pulse due to a reduction in wave breaking as β decreases further pole-

ward. However, if the jet mean position moves equatorwards, the leading mode also transitions to a pulse as the subtropical winds affect eddy propagation. As well as these changes in the jet, other characteristics of variability that can vary are the amount of variability displayed by the mean jet position, shape and strength (Barnes and Polvani, 2013).

There are a variety of mechanisms proposed to explain shifts in jet latitude in response to forcing. One theory is that changes in low-level baroclinicity can drive shifts in the storm track (Yin, 2005; Brayshaw et al., 2008; Frierson, 2008; Lu et al., 2010; Butler et al., 2011). Another theory states that changes in the upper troposphere and lower stratosphere modify wave propagation characteristics, leading to storm track shifts (e.g. Kushner and Polvani, 2004; Chen and Held, 2007; Butler et al., 2010; Ceppi and Hartmann, 2013). Chen et al. (2008) suggest that in the upper troposphere, an increase in meridional temperature gradient plays a key role in shifting the jet poleward. Lorenz and DeWeaver (2007) suggest that raising the tropopause height is responsible for a shift in the midlatitude jet, but the exact dynamical reason for this is as yet undetermined, suggesting also that near-surface changes in the meridional temperature gradient play a secondary role in the jet shift. However, Mbengue and Schneider (2017) show a near collocation between the maxima of near-surface eddy kinetic energy and near-surface extratropical temperature gradients in their climate change simulations, suggesting that near-surface meridional temperature gradients may play more than a secondary role. Other theories suggest that an increase in eddy length scales under global warming may contribute to storm track shifts (Kidston et al., 2010; Rivière, 2011), that increased Rossby wave reflection of low phase speed waves across the jets act to cause the shift (Lorenz, 2014) or that the effective diffusivity, as diagnosed using finite-amplitude wave theory, increases and advances upward and poleward, leading to a poleward displaced potential vorticity gradient peak in the upper troposphere (Lu et al., 2014).

There is evidence that stratospheric dynamics and coupling to the troposphere can influence the eddy-driven jet (e.g. Polvani and Kushner, 2002; Gerber and Polvani, 2009;

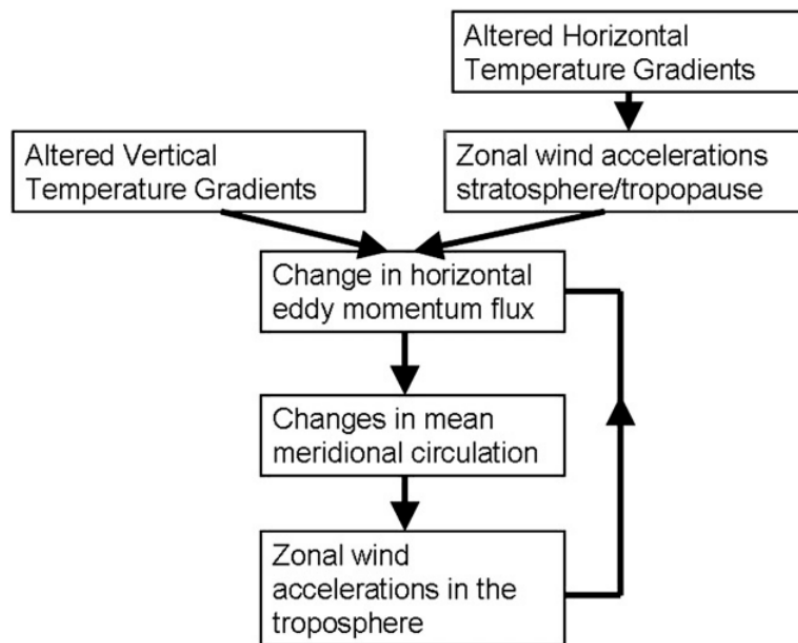


Figure 2.6: Summary of the mechanism by which the heating perturbation in the stratosphere influences tropospheric circulation. Figure reproduced from Simpson et al., 2009.

Son et al., 2009; Manzini et al., 2014). Whilst this thesis will concentrate on tropospheric drivers, it is useful to acknowledge that there are stratospheric processes that can also affect the circulation. Polvani and Kushner (2002) show that a sufficiently strong cooling in the polar winter stratosphere can act to shift the jet poleward and strengthen it at the surface. They show that the tropospheric response is highly correlated with the leading mode of internal variability. This is consistent with Baldwin and Dunkerton (2001), who show that stratospheric polar vortex precursors in the Northern Hemisphere precede certain extreme midlatitude tropospheric weather regimes in winter. However, Simpson et al. (2010) show that the response to stratospheric forcing is affected by the tropospheric climatology. Introducing topography into the GCM, so breaking the zonal symmetry, reduces the magnitude of the response to stratospheric heating. It is also found that lower-latitude jets/narrower jets exhibit a much larger response to stratospheric heating than higher-latitude/wider jets. The exact mechanism by which the jet structure leads to the difference in response is still an active area of research. A gen-

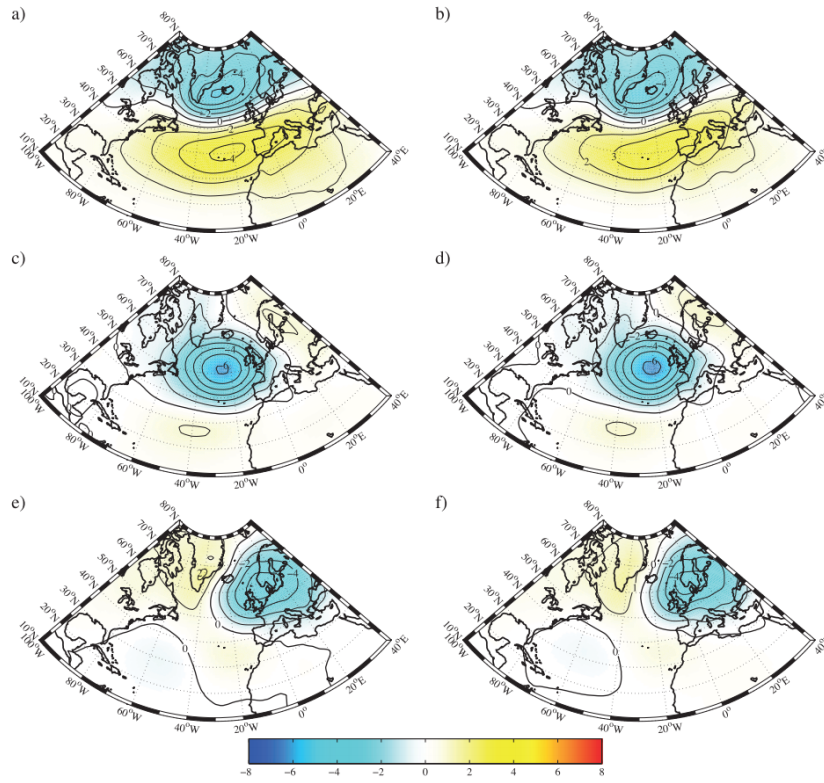


Figure 2.7: The three leading EOFs of the sea level pressure field for the winter months (DJF). (left) The NCEP reanalysis for the period 1948-2011 and (right) the 20CR for the period 1871-2008. (a,b) The leading EOF represents the NAO teleconnection, while the (c,d) second and (e,f) third represent the East Atlantic and Scandinavian teleconnections, respectively. Figure reproduced from Moore et al., 2013.

eral mechanism of tropospheric response to stratospheric perturbations is outlined by Simpson et al. (2009), although their model does not include a polar vortex. Using a spin-up ensemble experiment, where the perturbation is applied at day zero and then the evolution of the response analysed, they find that the changing eddy momentum fluxes are key in propagating through the chain from perturbation to response. The chain is summarised in Fig. 2.6.

2.3 The North Atlantic

The North Atlantic eddy-driven jet strength and position are tied to the modes of atmospheric variability over the Atlantic Ocean. The primary pattern of variability is the

North Atlantic Oscillation (NAO), a meridional sea level pressure dipole with centres of action near Iceland and over the subtropical Atlantic (Visbeck et al., 2001), shown in Figs. 2.7a,b, and a result of the North Atlantic storm track and associated eddy-driven jet (Vallis and Gerber, 2008). An NAO positive index corresponds to an increased pressure gradient between the two centres, and a negative index corresponds to a decreased pressure gradient. The NAO comes out as the dominant pattern of variability in the North Atlantic when using an empirical orthogonal function (EOF) analysis (which is commonly used to identify leading modes of atmospheric variability) on sea level pressure over the North Atlantic sector. Gerber and Vallis (2009) suggest that the North Atlantic storm track and the NAO are in fact the same phenomenon, and that the index describing the NAO also describes the interaction of the subtropical jet and the eddy-driven jet. They suggest a positive NAO corresponds to a jet splitting and a negative NAO a jet merging. However, Woollings et al. (2010) show that in order to account for variability in the jet speed and position, the East Atlantic (EA) pattern must be considered too, and even then it may still not be exactly possible to describe the state of the jet. The EA (Figs. 2.7c,d) is the second mode of variability in the North Atlantic, and is variously defined as a tripole (Wallace and Gutzler, 1981), dipole (Barnston and Livezey, 1987) or monopole (Moore et al., 2013) pattern of sea level pressure, with the main centre of action on the nodal line of the NAO off the west coast of Ireland. Woollings et al. (2010) show that the NAO and EA both describe combined changes in the latitude and speed of the jet, and Moore et al. (2013) suggest the Scandinavian pattern (Figs. 2.7e,f), the third mode of variability, with a centre west of Bergen (Barnston and Livezey, 1987), is tied to the jet too.

With the behaviour of the jet linked closely to the NAO, it is insightful to examine the variability in the NAO and to what extent it is predictable. Hurrell and Deser (2009) review the modal variability of the NAO, concluding that the variability arises from internal, non-linear dynamics of the extratropical atmosphere. Recent skill in the seasonal forecasts of the winter NAO (Scaife et al., 2014) suggests that much of the variability

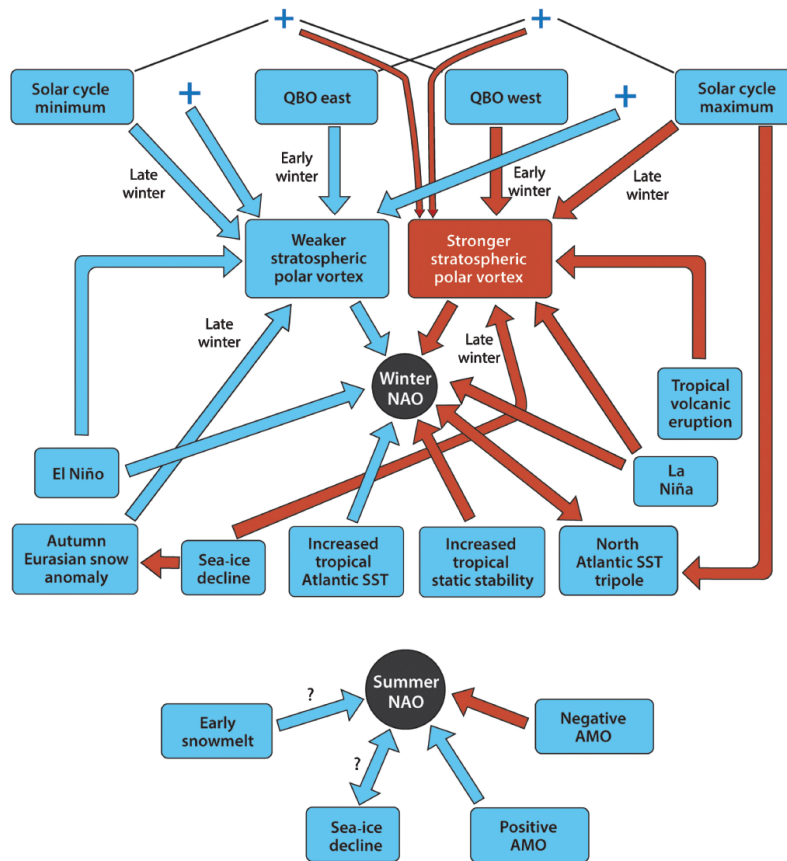


Figure 2.8: A conceptual model of some of the potential drivers of winter eddy-driven jet and NAO variability derived from the literature. A black (red) arrow indicates a strengthening of the target box, while a grey (blue) arrow indicates a weakening. (b) As above but for the summer eddy-driven jet and NAO, NAO can be used as a surrogate for eddy-driven jet variability. Arrows are not proportional to strength of potential forcing or confidence attached to the potential forcing. Figure reproduced from Hall et al., 2014.

in the NAO is indeed forced by a predictable driver, implying that the NAO is not just a manifestation of the aggregation of stochastic weather events, as had been postulated as a potential explanation for the existence of the NAO (Stephenson et al., 2000). Although the existence of the NAO is entirely down to atmospheric processes (Thompson et al., 2003), the North Atlantic Ocean still has an influence on atmospheric circulation. In winter, Atlantic SSTs appear to force the NAO, explaining much of the multiannual and multidecadal variability of the NAO (Rodwell et al., 1999). Coupling between the NAO and the SSTs means that the phase of the NAO can force a tripole pattern in North Atlantic SSTs through anomalous heat fluxes and wind stress curl (Kushnir et al., 2002).

This pattern then persists at depth over the summer and its re-emergence the following winter gives some element of predictability of the NAO (Cassou et al., 2007).

The decadal and longer timescale variability is investigated by Woollings et al. (2014), who show that on interannual-decadal (<30 year) timescales, the winter NAO describes predominantly latitudinal variations in the jet, whilst on longer timescales it represents changes in the jet strength. These longer timescales appear in the ocean as the Atlantic Multidecadal Oscillation (AMO), a basin-wide warming and cooling in the North Atlantic. Statistical analyses show that there exists a negative correlation between the AMO and winter NAO (Peings and Magnusdottir, 2014), with the AMO leading the NAO by 10-15 years, providing a window of predictability for decadal forecasting (although this result should be treated with caution due to the shortness of the time period used). The negative correlation means an increase in the AMO results in a more frequent negative NAO, (and vice versa for a decrease in the AMO) which in turn reflects the state of the eddy-driven jet.

In summer, the change in phase of AMO from negative to positive in the 1990s was accompanied by anomalously wet summers in northern Europe and hot, dry summers in southern Europe (Sutton and Dong, 2012), leading to the question as to whether the warmer Atlantic Ocean caused this change in climate via changes in the circulation. The NAO index is primarily used in the winter months, but it is possible to define a summer NAO, which is characterized by a more northerly location and a smaller spatial scale (Folland et al., 2009). The summer NAO is shown to exert strong influence on northern European summer climate and its variations are, in part, linked to the AMO. A summary of some of the potential drivers of the winter and summer NAO and their seasonalities is shown in Fig. 2.8.

Whilst an understanding of the processes affecting the NAO is important, ultimately, the NAO is a zonal index that reflects changes in both the jet speed and latitude. However, both jet speed and latitude appear to have different seasonal cycles and interannual variability (Woollings et al., 2014) and so the NAO index cannot capture the altogether

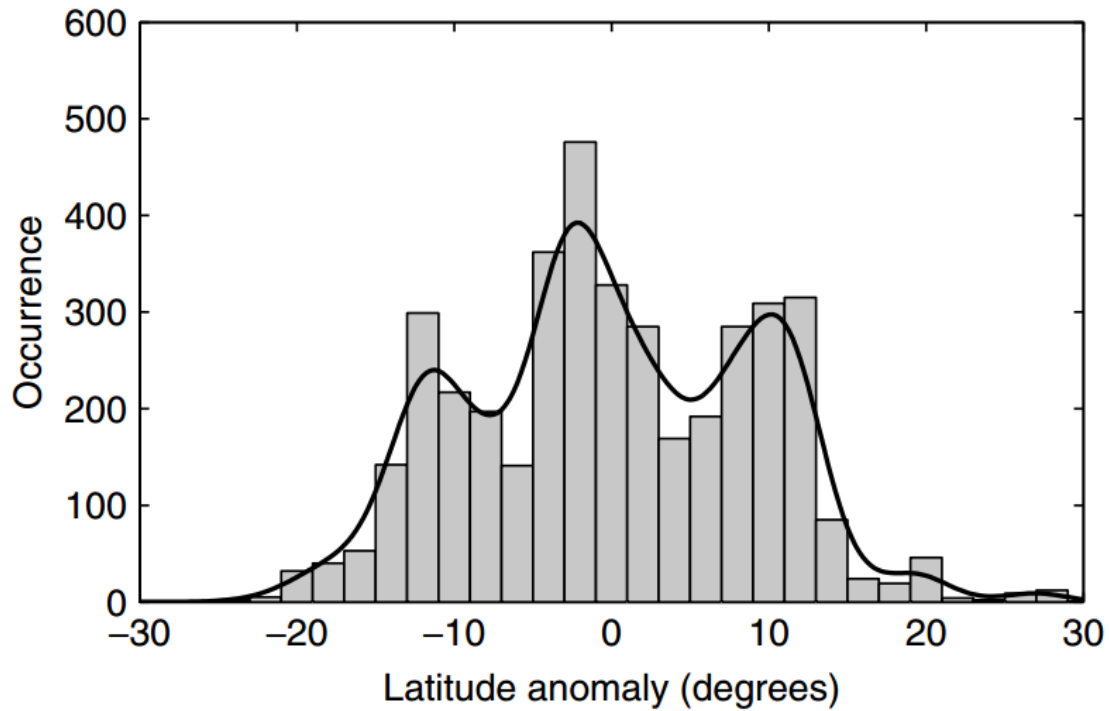


Figure 2.9: Distribution of the daily DJF eddy-driven jet latitude anomalies in ERA-40 (averaged over 0-60°W and 925-700 hPa). Figure reproduced from Woollings et al., 2010.

more complex variability displayed by the eddy-driven jet. In this thesis, we will therefore begin by investigating the eddy-driven jet speed and latitude separately so that we can understand the physical mechanisms affecting the circulation sensitivity to forcing. We will then look at the NAO and see to what extent it captures the jet changes, and whether the physical mechanisms explaining jet changes also apply to the NAO.

In winter, the North Atlantic eddy-driven jet is unique in exhibiting a trimodal distribution in latitudinal position (Woollings et al., 2010). The distribution is shown in Fig. 2.9; the southern location corresponds to the negative phase of the NAO, and the central and northern locations closely resemble the positive and negative phases of the EA pattern respectively (rather than more positive NAO values). These phases coincide with the occurrence of blocking over Greenland, an undisturbed state, and blocking over southwest Europe respectively. Blocking further north and east within Europe is decoupled from the latitude of the eddy-driven jet. The mean latitude and mean speed

of the jet undergo seasonal cycles, with considerable variation in each month as well as between months (Woollings et al., 2014). Variations in mean latitude are small, with the mean monthly positions separated by only 5° over the year, with the most northerly mean in September and southerly mean in March. Jet speeds display a 4 ms^{-1} variation over the year, peaking in January before falling rapidly to the minimum in May. The spring distribution of jet latitude displays a similar trimodal distribution to winter, but with different amplitudes in the peaks. Summer and autumn appear more bimodal, but their peaks are in the same locations as the trimodal distributions, suggesting the preferred jet positions are related to physical features such as coastlines or orography. Considering jet variability and trends on a longer timescale, reanalyses show a slight poleward shift in jet latitude (Woollings et al., 2014), and an increase in jet speed (Strong and Davis, 2007), although these results must be treated with caution due to uncertainty in the early historical reanalysis data.

2.4 Teleconnections

The NAO is an example of an atmospheric teleconnection, that is to say, a long distance relationship between atmospheric anomalies. The NAO, Pacific-North American (PNA) pattern, and Southern Oscillation are three of the most well recognised teleconnection patterns which manifest in correlated variations in sea level pressure between remote centres of action. Early work by Exner (see retrospective in Hurrell and Deser, 2015) identified the NAO teleconnection pattern, and further work summarised in Walker (1928) demonstrated the existence of the Southern Oscillation and a North Pacific Oscillation in sea level pressure data. Expanding on this, Wallace and Gutzler (1981) identified numerous Northern Hemisphere teleconnection patterns in the 500mb geopotential height field. Compared to sea level pressure teleconnection patterns, they found these teleconnections to be more regional and wavelike with multiple centres of action, resembling forced stationary waves on a sphere. Blackmon et al. (1984a) inves-

tingated the horizontal structure of 500mb height fluctuations on three timescales, 2.5-6 days, 10-30 days, and 30 days plus. Their results confirm those of Wallace and Gutzler (1981), that the long timescale teleconnection patterns represent discrete standing oscillations, such as the NAO, at geographically fixed centres. In contrast, the 10-30 day fluctuations, or intermediate variability, is dominated by zonal wavetrains (wavenumbers 4-6) with no preferred geographical location, but propagating along waveguides originating in jet entrance regions. This intermediate variability exhibits the characteristics of Rossby wave dispersion (Blackmon et al., 1984b).

These intermediate timescale teleconnections can be communicated via Rossby waves which propagate in the waveguides, regions of strong zonal wind shear occurring in the vicinity of jets (Hoskins and Ambrizzi, 1993). These patterns can extend circumglobally, and in doing so include a north-south dipole closely resembling the NAO (Branstator, 2002), suggesting that the NAO may be a local manifestation of a hemispheric wide mode. A key mechanism by which the tropics can affect the extratropics remotely in the form of a teleconnection is via tropically forced Rossby waves, as demonstrated in a baroclinic model by Hoskins and Karoly (1981). The vertical and horizontal structure of tropical heating affects the divergent flow aloft, whose interaction with the mean vorticity gradients represents the source of Rossby waves (Sardeshmukh and Hoskins, 1988). Tropical heating is balanced by vertical motion and strong divergence in the upper troposphere, creating vorticity which generates the Rossby waves (Hoskins et al., 1977; Hoskins and Karoly, 1981). A distinct potential mechanism was demonstrated by Simmons et al. (1983) and Sardeshmukh and Hoskins (1988) using linear barotropic models. They associated the extratropical response to the fast growing mode of tropical instability that excites local anomalies through barotropic instability and energy conversion. Regardless of the distinct forcing mechanism, the fact that wave propagation triggered by tropical convection can exist is clear, with waves able to propagate from the tropics into the midlatitudes.

The Madden-Julian Oscillation (MJO) is an example of a tropical-extratropical tele-

connection in natural climate variability (Madden and Julian, 1971) where tropical forcing impacts the NAO (Cassou, 2008; Lin et al., 2009). The MJO is characterised by an envelope of organized convection confined to the equatorial region that propagates eastward around the globe. In certain phases, the changes in divergent flow associated with the convection can force Rossby waves which propagate into the extratropics and modulate the weather there (Seo et al., 2016). Enhanced understanding of these wavelike teleconnections is important for seasonal forecasting; recently, tropically forced Rossby waves have been shown to contribute to NAO forecast skill (Scaife et al., 2017). We will return to the idea of tropically forced Rossby waves affecting extratropical circulation in Chapter 6, but for now, we just note that there are teleconnection mechanisms by which the extratropics can be remotely influenced from the tropics.

2.5 Forcing the circulation

The eddy-driven jet and associated storm track contribute to much of the weather and climate in the midlatitudes. An active and important area of research is how the North Atlantic eddy-driven jet will respond to anthropogenic forcing. Changes in the mid-latitude eddy-driven jet latitude are one of the most robust circulation signs of climate change. The consensus amongst current climate models is a modest change in the position and strength of the North Atlantic eddy-driven jet; a poleward shift in summer, and a strengthening and lengthening in winter (Woollings et al., 2012; Simpson et al., 2014). This has been found in numerous idealized modelling studies and sensitivity studies, discussed in detail in this section. However, it is necessary to impose tighter constraints on models, in particular constraints on ocean circulation to reduce uncertainties in the response. On top of this, modelling studies exhibit large spreads, particularly in winter; model responses are often weaker than their biases when compared to reanalysis and can be sensitive to the methodology used (Woollings and Blackburn, 2012).

The IPCC Fifth Assessment Report states that, under Representation Concentration

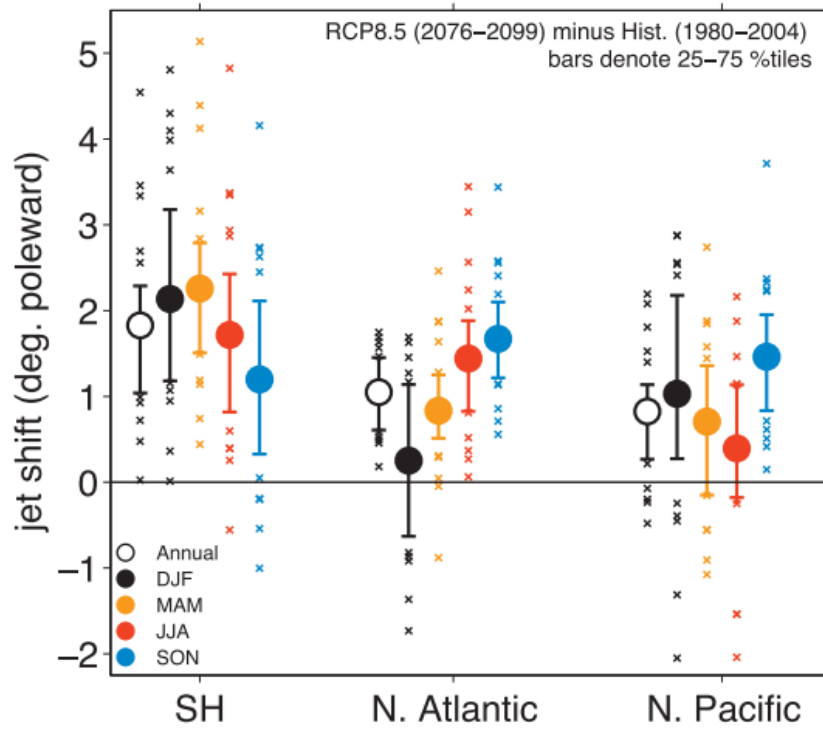


Figure 2.10: Seasonal-mean jet shift (degrees poleward) between the Historical and RCP8.5 CMIP5 experiments for the three sectors. The bars denote the 25th–75th percentile range of the multimodel spread (22 models total) and the diagonal crosses denote the models outside of this range. Figure reproduced from Barnes and Polvani, 2013.

Pathway 8.5, poleward shifts of the eddy-driven jets are likely by the turn of the twenty-first century (IPCC, 2013). In the Coupled Model Intercomparison Project (CMIP5), the majority of models find a poleward shift of the North Atlantic eddy-driven jet (Simpson et al., 2014), see Fig. 2.10, but no robust response in jet speed. The variability in jet position is found to decrease in the CMIP5 models as the jet shifts poleward, coupled with a modest increase in variability in jet speed (Barnes and Polvani, 2013). However, the magnitudes of these changes are uncertain (Grise and Polvani, 2016), and the mechanisms unclear. This is especially true in the North Atlantic, where compensating effects and greater natural variability make it harder to attribute changes to anthropogenic forcings. Jet changes (both forced and due to internal variability) are a significant source of uncertainty in predictions of future regional climate (Karpechko, 2010; Deser et al., 2012). Our inability to project the change in the jets thus limits our ability to quantify

the effect of anthropogenic and natural forcings on the jets. Therefore, the degree to which weather and climate will be affected over regions such as northwestern Europe, and the real societal impact on those living in the midlatitudes, cannot yet be quantified. This makes it imperative to understand the response of the jets to forcing.

2.5.1 Idealized modelling studies

Idealized models are heavily simplified models which allow experiments to be conducted without being affected by the choice of parameterization and so allow the dynamics to be focused upon. The dynamics are often simplified too, which can be used to determine the robustness of results across a hierarchy of models, and also to identify key processes which may be missing in simplified models. The dynamics can be investigated under different background states, which are all simply set up in the general circulation model (GCM). The dry dynamical core set-up first pioneered by Held and Suarez (1994) is the go-to model for idealized modelling experiments (e.g. Polvani and Kushner, 2002; Lorenz and DeWeaver, 2007; Butler et al., 2010; Simpson et al., 2010; McGraw and Barnes, 2016). The model uses a simple Newtonian relaxation of the temperature field to a zonally symmetric state, and a Rayleigh damping of low-level winds to represent boundary-layer friction. The simplistic set-up makes it possible to perform long runs without being computationally intensive.

The response to anthropogenic climate change is multifaceted, involving complex interactions between many different systems. In general, moist processes should be considered when studying the eddy-driven jet (Shaw and Voigt, 2015) and associated storm track (Shaw et al., 2016). However, since Mbengue and Schneider (2013) show that 80% of the storm track shift seen in CMIP5 simulations can be reproduced using a model with only dry dynamics, much can be learnt about the response of the eddy-driven jets using a dry idealized model. Dry dynamical models are useful tools for understanding the response of the jets to heating, which in more complex systems may arise due to moist

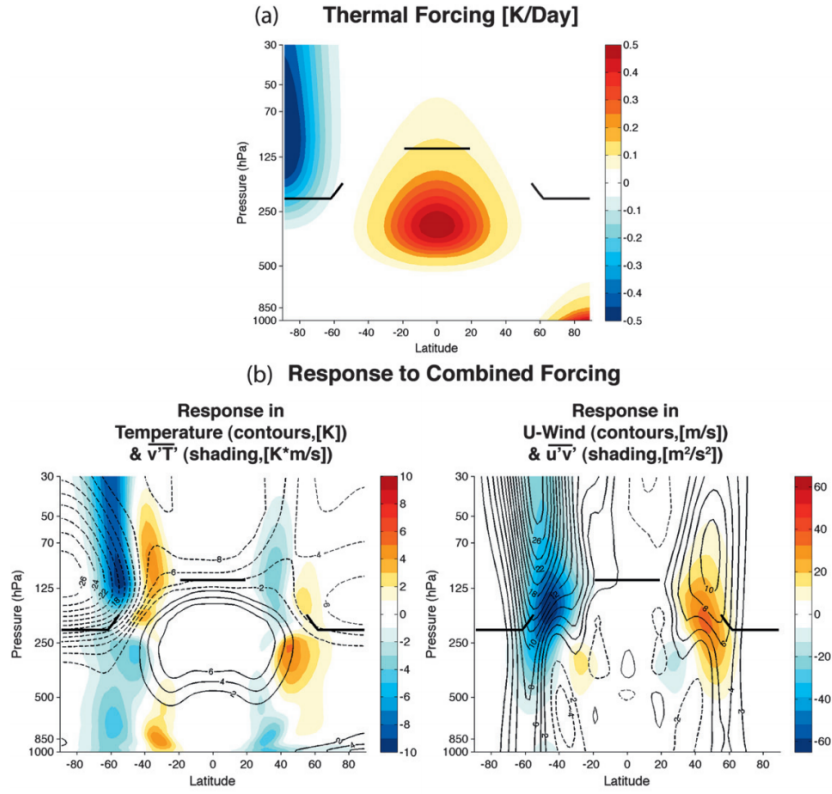


Figure 2.11: Steady state response to heating in a zonally symmetric dry idealized GCM: (a) The combined forcing. (b) The responses to the combined forcing. Bold black lines represent the control run tropopause height. (left) The total eddy heat flux response (shading) and temperature response (contours). (right) The total eddy momentum flux response (shading) and the zonal-mean zonal wind response (contours). Figure reproduced from Butler et al., 2010.

processes. Previous work has shown that idealized dry dynamical models (e.g. Lorenz and DeWeaver, 2007; Butler et al., 2010; Mbengue and Schneider, 2013; McGraw and Barnes, 2016) do respond similarly to fully complex climate models (e.g. Barnes and Polvani, 2013; Simpson et al., 2014; Harvey et al., 2015). For example, Butler et al. (2010) find both the poleward shift and strengthening of the jet in response to expected anthropogenic forcings (Fig. 2.11). This suggests that an idealized model could be used to further investigate the sensitivity of the jets to forcing in different regions.

The northern annular mode is a hemispheric teleconnection pattern of sea level pressure anomalies centred at the poles, commonly defined as the first EOF of sea level pressure over the whole Northern Hemisphere. Many previous studies focus on the response

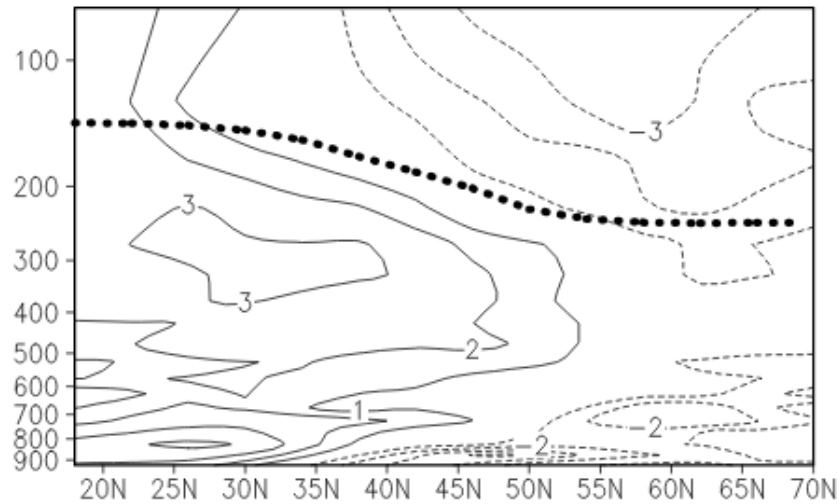


Figure 2.12: Zonal-mean zonal wind response (relative to a control run) of idealized dry GCM experiments with applied, localized heating. The zonal wind response is projected on EOF1 of the control run as a function of both latitude and pressure. The applied heating is 20° latitude wide and 150 hPa deep. The units are the maximum EOF1 response in ms^{-1} (i.e., the amplitude of zonal wind response at the centres of action of EOF1). The tropopause is plotted with a thick dotted line. Figure reproduced from Lorenz and DeWeaver, 2007.

of the northern annular mode to forcing (e.g. Lorenz and DeWeaver, 2007; Ring and Plumb, 2008), see results from Lorenz and DeWeaver (2007) in Fig. 2.12. However, this fails to isolate latitude jet shifts from the changes in jet speed (Monahan and Fyfe, 2006; Woollings et al., 2010). To distinguish between these two changes it is necessary to use two distinct indices, one for jet speed and one for jet latitude. An alternative approach would be to use two or more EOFs (Vallis et al., 2004; Sheshadri and Plumb, 2017). Studies that do separate the speed and latitude of the jet do so by applying forcings that are meant to mimic the expected forcings from anthropogenic climate change (e.g. Son and Lee, 2005; Butler et al., 2010; Allen et al., 2012; Wang et al., 2012; Sun et al., 2013; Tandon et al., 2013; McGraw and Barnes, 2016; Burrows et al., 2017). The shortfall here is that the exact anthropogenic forcing in the atmosphere is not known. This makes the results found in these studies highly dependent on the exact location and magnitude of the forcing applied. Sensitivity studies are important in this regard, as with a broader range of forcings, greater understanding of the possible range of re-

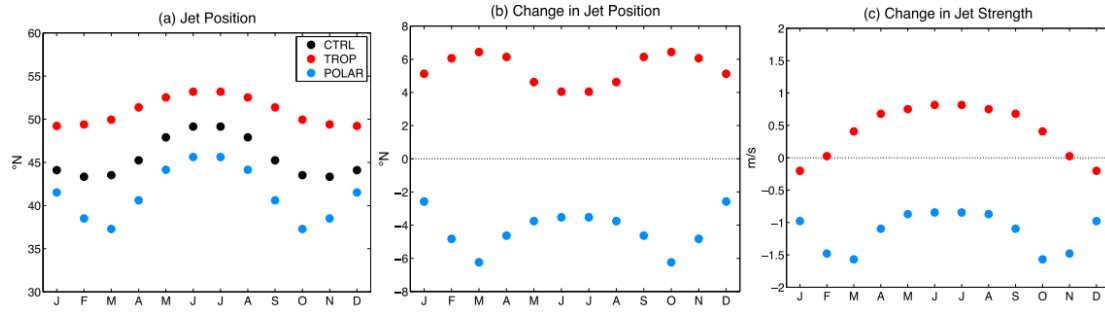


Figure 2.13: (a) Jet position, (b) change in jet position, and (c) change in jet strength for tropical and polar heating experiments. Figure reproduced from McGraw and Barnes, 2016.

sponses can be determined. In addition to climate change-like forcings, quantifying and characterising jet sensitivities is also useful in understanding the response of the jets to natural variability such as the Atlantic Multidecadal Oscillation (Sutton and Dong, 2012), El Niño–Southern Oscillation (Lu et al., 2008) and in assessing the sensitivity to model biases in the position of diabatic heating (Hawcroft et al., 2016). More generally, heating experiments can also be used in understanding the role of diabatic heating in modifying the atmospheric mean state (Hoskins and Valdes, 1990) and its variability (e.g. Woollings et al., 2016).

The response of the jet is not uniform across all seasons. Using an idealized model, McGraw and Barnes (2016) investigate the seasonality of the response of the jet to tropical and polar forcings, finding the response of the latitude to be strongest and non-linear in transition seasons, but approximately linear in winter months (Fig. 2.13). Butler et al. (2010) also note differences between equinoctial and winter responses, with a reduced shift, but greater speed increase in winter. Seasonal variations are also found in CMIP5; Simpson et al. (2014) show that the North Atlantic jet exhibits robust shifts in March–April–May, June–July–August and September–October–November, but there is no model consensus on a shift in December–January–February, but on a strengthening and lengthening of the jet instead. These changes in the jet appear to have an important contribution from changes in low-frequency/stationary waves but the mechanisms are not yet well understood.

The Green's function type approach is a powerful tool in building up the spatial sensitivity pattern to forcing. Hassanzadeh and Kuang (2016) employ this in the atmosphere to study the circulation response to thermal and mechanical forcing. The sensitivity map informs how the circulation responds linearly to forcing in different locations. From the sensitivities, it is possible to determine what forcing is necessary to cause a certain atmospheric state, which has an immediate use for attributing changes to specific forcing distributions. As an example, the forcings required to produce a poleward shift in the jet can be determined and compared to anthropogenic climate forcings to see if the jet shift can be explained by these forcings. Another key application is in isolating eddy-feedbacks to determine a mechanistic understanding of how eddy-mean flow interactions cause, for example, a poleward shift of the midlatitude jets under climate change.

Slightly more complex models include SSTs to investigate how their changes can affect the jet. Through aquaplanet simulations in equinoctial conditions, Brayshaw et al. (2008) find that increases in midlatitude SST gradients lead to stronger, poleward shifted, storm tracks. However, for increased SST gradients close to the subtropical jet, the eddy-driven jet shifts equatorward as the Hadley cell and subtropical jet strengthen. If the SST gradients are decreased near the subtropical jet then the storm track shifts poleward and a well separated eddy-driven jet occurs. The difference in responses implies that SST anomalies close to the latitude of the subtropical jet are very effective at modifying the storm track. These conclusions are supported by the work of Graff and LaCasce (2012), who perform similar experiments by increasing SSTs by 2 K in specific latitude bands. A uniform global 2 K SST increase strengthens the Hadley cell and the subtropical jets, and shifts the storm track poleward.

Spin-up ensembles can be used to determine the mechanism of response to a range of forcings. By initialising a 240 member ensemble with anomalous Northern Hemisphere sea ice extent, Deser et al. (2007) investigate the transient response of the circulation to SST and sea ice perturbations in winter, showing that the initial baroclinic response grows to a hemispheric scale equivalent barotropic response. Strong and Magnusdot-

tir (2010) used the same simulations to identify the mechanism by which the response manifests. They see changes in the development of a localized baroclinic response which leads to the dominant eddies along 55°N shifting to lower phase speeds, decreasing anticyclonic Rossby wave breaking (RWB). In contrast at 35°N , eddy phase speeds increase and anticyclonic RWB increases. This change in RWB relative frequency forces the background zonal flow, with the response in RWB accounting for over 80% of the spatial and temporal pattern of the zonal wind response. Using a 200 member spin-up ensemble in a simple aquaplanet setup, Chen et al. (2013) diagnose the mechanism of storm track response to a uniform 4 K increase in SST. The model responds with a fast poleward shift in midlatitude eddy-driven winds and the edge of the Hadley cell, with a slow equatorward contraction of the upward branch of the Hadley cell and ITCZ. The shift in eddy-driven westerlies is deduced to come as a result of more upward wave propagation, rather than a reduction in lower-tropospheric baroclinicity. A reduction in eddy dissipation of wave activity due to a weaker meridional potential vorticity gradient causes this increase in waves propagating away from baroclinic generation regions, leading to more poleward momentum flux and the shift in eddy-driven winds and Hadley cell edge.

The power of idealized models is in having a simple and efficient way of investigating atmospheric and oceanic dynamics. However, as they are heavily simplified compared to the real world, results from these modelling experiments should act as a guide as to what happens in the real world. Comparison and similarities between more idealized and more complex models can be used to infer the robustness of mechanisms and responses across model hierarchies (Thomson and Vallis, 2018a,b), as some model responses display dependence on the basic-state, particularly in the midlatitudes (Thomson and Vallis, 2018a). To say anything with more certainty requires the use of more complex models which can simulate the real world to a much higher degree of accuracy.

2.5.2 Sensitivity studies in complex models

The Coupled Model Intercomparison Project (CMIP) was established as a standard experimental protocol for studying the output of coupled atmosphere-ocean general circulation models. In response to the commonly used midrange emissions scenario, A1B, taken from the Special Report on Emissions Scenarios, the CMIP3 (phase 3 of the CMIP) models display a poleward shift of the eddy-driven jet (Woollings and Blackburn, 2012). The response is greatest over Europe, downstream of the jet core, signalling an extension of the storm track. However, there is considerable spread amongst the models, with the standard deviation larger than the response everywhere except in the region of greatest response. Many of the models exhibit an equatorward bias in DJF climatology, an Atlantic jet that is too strong and a jet that tends to be too zonal. In CMIP5, the 5th phase, this equatorward jet bias is especially pronounced for the North Atlantic eddy-driven jet (Barnes and Polvani, 2013). As was found in Woollings and Blackburn (2012) for the CMIP3 models, in CMIP5 there is no robust response of jet speed over the North Atlantic. The variability in jet position is found to decrease in the CMIP5 models as the jet shifts poleward, coupled with a modest increase in variability in jet speed. The mechanism for this, suggested by Barnes and Hartmann (2011), is due to a decrease in background vorticity gradient on which Rossby waves propagate due to geometric effects. Concretely, a more poleward jet is closer to the turning latitude associated with the β -effect, so waves turn instead of break. As waves are unable to break on the poleward flank of the jet, the feedback between waves and jet is broken, preventing the jet from wobbling as much.

In response to global warming, SSTs are expected to increase and develop into pronounced patterns (Xie et al., 2010). There is considerable uncertainty as to the sign of the response to SSTs in the eddy-driven jet amongst CMIP3 models, especially in the Northern Hemisphere (Delcambre et al., 2013). Ciasto et al. (2016) also show that in winter, there is uncertainty between different models in how the North Atlantic storm

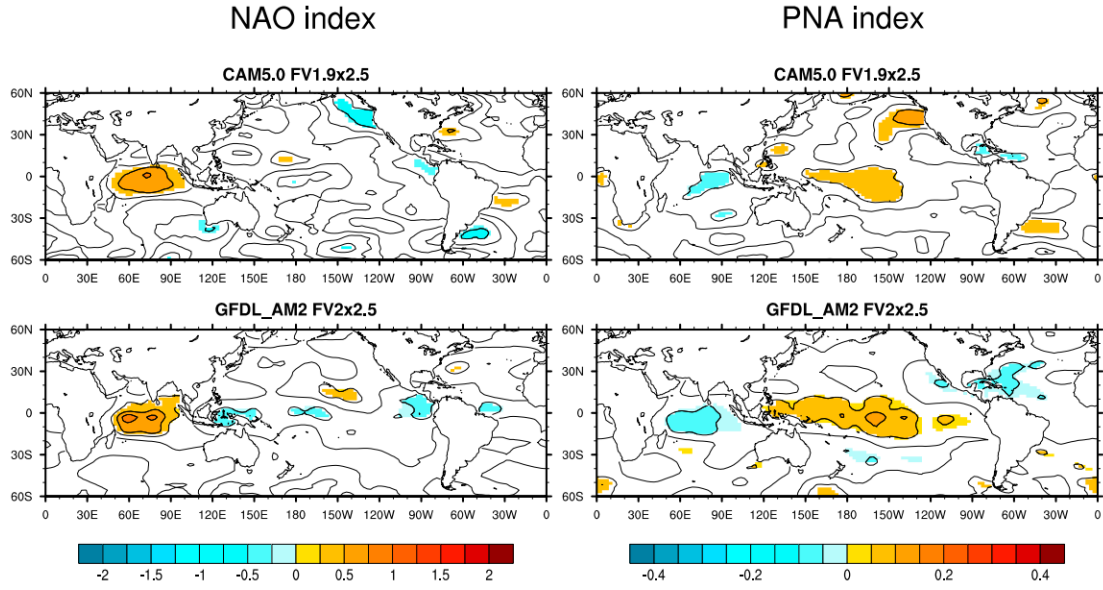


Figure 2.14: Sensitivity of the wintertime (DJF) NAO and PNA indices to the SST anomaly from CAM5 and GFDL AM2. For each model, 200 ensemble members are used to obtain the sensitivity map. Shaded regions refer to where the sensitivity is significant using a two-tailed t-test at 5% significance level. [Units of the sensitivity map of NAO and PNA indices are $(1 \times 10^5 \text{ km}^2 \text{K})^{-1}$]. Figure reproduced from Li and Forest, 2014.

track will change in response to SST changes. They demonstrate that local SST changes fail to reproduce the changes observed in different models, showing that global SST and sea ice changes must be considered to reproduce the responses. However, Haarsma et al. (2015) suggest that in summer the slowdown of the Atlantic meridional overturning circulation (AMOC) will be the main cause of atmospheric circulation changes over western Europe.

A simple method to investigate the sensitivity of atmospheric responses to SSTs is to apply localized SST perturbations and combine them linearly to produce sensitivity maps. Simmons et al. (1983) pioneered the use of a Green's function type approach, systematically applying localized thermal forcings to mimic SST forcings. The individual responses are combined to investigate the sensitivity and linearity of the atmospheric response to SST patterns, in this case, the PNA and EA teleconnection pattern responses. Barsugli and Sardeshmukh (2002) applied 1.5 K patches across the tropical Indian and Pacific Ocean basins and use the Green's function approach to focus on

the response of precipitation. Both studies highlight the importance of tropical SSTs in affecting midlatitude atmospheric phenomena. Barsugli et al. (2006) demonstrate the importance of considering the seasonality of responses, and also that the pattern of SST is of fundamental importance to the response. This is shown by the fact that the response to a uniform 1 K warming of SSTs in the tropics produces a smaller response than to that of the optimal forcing pattern in the tropics. Li et al. (2012) develop an alternative method, the random perturbation method (RPM), by generating random SST patterns instead of applying individual patches. The method is approximately twelve times more computationally efficient to achieve the same standard error in the estimated sensitivity. Using this method, Li and Forest (2014) find the response of the NAO to tropical SSTs is strongly sensitive to the Indian Ocean, as well as exhibiting sensitivity to the central Pacific Ocean and Caribbean Sea (Fig. 2.14). The most efficient basin at forcing the NAO is found to be the Indian Ocean, and the observed NAO time series is significantly correlated with the time series reconstructed from using the three basin's SST anomalies. This heavily implies that the North Atlantic jet is sensitive to these regions in these models, although further work is required to confirm this is the case in the real world.

Extratropical SSTs also play a role in forcing the atmospheric circulation (Kushnir et al., 2002). The response manifests through transient eddies, and is small compared to the internal variability of the atmosphere, meaning that the response is only noticed through changes in the probability distributions of the internal modes of variability. In an analysis of CMIP3 models, Woollings et al. (2012) find half the differences between storm track responses to global warming are associated with uncertainties in ocean circulation changes. The uncertainty of the change in the AMOC is found to have a large impact on the uncertainty of the storm track shift. In winter, Woollings et al. (2012) find a surprising southward shift in storm track, but strengthening and extension of the eddy-driven jet in response to the slowdown of the AMOC. Work by Jackson et al. (2015) confirms this; a slowdown of the AMOC is associated with a strengthening of the eddy-driven jet in winter, with no shift. In summer they find a poleward shifted eddy-driven

jet.

Another potentially important driver of circulation changes in the midlatitudes is sea ice loss. Although we do not investigate this explicitly in this thesis, it is important to be aware of the hypothesised responses. Various studies come to conflicting results regarding the direction of North Atlantic storm track shift with a reduction in Arctic sea ice (Bader et al., 2011; Barnes and Screen, 2015). Reduced sea ice cover in summer/autumn is shown to lead to a negative NAO-like pattern in autumn/winter. This is attributed to a warming in the lower Arctic troposphere with reduced sea ice cover (Deser et al., 2010). A negative NAO should lead to an equatorward shift in the storm track and jet, however further work is required to isolate the storm track shift with a reduction in sea ice cover. Recent modelling studies do find links from Arctic sea ice to the midlatitude circulation, but, due to the short observational record, it is difficult to say if these links exist in the real world (see Barnes and Screen, 2015, and references therein).

It is clear that it is difficult to say with certainty how and why the North Atlantic circulation will respond to the predicted changes due to climate change. Part of the problem arises from the multitude of drivers, many due to anthropogenic forcing, which have competing influences on the circulation. Also, the challenge of accurately and efficiently modelling the coupled ocean-climate system means that it is impossible to say with complete certainty that a particular result from an experiment will manifest in the real world. Building on the studies presented in this chapter, the sensitivity analyses presented in this thesis will provide a useful tool to further aid the understanding and quantification of the future circulation changes.

Chapter 3

Eddy-driven jet sensitivity in an idealized GCM

3.1 Introduction

This chapter investigates the sensitivity of the eddy-driven jet by applying a series of idealized thermal forcings to a simple GCM. We follow Lorenz and DeWeaver (2007) and Hassanzadeh and Kuang (2016) in applying localized perturbations systematically throughout the troposphere. We differ from their studies by evaluating the sensitivity of the different jet characteristics (i.e. speed and latitude) separately, and by applying finite-amplitude forcings on the order of the magnitude of latent heating occurring in the atmosphere. This ensures a high signal-to-noise ratio, but does not guarantee the linearity of the responses. In particular, we will look at the sensitivity of the jet strength, position, and variability to diabatic heating. A large source of uncertainty in the response of the jet to climate change arises from the competing effects of tropical upper tropospheric heating and Arctic amplification. Developing an understanding of the sensitivity of the jet to these competing influences allows the response of the jet to be quantified more accurately.

Idealized modelling studies find the response of the jet is not uniform across all

seasons (e.g. Butler et al., 2010; McGraw and Barnes, 2016). Seasonal variations in jet responses are also found in CMIP5 (Simpson et al., 2014). These changes in the jet appear to have an important contribution from changes in low-frequency/stationary waves but the mechanisms are not yet well understood. In our idealized simulations, the meridional near-surface radiative-equilibrium temperature profile is set so that the resulting temperatures mimic a North Atlantic wintertime and summertime temperature profile in the southern and northern hemisphere respectively.

The focus of this work is on the jet sensitivity to tropospheric forcings. There is evidence that stratospheric dynamics and coupling to the troposphere can influence the eddy-driven jet (e.g. Polvani and Kushner, 2002; Gerber and Polvani, 2009; Son et al., 2009; Manzini et al., 2014). Given the potential complexity of stratosphere-troposphere interactions (Mitchell et al., 2013), a much more realistic model might be needed to assess the stratospheric sensitivities. Here the model does not attempt to reproduce the dynamics seen in the stratosphere, and allows the sole focus to be on the tropospheric dynamics.

The storm tracks are intimately connected via eddy fluxes to the jet. By using eddy flux quantities to quantify the sensitivity of the storm track, greater understanding of how the jet shifts occur is developed. Section 3.2 outlines the methods used to investigate the jet and storm track sensitivities including the model setup and experimental design. Sensitivity results are presented in Section 3.3, along with an in-depth look at four case study simulations to further understand the dynamics behind the sensitivities. A series of diagnostics are used to understand the sensitivities that these simulations produce. This section also explores the linearity of the responses to the amplitude of the forcing, and to combinations of forcings. Some existing theories are tested in Section 3.4, and we discuss how the sensitivities determined by this work support the findings of previous studies. Finally, the conclusions are presented in Section 3.5.

3.2 Methods

3.2.1 Model

Simulations are conducted using the Geophysical Fluid Dynamics Laboratory’s Flexible Modelling System dry dynamical core. The model solves the equations of motion using a spectral transform method at T42 resolution with 37 unevenly spaced sigma levels. Temperatures are relaxed to a zonally-symmetric radiative-equilibrium thermal profile. A quasi-equilibrium convection scheme relaxes temperatures to a prescribed lapse rate $\gamma\Gamma_d$, where Γ_d is the dry adiabatic lapse rate (g/c_p) and γ is a number between 0 and 1. Here, $\gamma = 0.7$ is used. Vertical diffusion of momentum and dry static energy acts within the boundary layer up to 2.5 km. A full description of the model used can be found in Schneider and Walker (2006). The simplistic set-up makes it possible to perform long ensemble simulations without being computationally intensive.

Temperatures are relaxed toward radiative equilibrium on a time scale of 7 days near the surface at low latitudes and 50 days in the rest of the atmosphere. The surface temperature in radiative equilibrium varies with latitude ϕ as

$$T_s^e(\phi) = \tilde{T}_s^e + \Delta_h[\cos^2 \phi + 2 \sin \phi_0(1 + \sin \phi)], \quad (3.1)$$

where $\tilde{T}_s^e = 230$ K is the surface temperature at the south pole, $\Delta_h = 95$ K is the pole-equator thermal contrast and $\phi_0 = 10^\circ$ is the latitude of maximum solar irradiance. We choose $\phi_0 = 10^\circ$ as it gives model surface temperature and zonal wind profiles which represent the observed winter and summer seasonal averages over the North Atlantic. Aside from this altered surface temperature profile, the model setup is that of Schneider and Walker (2006). The radiative-equilibrium profile is detailed below for completeness.

The radiative equilibrium temperature away from the surface is given by

$$T^e(\phi, p) = T_t^e \left[1 + d_0(\phi) \left(\frac{p}{p_0} \right)^\alpha \right]^{1/4}, \quad (3.2)$$

where

$$d_0(\phi) = \left[\frac{T_s^e(\phi)}{T_t^e} \right]^4 - 1, \quad (3.3)$$

and

$$d = d_0(\phi) \left(\frac{p}{p_0} \right)^\alpha \quad (3.4)$$

is a latitude-dependent optical thickness that depends on the near-surface radiative equilibrium temperature $T_s^e(\phi)$ (Schneider, 2004). The constant $p_0 = 1000$ hPa is a reference surface pressure, and the exponent $\alpha = 3.5$, the ratio of the pressure scale height to the partial pressure scale height of the dominant infrared absorber, for example water vapour on Earth, controls the lapse rate of the radiative equilibrium state. The optical thickness is chosen so that the radiative equilibrium temperature T_t^e at the top of the atmosphere is constant. Here, $T_t^e = 200$ K, assuming a fixed longwave emission temperature for a gray atmosphere is proportional to the temperature at the top of the atmosphere (Schneider, 2007).

All simulations use the values for the parameters in Eq. 3.1, which have been explicitly chosen to create a state which mimics the North Atlantic. This creates a background state with a winter hemisphere and summer hemisphere with conditions generally resembling those from reanalysis data. Fig. 3.1 compares the temperature at the lowest model level in a statistically steady state to surface temperature data taken from the NCEP2 Reanalysis (NCEP Reanalysis 2 data provided by the NOAA/OAR/ESRL PSD, Boulder, Colorado, USA, from their website at <http://www.esrl.noaa.gov/psd/>), time averaged from 1979-2015 and averaged over the North Atlantic region 0-60°W, showing they are in good agreement with each other. The North Atlantic state is chosen, rather than a zonal mean state, in order to represent an ocean storm track system. Therefore,

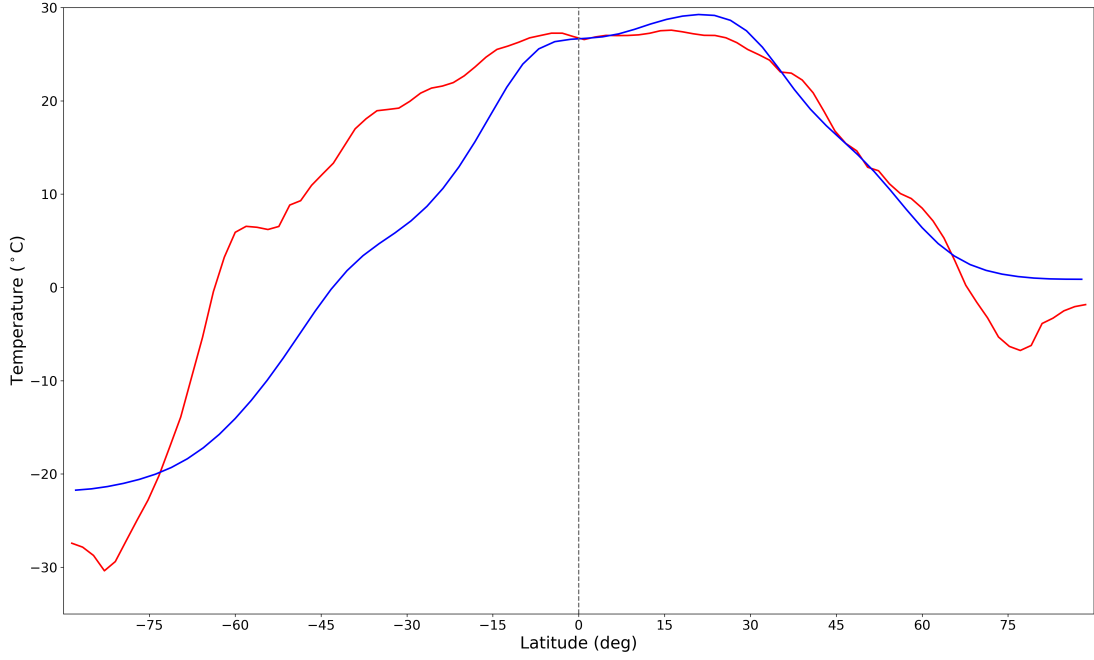


Figure 3.1: Comparison of reanalysis (red) and model (blue) surface temperature. The dashed grey line shows where the winter and summer reanalysis data over the North Atlantic have been stitched together, such that the left half of the figure shows winter data, and the right half shows summer data. Model data is the temperature of the lowest model level taken from a 6-year control simulation. The reanalysis surface temperature data is taken from the NCEP2 Reanalysis time averaged from 1979-2015 and averaged over the North Atlantic region 0-60°W.

whilst the background state resembles the North Atlantic, the results here apply more generally to ocean storm tracks.

The model is run in perpetual winter (southern hemisphere of model) and summer (northern hemisphere of model) conditions with no diurnal or seasonal cycle. Horizontal hyperdiffusion, ∇^8 , damps the smallest resolved scales. In addition, there are no continents or ocean, and no zonal asymmetries to excite stationary waves. Idealized modelling studies with no topography and/or low resolution often produce jets with too persistent annular modes (e.g. Gerber and Vallis, 2007; Gerber et al., 2008; Gerber and Polvani, 2009; Chan and Plumb, 2009). In our setup, the decorrelation timescale (calculated using the e-folding time, as in Baldwin et al., 2003) of the jet indices in the control simulation are: 7.4 and 9.6 days for winter and summer jet latitude respectively. These timescales are on the same order of magnitude as those found in Keeley et al. (2009) for

the NAO (their Fig. 1) of 10 and 5 days for winter and summer respectively.

3.2.2 Control simulation

A spin-up is run for 10 years to allow the model to reach a statistically steady state, then a 6-year control simulation is performed. A comparison of the zonal mean zonal wind, $\bar{u}$, with the reanalysis data, time averaged from 1979-2015 and averaged over the North Atlantic region 0-60°W, is displayed in Fig. 3.2. The model qualitatively reproduces the zonal wind field seen in the reanalysis, with the jets at a similar latitude to the reanalysis; both model jets are stronger than the reanalysis jets, but with similar broad structure. This discrepancy in strength could arise for several reasons. With no ocean transport in the model, the eddies must be stronger than in the reanalysis in order to transport the same amount of heat poleward to reproduce the temperature profile seen in the reanalysis. The winter jet in the model is stronger than observed because the maximum near-surface temperature gradient in the baroclinic zone of the model is larger. There are also no stationary eddies and no moisture in the model. Moisture allows more poleward energy transport through latent transport for the same divergent circulation strength. The lack of topography could also cause this discrepancy in strength, cf. the stronger austral jets to the weaker boreal jets. The subtropical jets are visible on the equatorward flank of the eddy-driven jets. Note the summer jet displays a slight poleward bias in its position, similar to most GCMs including the CMIP suite, which often display poleward biases of up to 5° in the summer jet latitude (Hannachi et al., 2013).

Jet indices are defined for the strength and position of the jet using the maximum of the zonally averaged zonal wind at $\sigma = 0.85$. This height is chosen to be just above the model boundary layer but low enough to separate the eddy-driven jet from the subtropical jet (which has its maximum around $\sigma = 0.2$). The two jet types can be identified in Fig. 3.2; the eddy-driven jets extend down towards the surface, whilst the subtropical jets appear as lobes confined to the upper-troposphere on the equatorward sides of the

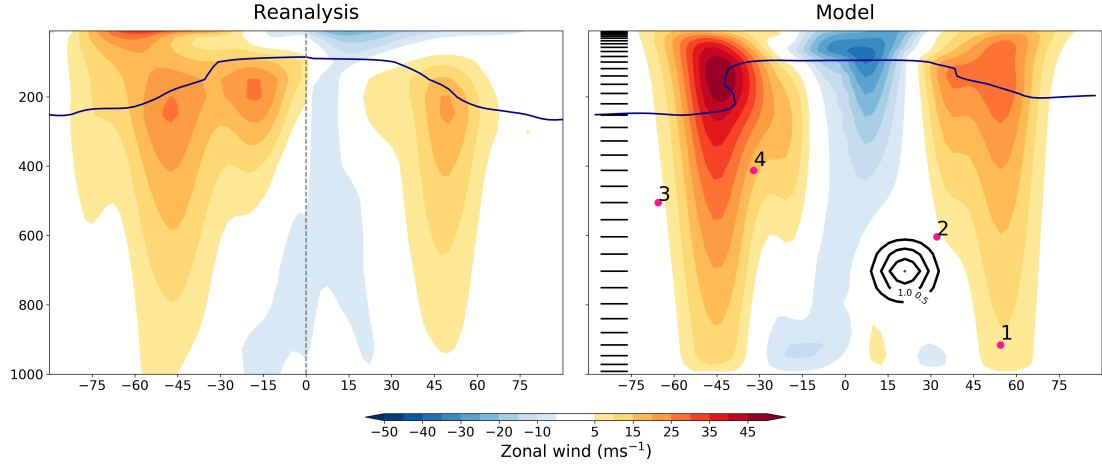


Figure 3.2: Comparison of reanalysis and model $\bar{u}$. The blue line indicates the tropopause, computed as the sigma level at which the lapse rate, $\Gamma = 2 \text{ K km}^{-1}$, the World Meteorological Organization criterion. The dashed grey line shows where the winter and summer reanalysis data over the North Atlantic have been stitched together, such that the left half of the figure shows winter data, and the right half shows summer data. Model data is taken from a 6-year control simulation. The reanalysis data is taken from the NCEP2 Reanalysis time averaged from 1979-2015 and averaged over the North Atlantic region 0-60°W. Also shown in black are the model levels. A sample heating is plotted in the black contours to show the spatial extent of the forcings applied.

eddy-driven jets. The maximum is found by fitting a second order polynomial around the maximum of zonal wind at $\sigma = 0.85$ in each hemisphere, and taking the maximum of the polynomial fit as the jet strength and its latitude as the jet position. Throughout this work we refer to changes in the eddy-driven jet as the changes in the maximum of the zonal mean zonal wind at $\sigma = 0.85$; subtropical jet changes have not been investigated. Storm track indices are also defined using two different measures: $\overline{v'T'}$ taken at $\sigma = 0.85$ and $\nabla \cdot \overline{u'v'}$ at $\sigma = 0.25$ (primed quantities represent deviations from the time mean, the overbar indicates a zonal and temporal average).

3.2.3 Heating experiments

Tropospheric thermal forcings are applied in a series of 306 heating experiments. The forcings are zonally uniform and are applied in the latitude-sigma plane, across 34 different latitudes and at 9 sigma levels, and are held constant over the length of the sim-

ulation. The forcings are applied as a diabatic source term via a Gaussian patch with heating following Hassanzadeh and Kuang (2016), given by

$$\mathcal{H}(\phi, \sigma) = q_0 \exp \left\{ - \left[\frac{(\phi - \phi_0)^2}{2\chi_\phi^2} + \frac{(\sigma - \sigma_0)^2}{2\chi_\sigma^2} \right] \right\}. \quad (3.5)$$

Here, ϕ and σ represent the latitude (in radians) and the sigma level respectively. For each simulation the heating is applied at coordinates (ϕ_0, σ_0) . In all simulations, the magnitude of the heating is given by $q_0 = 2 \text{ K day}^{-1}$, and the spatial size of the heating is determined by $\chi_\phi = 0.12341 \text{ rad}$, $\chi_\sigma = 0.053$ (an example of the heating is shown in Fig. 3.2). The spatial size of the patches used is taken from Hassanzadeh and Kuang (2016), where a series of simulations across the latitude-sigma plane in the linear regime are conducted, in order to build a linear response function. The magnitude of the forcing Hassanzadeh and Kuang (2016) use varies with position to ensure all simulations are within the linear regime. In this work, the magnitude of the forcing is kept constant at 2 K day^{-1} , a similar order of magnitude to Hassanzadeh and Kuang (2016) (who use forcings varying from $0.1\text{-}3.5 \text{ K day}^{-1}$), to ensure the individual experiment responses are all to an equal magnitude of forcing. A similar set of simulations run with $q_0 = 0.5 \text{ K day}^{-1}$ (not shown), produced very similar patterns of sensitivity (see Subsection 3.3.1) as with $q_0 = 2 \text{ K day}^{-1}$. $q_0 = 2 \text{ K day}^{-1}$ is used to give a larger response signal; as shown for specific case studies in Subsection 3.3.2, the response of the jet to varying magnitude of thermal forcing is approximately linear, and so there are no linearity issues with using the larger forcing. Whilst the amplitude of the forcing used here is greater than that used in Butler et al. (2010) ($q_0 = 0.5 \text{ K day}^{-1}$), the spatial extent is substantially smaller, and so the forcings produce temperature changes of a similar maximum amplitude to those of Butler et al. (2010) (not shown).

All simulations are started from the end of the 10-year spin-up period. For each simulation, the patch is then applied and a further year is run before collecting 6 years of model output. From each simulation, $\bar{u}$ is used to calculate the jet indices, and the sen-

sitivity is determined from the difference in the jet indices from the control simulation. Following Woollings et al. (2008), statistical significance in the sensitivity is estimated using the daily jet indices by applying a two-tailed t -test at the 1% significance level. Each time series has an effective sample size:

$$N_i^{eff} = N \left(\frac{1 - \rho_i}{1 + \rho_i} \right), \quad (3.6)$$

where N is the actual sample size and ρ_i is the lag-1 autocorrelation coefficient of the time series. The t statistic is calculated as:

$$t = \frac{\mu_i - \mu_j}{\sqrt{\frac{\sigma_i^2}{N_i^{eff}} + \frac{\sigma_j^2}{N_j^{eff}}}}, \quad (3.7)$$

where μ_i is the mean of time series i and σ_i the standard deviation.

3.3 Results

3.3.1 Sensitivities

Fig. 3.3 summarises all the simulations by showing sensitivities of the jet indices to heating in all regions. The shading at a given point quantifies the change in the jet when heating is applied at that location. The change in the winter jet is measured in the upper panels, and the change in the summer jet in the lower panels. In all figures, the left hand hemisphere (negative latitudes) is the winter hemisphere and the right hand hemisphere (positive latitudes) is the summer hemisphere. For example, when a thermal forcing is applied at $\sigma = 0.8$ and $\phi = 30^\circ$ (a forcing in the summer hemisphere), Fig. 3.3 shows that both the winter jet (top panels) and summer jet (bottom panels) shift poleward, with a strengthening of the summer jet. The control zonal mean zonal wind is contoured for orientation. It is immediately apparent that the latitude and speed of the jet have different

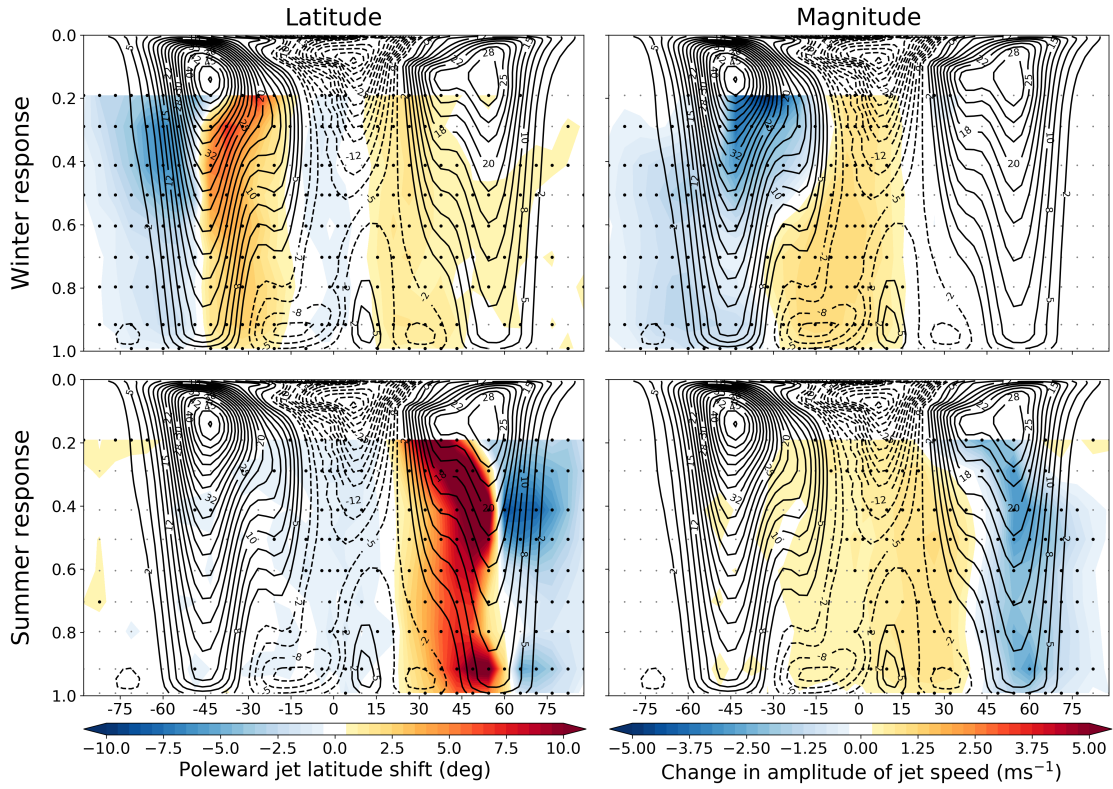


Figure 3.3: Sensitivity of the jet indices to heating experiments (colours) in the latitude-sigma plane. Contours indicate the control $\bar{u}$. Black dots mark experiments where the simulation was significantly different from the control, grey dots where the simulation was not significant. Positive values correspond to poleward shift or strengthening of the jet in the target hemisphere.

sensitivity distributions. This highlights the need to consider the two indices separately unlike in the annular mode and NAO frameworks. For the idealized GCM, the annular mode change (not shown), is mostly reflective of the jet latitude shift.

The largest response in the latitude of the jet in both summer and winter comes from heating on the flanks of the jet core. About the jet core, the response is highly sensitive to the position of the heating; heating equatorward of the core shifts the jet poleward, and heating poleward of the core shifts the jet equatorward. The latitude sensitivity of the summer jet is confined to the region close to the jet, whereas the winter jet shifts poleward due to heating in the subtropics of either hemisphere. The climatological jet position has been shown to impact its response to external forcing (e.g. Garfinkel et al., 2013; McGraw and Barnes, 2016). The climatological jet is further poleward in summer

due to the seasonal cycle, and so the response of the jet latitude is less sensitive to heating in the tropics, but more sensitive to heating at the pole. This is in agreement with Petrie et al. (2015), where a response to sea ice loss in the Arctic is found to drive an equatorward shift in the summer jet, but with no significant change in winter. Changes in jet latitude in winter due to polar heating near the surface still elicit an equatorward jet shift, as found in, for example, Kim et al. (2014) and Sun et al. (2015). It should be noted that under anthropogenic climate change, the heating due to sea ice loss in the Arctic is expected to be larger in winter (Deser et al., 2010), and so whether a greater jet shift might be seen in summer or winter depends on the two factors: jet sensitivity, and amplitude of forcing.

Heating over the pole weakens the jet, although the additional change in jet latitude shows that the response is more complicated than a simple weakening of the jet, as in Francis and Vavrus (2012) for example, who just consider a weakening of the jet with no change in jet latitude. In the deep tropics, heating affects the strength of the jet to a far greater degree than the latitude of the jet, with jet shifts less than 1° until the forcing is further than 15° from the equator. In general, heating equatorward of the midlatitude westerlies and throughout the tropics produces a jet strengthening, and heating in the midlatitude westerlies and further poleward produces a weakening. Note the offset between the change of sign of sensitivity of the latitude and magnitude of the jet; the latitude shift changes sign about the core of the jet, whereas the magnitude response changes sign on the equatorward flank of the westerlies. This suggests that for some range of forcing the jet shifts poleward and weakens.

Heating at different levels generally gives the same sign of response, and the sensitivity often increases with height up to around $\sigma = 0.4$ in the midlatitudes. This could be due to changes in static stability, as heating higher in the atmosphere will increase the stability in the storm track region below the heating. Another reason could be due to near-surface friction affecting eddy amplitudes and leading to differences between simulations with heating in the lower troposphere and the upper troposphere (Held, 2007).

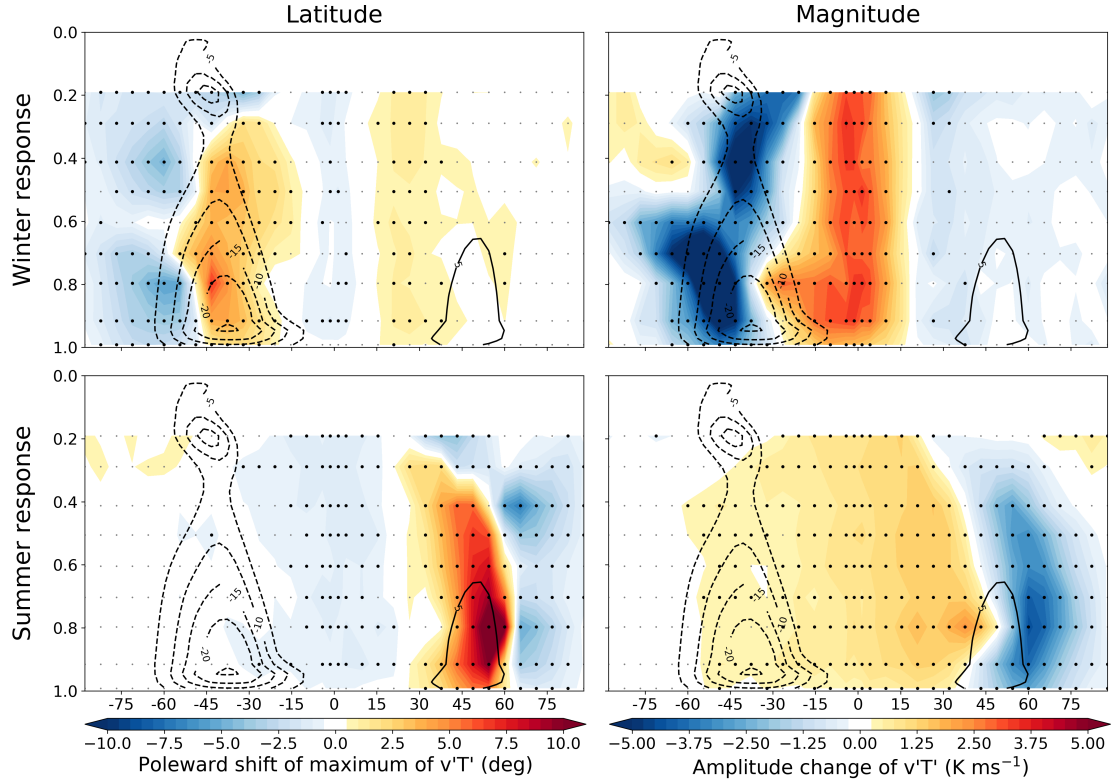


Figure 3.4: Same as Fig. 3.3 but for $\overline{v'T'}$. Positive values correspond to poleward shift or an increasing of the amplitude of $\overline{v'T'}$ in the target hemisphere.

However, neither of these two mechanisms are confirmed, especially as the sensitivity with height mainly affects the latitude shifts and not the changes in magnitudes. Above $\sigma = 0.4$, the heating has the potential to significantly alter the tropopause height, affecting the dynamics. There is also a peak in the amplitude of the response when the heating is applied at $\sigma = 0.9$ near to the jet core, especially prevalent in the summer hemisphere. These sensitivities are explored in Subsection 3.3.2.

To allow comparison with the jet sensitivities, storm track sensitivity maps are plotted using the two measures defined, $\overline{v'T'}$ and $\nabla \cdot \overline{u'v'}$ (Fig. 3.4, Fig. 3.5). There is a broad agreement between the sensitivity of the jet and storm track; for the magnitude of the forcing used here, local heating is needed to affect the storm track position, whereas the magnitude is sensitive to heating in the tropics as well. As with the jet sensitivities, the magnitude of the storm track is more sensitive in winter, and the latitude more sensitive in summer. These contrasts potentially arise from differences in the magnitude of the

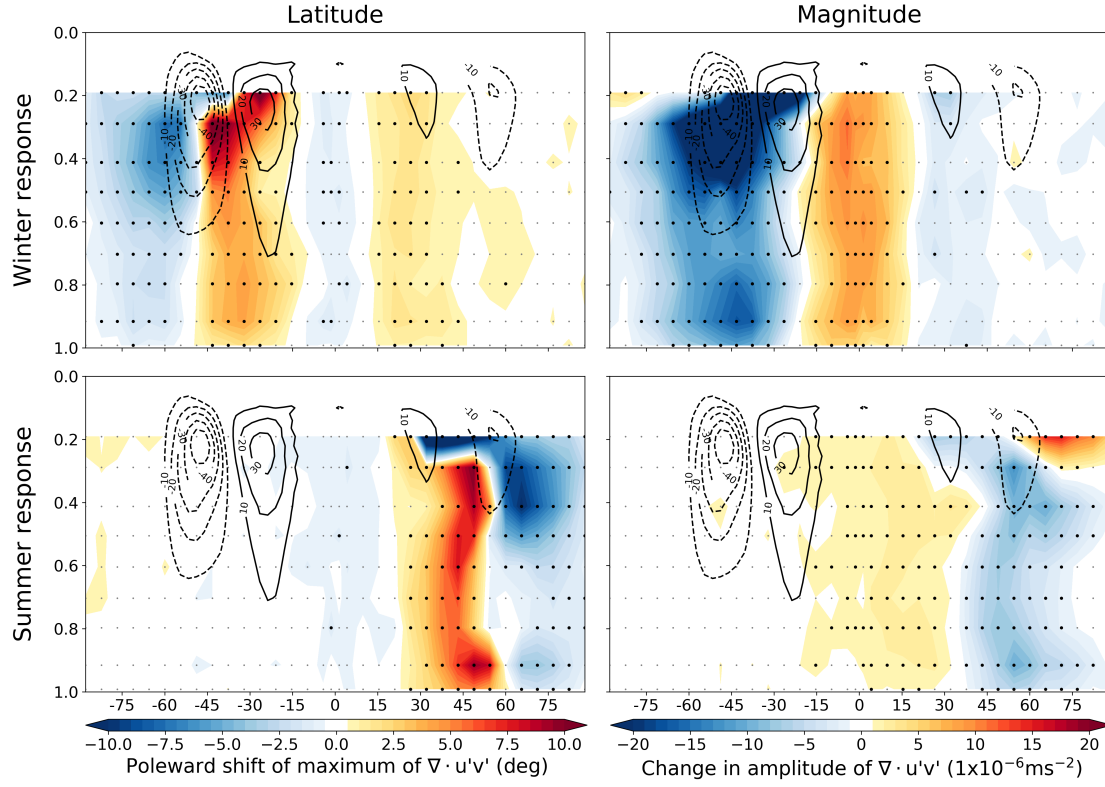


Figure 3.5: Same as Fig. 3.3 but for $\nabla \cdot \overline{u'v'}$. Positive values correspond to poleward shift or an increasing of the amplitude of $\nabla \cdot \overline{u'v'}$ in the target hemisphere.

temperature gradients in the two hemispheres. The climatological maximum of $\overline{v'T'}$ is offset slightly equatorward of the climatological $\overline{u}$ maximum. The maximum magnitude of $\nabla \cdot \overline{u'v'}$ is offset slightly poleward of the $\overline{u}$ maximum. This may help to explain the difference in position of the latitude where the sensitivity of the magnitude and latitude shifts changes sign. Heating on the equatorward flank of the jet lies on the region of maximum $\overline{v'T'}$. Hence heating here acts to push the temperature gradient maximum, and hence $\overline{v'T'}$ maximum, further poleward. Heating either side of the $\overline{v'T'}$ maximum will give changes in the strength of the temperature gradient and hence changes in the magnitude of $\overline{v'T'}$. The strong jet weakening in response to upper level heating on the equatorward jet flank is accompanied by a strong weakening of the storm track. This is likely due to the increased static stability below the heating, which in this case lies directly over the $\overline{v'T'}$ maximum.

The change in variability, or the spread of the distribution of the jet latitude and

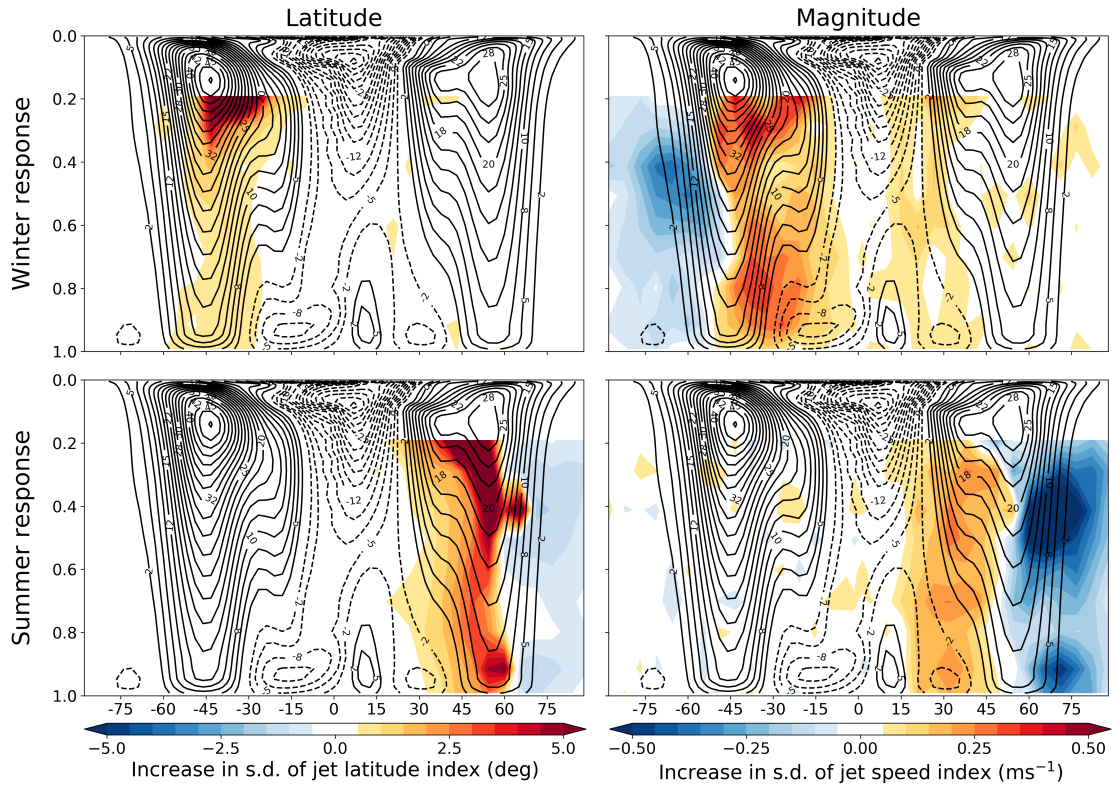


Figure 3.6: Same as Fig. 3.3 but for jet spread. Positive values correspond to an increase in spread of the jet position or strength in the target hemisphere.

speed, is assessed by considering the change in standard deviation of the indices. For each simulation, the daily time series of jet indices are used to calculate the standard deviation of the jet speed and latitude. The change in spread is calculated by subtracting the control standard deviation from the forced simulation standard deviation. Fig. 3.6 shows the sensitivities of the spread of the jet indices in response to the thermal forcings. To affect the jet spread by an appreciable amount, thermal forcing must be applied outside the tropics. The greatest increase in variability comes from heating inside the jet core and on the equatorward flank of the jet, causing an increase in both jet speed and latitude spread. There are no locations where the thermal forcing causes a decrease in the winter jet latitude spread. However for the jet speed in winter, and both latitude and speed in summer, heating on the poleward flank of the jet and over the poles causes a decrease in the spread of the jet latitude and speed.

Barnes and Hartmann (2011) use a barotropic model to show the leading mode of

variability transitions from a shift to a pulse as the jet is shifted poleward, implying a decrease in jet latitude spread. This does not support our findings that the jet latitude spread increases as the jet is shifted poleward. However, Barnes and Hartmann (2011) also show that as the eddy-driven jet moves away from the subtropical jet, the leading mode of variability of the eddy-driven jet shifts from a pulsing mode to a meridional shifting mode. In our simulations, we may be seeing a combination of these two effects influencing the jet spread. McGraw and Barnes (2016) use a dry idealized model to show the spread of the jet strength increases as the jet is shifted poleward, in agreement with our findings. Our findings are also in agreement with, for example, Kidston and Gerber (2010) and Burrows et al. (2017), who note a decrease in persistence time for a poleward shifted jet.

The increase in spread on synoptic timescales may result from increases in the mean meridional temperature gradient or the width of the baroclinic zone. Consider the baroclinic zone width in a simplified case. Take the baroclinic zone as occurring in the region with a strong meridional temperature gradient, with the jet at the latitude of steepest temperature gradient. In general, a heating on the equatorward side of the jet acts to increase the mean temperature gradient and, due to an increase in available potential energy which increases the amount of eddy activity, an increase in the baroclinic zone width. Both the increase in mean temperature gradient and the increase in baroclinic zone width may act to increase the spread of jet latitudes and magnitudes. In contrast, a heating poleward of the jet acts to reduce the extent in the baroclinic zone width and mean temperature gradient, thus reducing the spread of jet latitudes and magnitudes.

3.3.2 Case studies

To investigate the mechanisms behind the shifts and strengthening of the jet discussed above, it is instructive to consider a series of case studies displaying different behaviour due to the location of the thermal forcing (case study forcing locations displayed in Fig.

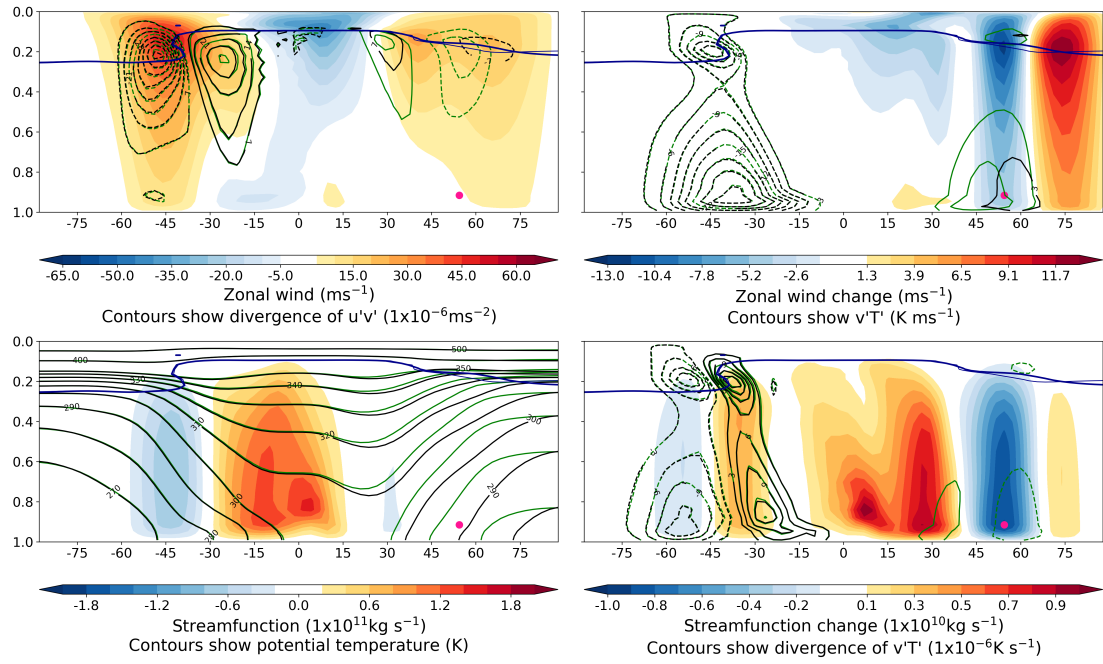


Figure 3.7: Summary plot for heating applied at $\sigma = 0.9$ and $\phi = 54^\circ$ (pink dot). Upper left: simulation $\bar{u}$ (colours), $\nabla \cdot \overline{u'v'}$ (black contours for simulation, green for control). Upper right: change in $\bar{u}$ (colours), $\overline{v'T'}$ (black contours for simulation, green for control). Lower left: simulation streamfunction (colours), potential temperature (black contours for simulation, green for control). Lower right: Change in streamfunction (colours), $\nabla \cdot \overline{v'T'}$ (black contours for simulation, green for control). The thick blue line is the tropopause in the simulation and the thin blue line is the control tropopause.

3.2).

The first of these, displayed in Fig. 3.7, examines the response to near surface heating in the region of large summer jet latitude response (pink dot). Here, the forcing causes a poleward shift of the summer jet latitude, but a decrease in its speed. The thermal forcing weakens the low level meridional temperature gradient in summer, and shifts the location of the steepest isentropes poleward. This acts to weaken $\overline{v'T'}$ and to shift it poleward, causing the weakening and shift of the jet. Changes in jet speed may be mediated via the following mechanism: a decrease in $\overline{v'T'}$ causes a downward Eliassen-Palm flux anomaly which in turn weakens upper level momentum fluxes. The weakened polewards heat transport also acts to set up an anomalous overturning circulation, with the lower branch moving equatorwards. Coriolis then turns this equatorward moving air toward the west, thus weakening the lower level westerly winds and hence the jet, as in

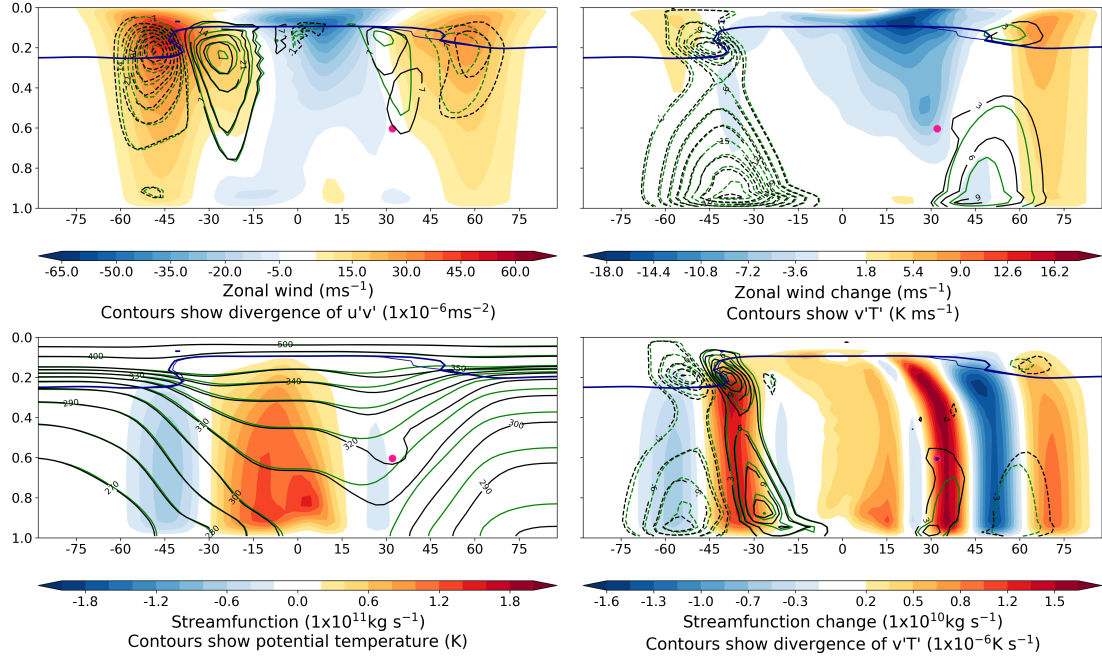


Figure 3.8: As in Fig. 3.7 but for heating applied at $\sigma = 0.6$ and $\phi = 32^\circ$ (pink dot).

Edmon et al. (1980). For an increase in jet speed, a strengthening in $\overline{v'T'}$ sets off the same cycle, but with opposite signs. This forcing is between the maxima of $\overline{v'T'}$ and $\overline{u}$, causing a weakening but poleward shift of the jet. More commonly, a poleward shift of the jet would be associated with the jet strengthening (due to eddy phase speed arguments such as in Chen and Held, 2007). Heating above this level is above the maximum in $\overline{v'T'}$, and so the responses are weaker until they begin to strengthen again up towards $\sigma = 0.4$, as discussed in Section 3.3.1.

In the second example, displayed in Fig. 3.8, the summer jet responds to the mid-tropospheric subtropical forcing by strengthening and shifting poleward, and the winter jet shifts poleward (see Fig. 3.3). The heating is now equatorward of the maximum in meridional heat flux, thus $\overline{v'T'}$ strengthens, causing an acceleration of the jet as described above. The isentropes steepen poleward of the heating, causing the jet to shift poleward to the steeper region. The effect on the winter jet arises as the thermal forcing modifies global isentropes, conveying the shift in jet latitude through the overturning circulation. An increase in inter-hemispheric energy asymmetry leads to a stronger cross

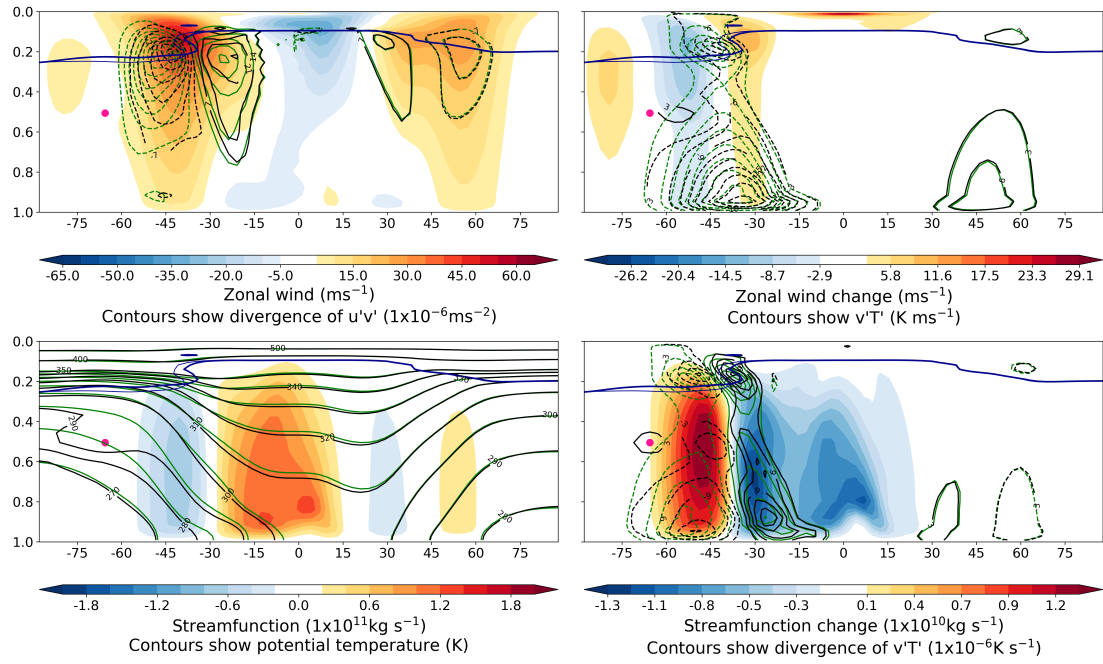


Figure 3.9: As in Fig. 3.7 but for heating applied at $\sigma = 0.5$ and $\phi = -66^\circ$ (pink dot).

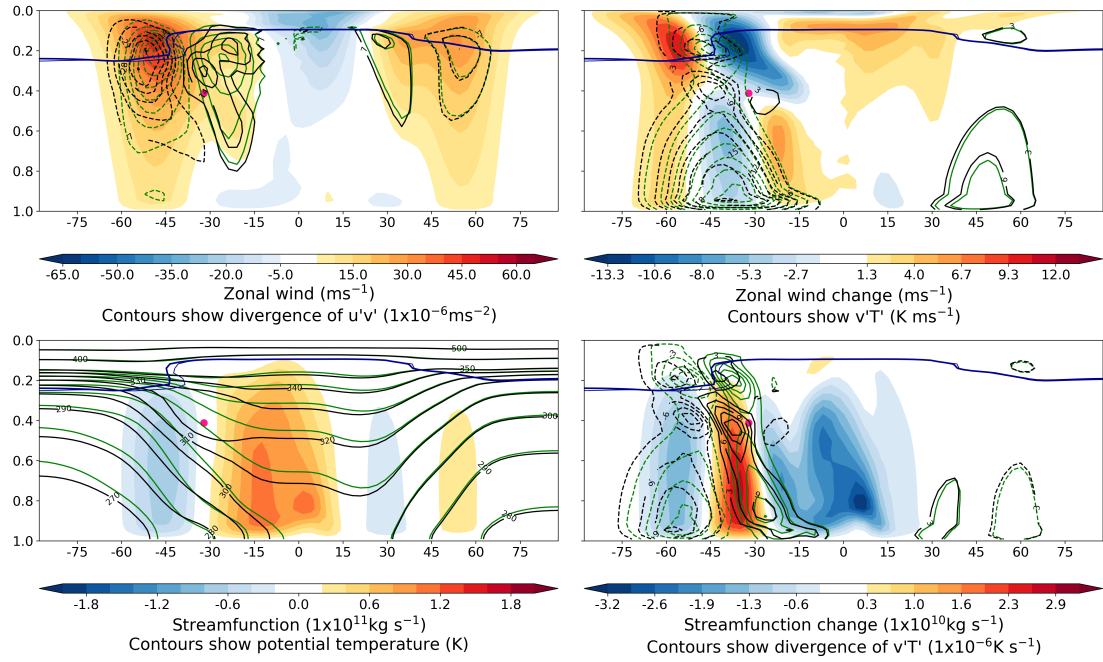


Figure 3.10: As in Fig. 3.7 but for heating applied at $\sigma = 0.4$ and $\phi = -32^\circ$ (pink dot).

equatorial flow, with the ascending branch of the Hadley cell displaced into the summer hemisphere. The Hadley cell, which relaxes temperature gradients in the deep tropics towards zero, communicates changes in the summer hemisphere to the winter hemisphere

through a modification of the temperature gradients in the winter hemisphere. This effect occurs due to sub-tropical heating in summer hemisphere, where the Hadley cell acts to transport the energy into the winter hemisphere. Heating at the same latitude in the winter hemisphere however, does not produce an effect on the summer jet as then the heating would reduce the Hadley circulation strength.

The third case study examines mid-tropospheric heating on the poleward flank of the winter jet (Fig. 3.9). This causes a flattening of the isentropes at the latitude the heating is applied, shifting the maximum of the baroclinic zone equatorward and contracting the baroclinic zone, thus shifting the jet equatorward. The low level temperature gradient across the jet weakens, weakening the eddy transports and thus weakening the jet via thermal wind balance. Heating at the same latitude but higher in the atmosphere (not shown), creates a region of high static stability below the heating, thus compressing the local isentropes which acts to flatten the near surface isentropes causing an equatorward shift of the mean temperature gradient, shifting the jet further equatorward. Heating at the same level, but further equatorward, acts to greatly weaken the low level meridional temperature gradient and the momentum fluxes on the equatorward flank of the jet (not shown), causing a greater weakening of the jet.

The final case study looks at the response to a mid-tropospheric midlatitude forcing on the equatorward flank of the winter jet, Fig. 3.10. A poleward shift in the jet arises due to the latitude of steepest isentropes shifting poleward, which acts to shift $\overline{v'T'}$ poleward. However, the low level meridional temperature gradient weakens, causing $\overline{v'T'}$ to weaken, giving a weakening of the jet via the same mechanism as in the first case study. Thermal forcings at the same level, but moving closer to the equator progressively, act less to weaken the low level meridional temperature gradients. Once past a critical latitude, the thermal forcing begins to act to increase the low level meridional temperature gradients, thus causing the strengthening of the jets seen for forcings in the tropics.

By considering the four case studies, it is apparent that many of the jet shifts seem to be associated with meridional shifts in the maximum low level meridional tempera-

ture gradients, shifting the position in maximum meridional heat flux. Jet speed changes seem to be associated with changes in the magnitude of the low level meridional temperature gradients which lead to changes in $\overline{v'T'}$ relative to the climatological maximum.

Due to the number of simulations, it is unfeasible to examine the mechanisms by which the jet responds in all simulations, and it is possible that other mechanisms are at play in the other heating simulations. All results considered here concern the fully equilibrated response. Some spin-up ensemble simulations, similar to Sun et al. (2013), have been performed for the four case studies using a 200 member initial condition perturbation ensemble (presented in Chapter 5). Heating is applied at day zero and the ensemble mean response at each day is used to investigate the mechanisms of how the jet responds to the heating. Changes in $u'v'$ associated with upper level wind changes are observed. However, the response is apparent earliest in $v'T'$, suggesting that in the ensemble, changes in baroclinic growth are instrumental in initiating the jet response. We explore exactly how these responses are set up in Chapter 5.

3.3.3 Linearity of simulations

In order to determine whether the response of the jet to the forcings behaves linearly, two further sets of simulations are run. The first of these varies the strength of the thermal forcing at each of the four heating locations used in the case studies. q_0 is set to 0.1, 0.5, 1.0, 1.5, 2.0 and 2.5 K day⁻¹, and the resultant changes in jet indices are plotted in Fig. 3.11. In the regime of heating between 0-2.5 K day⁻¹, the response to forcing in both jet speed and position, summer and winter, is approximately linear. It should be noted that for the stronger heatings there are some instances where the response of the summer jet latitude does deviate from linearity.

In the second test of linearity, the forcings applied in each of the individual case studies are applied in all possible combinations and simulated as in the original sensitivity simulations. Fig. 3.12 displays the change in jet index in each combined simulation

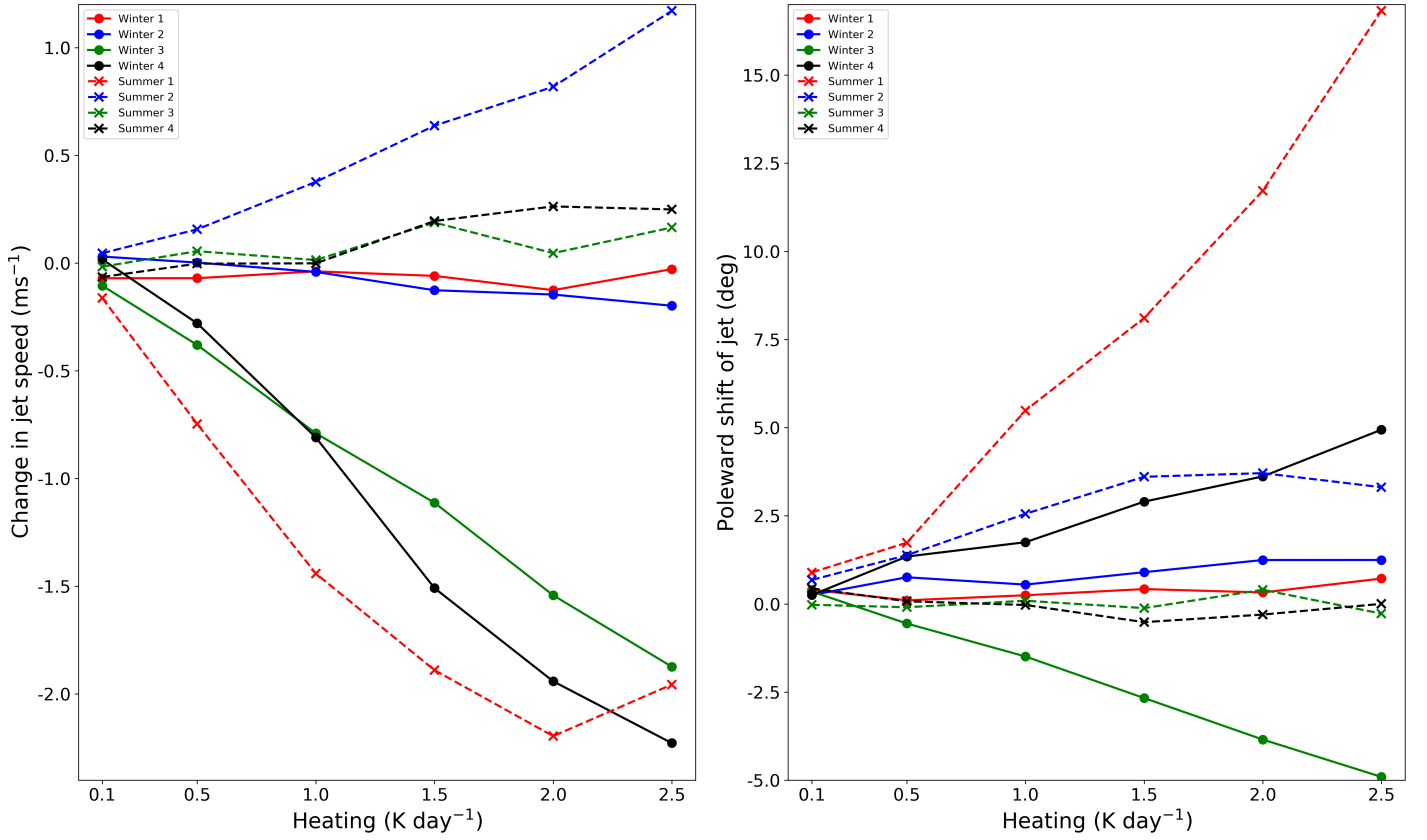


Figure 3.11: Linearity of response with respect to the amplitude of forcing. Left panel: change in jet speed for each of the four case studies in winter and summer as q_0 varies. Right panel: poleward jet shift for each of the four case studies in winter and summer as q_0 varies. Solid lines indicate winter jet responses, dashed lines indicate summer jet responses.

against the sum of the changes in jet indices from the respective individually forced simulations. Aside from the summer responses in simulations that include the forcings in the first two case studies, the results are also approximately linear. The responses to simulations that do contain the forcings from the first two case studies display smaller amplitude responses than in their respective linear combinations which is possibly due to the proximity of the two forcings, and the similar mechanism by which they affect the summer jet, or possibly as the limit of linearity has been reached. Note of course, only a few test cases have been combined in this work, and in some scenarios we are combining forcings from opposite hemispheres.

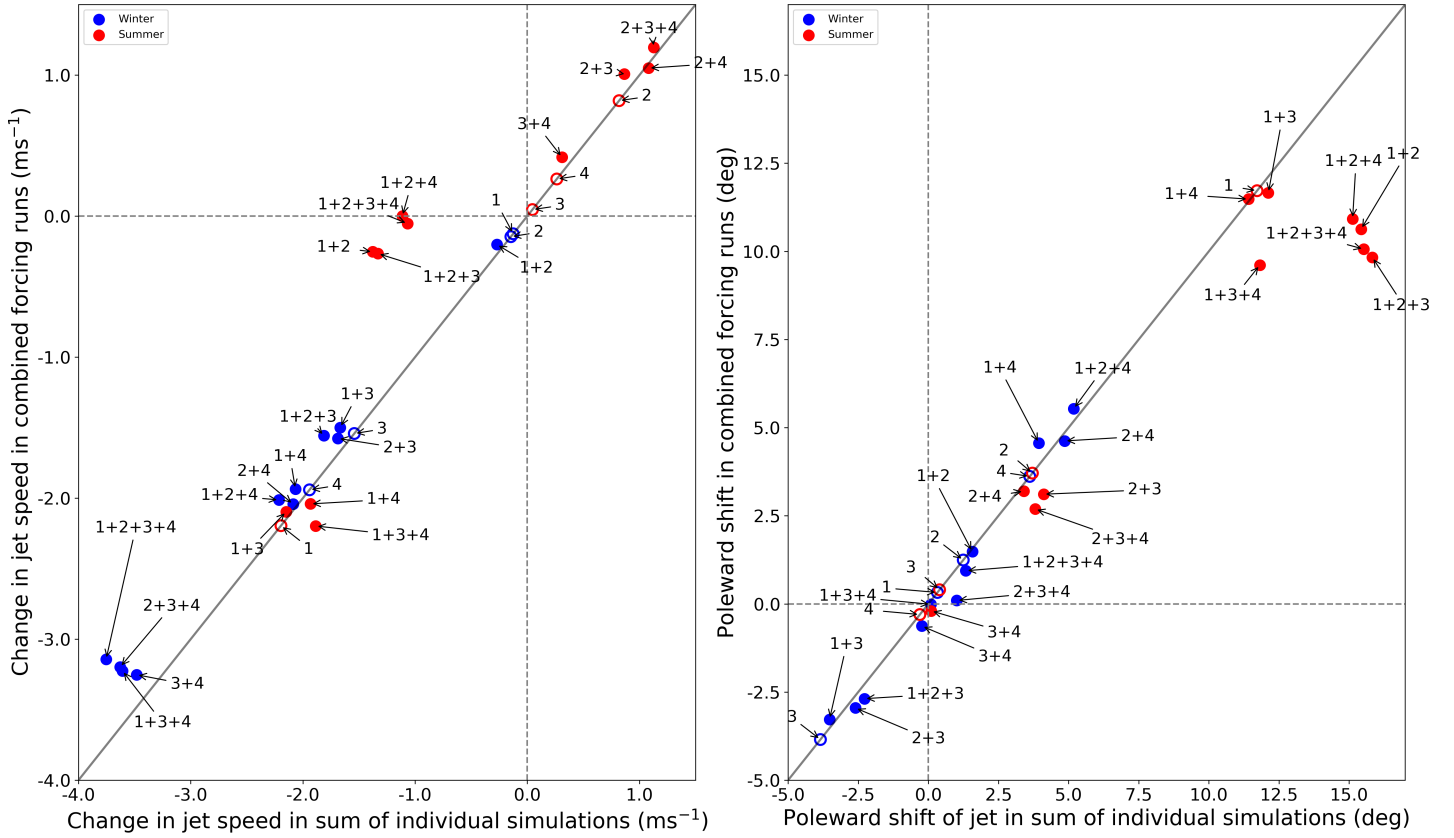


Figure 3.12: Linearity of response with respect to the combination of forcings. Left panel: scatter graph showing the change in jet speed in simulations where the forcings used in the case studies are combined against linear combinations of the results from each individual case study. Right panel: scatter graph showing the poleward jet shift in simulations where the forcings used in the case studies are combined against linear combinations of the results from each individual case study. Blue dots indicate winter jet responses, red dots indicate summer jet responses. Open dots indicate the original case study responses. The labels indicate which case study forcings have been combined at that particular point. In all simulations, $q_0 = 2.0 \text{ K day}^{-1}$.

These linearity tests demonstrate that, for the case studies tested, the jet response to forcing scales approximately linearly with the strength of the forcing, and when forcings are applied in combination. As noted in Tandon et al. (2013), who applied a similar amplitude (but larger spatial scale) forcing, we note that our results do not exhibit a regime change at strong forcing as found by Wang et al. (2012).

3.4 Discussion

It is insightful to compare these results with results from some of the literature in more detail. This section informs on the robustness of results, by comparing this model to others, and also indicates where previous results can be interpreted in the context of the sensitivities described here.

Early work by Son and Lee (2005) used a dry idealized model with tropical heating and polar cooling. Applying heating in a very narrow band across the equator throughout the depth of troposphere produces a marginal equatorward shift of the jet, and a strengthening of the jet (their Fig. 4). Polar cooling acts to shift the jet poleward and marginally weaken the jet. These findings are consistent with the sensitivities found in this work, aside from the jet weakening due to polar cooling (assuming a cooling at the pole produces an equivalent magnitude but opposite sign response to polar heating). Son and Lee (2005) also performed a set of moist runs to investigate the effect of latent heating on the jet, removing their heating and cooling thermal forcing so the latent heating is the only forcing. As much of the effect of moisture manifests in the tropics, the impact is similar to adding the tropical heating thermal forcing. This demonstrates the utility of dry idealized models in developing understanding of the jet response to thermal forcings.

One important result from the sensitivity findings is that the response of the climate to thermal forcing will depend strongly on the position of the thermal forcing. Butler et al. (2010) apply thermal forcings in a dry idealized model, yet find, in direct contrast to Son and Lee (2005), the response to their tropical heating produces a poleward shift of the jet (their Figs. 2, 6, 7). The same is also true of McGraw and Barnes (2016), who find a poleward jet shift due to tropical heating (their Fig. 3). As shown by Fig. 3.3, the poleward extent of the tropical heating has an important effect on whether the shift is poleward or equatorward. If the heating is confined near to the equator, Fig. 3.3 indicates that the response should be an equatorward shift. As the extent of the heating

broadens, it will start to impinge on the zone where a heating will cause a poleward shift, and so the shape of the forcing will determine which of the two competing effects will dominate to determine the direction of shift. Sun et al. (2013) and Tandon et al. (2013) observe similar sensitivities with regard to the direction of jet latitude shift depending on the spatial extent of the thermal forcing. Tandon et al. (2013) suggest that this dependence is due to the Hadley cell edge shifting towards the anomalous diabatic cooling on the flank of the positive thermal forcing. Lu et al. (2008) examine the circulation responses in CMIP3 models to El Niño vs. global warming. The typical El Niño response is a equatorward shift of the jet, as opposed to the anticipated poleward shift and strengthening under global warming. The sensitivities found in this work are again consistent with this difference in response. As the El Niño SST pattern is centred on the equator, with narrow poleward extent, the response seen is due to the weakening in the Hadley cell coupled with an increase in latent heating due to increased precipitation over the equator.

The difference in sensitivity of the speed and latitude highlights the importance of moving away from the NAO or annular mode view to examine circulation changes in the midlatitudes. Lorenz and DeWeaver (2007) use the first EOF of zonal wind to characterize the sensitivity of the jet, and in doing so produce a sensitivity plot that is an amalgamation of the two jet indices (Fig. 2.12). This inhibits the ability to draw conclusions about the changes in the jet and the mechanisms behind them. Despite this, their results do broadly agree with the sensitivities in this work: heating equatorward of the jet causes a poleward shift, heating poleward of the jet causes an equatorward shift (since the annular modes in zonally symmetric models tend to reflect changes in jet latitude primarily). Heating near the surface at 60°N produces a large equatorward shift in the jet, and the largest poleward shift comes from heating at 300 hPa in the subtropics. As shown here, different mechanisms are responsible for driving the different changes in the jet, therefore, to accurately quantify and understand the response, the separate characteristics of the jet must be considered individually, not as an annular mode.

The variation in seasonal sensitivity indicates that the jet latitude is more sensitive in summer, and the jet speed more so in winter. This is demonstrated by the fact that the summer jet displays the largest shift in latitude, and the winter jet experiences the greatest jet speed amplitude change. Although the results in this work are just the sensitivities and not responses, this seems to be corroborated by predictions from the CMIP5 models, which predict a poleward shift of the North Atlantic jet in summer, and a strengthening and lengthening in winter (Barnes and Polvani, 2013). McGraw and Barnes (2016) investigate the jet response to a climate change-like forcing using a dry idealized model. They find that combined high tropical tropospheric and surface polar forcings produce a larger poleward shift in winter and a weakening of the jet in both summer and winter. This could be different to the CMIP5 responses as the positions of the forcings may not be concurrent with those in the CMIP5 models, and it again highlights the importance of understanding the jet sensitivities and the exact thermal forcing due to anthropogenic warming. Understanding the spatial position and extent of the thermal forcing is not trivial; as well as direct forcing due to radiative effects due to CO₂, cloud feedbacks also act to alter diabatic heating in the atmosphere, and this has been shown to have an impact on the jets (e.g. Voigt and Shaw, 2015; Ceppi and Hartmann, 2016; Tandon and Cane, 2017). This highlights the importance of the sensitivity results presented here, as these cloud feedbacks may not necessarily manifest in the tropical upper troposphere and Arctic where the radiative and Arctic amplification warmings occur due to anthropogenic global warming.

The mechanism postulated here acts through the changes in isentropic surfaces, with changes in low level meridional temperature gradients affecting the jet. Brayshaw et al. (2008) investigated the effect of changing the surface gradients through applying SST perturbations. Increased SST gradients in the subtropics caused an equatorward shift and strengthening of the storm track and vice versa for decreased gradients. Increasing SST gradients in the midlatitudes caused a poleward shift and strengthening of the storm track and vice versa for decreased gradients (their Table 2). The strengthening of

the storm track agrees with the results found in this work; by increasing the temperature gradient from equator to pole, the jet strengthens. Brayshaw et al. (2008) attribute the changes due to midlatitude gradients to lower-tropospheric baroclinicity, with the increase in midlatitude gradients strengthening the baroclinicity around the jet and shifting the maximum poleward with the storm track. The equatorward shift of the storm track in Brayshaw et al. (2008) is attributed to a weakening in the tropically driven component of the Hadley cell, thus shifting the maximum in temperature gradient equatorward and the storm track too. This shifting mechanism does not preclude the strengthening of the storm track or jet, which can indeed be seen in certain cases in this work where the jet does shift equatorward. Graff and LaCasce (2012) also perform experiments altering the SST gradients in the tropics and midlatitudes and come to the same conclusion.

The sensitivities determined here can be used to explain results found in more complex models. One of the conclusions from the sensitivity analysis is that heating has to be applied poleward of the storm track in order to weaken the jet, and there is a sharp edge to this region of sensitivity. Conversely, assuming signs can be reversed, a strengthening of the jet could arise due to a cooling, provided this occurs poleward of the storm track. This suggests that the sensitivity of the Atlantic storm track to weakening of the Atlantic meridional overturning circulation (Woollings et al., 2012) might be caused by cooling in the Nordic seas region, rather than in the subpolar gyre further south (see also Harvey et al., 2015). Note also that in this example, regional jet changes could be forced from the tropics via a Rossby wave mechanism (Ciaasto et al., 2016).

The implications of this work reach beyond that of climate change alone. Our results can also aid understanding of jet changes due to natural variability. The sensitivities are also in line with results from studies that use reanalyses to draw insights into changes in the jet. For example, Hall et al. (2016) perform a multiple linear regression to determine drivers of the North Atlantic eddy-driven jet in summer. Many of the predictors they find to be significant (their Table 1), can be explained from the sensitivities determined here. For instance, they suggest that the El Niño (see previous discussion) and

winter AMO (see discussion below) could act to drive the jet equatorward; increases in the Barents-Kara sea ice and increases in the Atlantic SST tripole could drive the jet poleward; and increases in tripole SST and tropical rainfall could strengthen the jet. Increasing sea ice acts as a cooling, therefore a poleward shift agrees with the sensitivity over the pole to thermal forcing as found in this work (see Fig. 3.3). An increase in their tripole SST index increases the equator to pole temperature contrast. As indicated by the mechanisms discussed here, this should strengthen the jet, and could cause the position of the low-level temperature gradient maximum to shift, causing the jet shift. Increases in rainfall implies greater latent heat release, which agrees with the sensitivity found that an increase in thermal forcing in the tropics acts to strengthen the jet.

Focusing more on the AMO, Peings and Magnusdottir (2014) show that an increase in the AMO can force a negative NAO response in winter. This likely reflects an equatorward shift of the jet, as also seen in summer (Sutton and Dong, 2012). There is debate over if and where the AMO does force the atmosphere. Gulev et al. (2013) show that the positive heat fluxes from the AMO correspond to an increase in temperature over the midlatitudes, suggesting that in the midlatitudes, the ocean drives the atmosphere. However, Davini et al. (2015) argue that the effect on the atmosphere from the AMO arises from forcing in the tropics, with little influence from the extratropics. From the sensitivities in this work, it seems that the ocean could force an equatorward jet shift due to warming in the mid/high latitudes on the poleward flank of the jet, or the deep tropics, and not the subtropics/midlatitudes on the equatorward jet flank. Clearly the results from the dry model do not necessarily explain all the facets of the problem, but they do allow insights to be made into where the AMO may be forcing the atmosphere.

Another use for the sensitivities in this work are to aid in assessing sensitivity to model biases in the position of diabatic heating. For example, Hawcroft et al. (2016) identify likely biases in the distribution of latent heating in both the HiGEM model and the ERA-Interim reanalysis. Given uncertainty in the true distribution of atmospheric latent heating, it is advisable to be aware of the sensitivity of the circulation to the loca-

tion of heating. One general conclusion from this work is that the atmosphere might be more sensitive to biases in the latitude of the heating than in its height.

3.5 Conclusions

In this chapter, the sensitivity of the eddy-driven jet in a zonally symmetric North Atlantic-like background state has been examined in an idealized dry GCM using a series of thermally forced simulations. The model was set up to allow the sensitivity in both summer and winter to be studied by keeping each hemisphere in a perpetual seasonal state.

The main findings in these experiments are the following:

1. The speed and latitude of the jet display different sensitivities to thermal forcing. This means that treating the response of the jet under forcing as simply a change in NAO or annular mode index will not capture the full response. It also means that the speed and latitude of the jet are affected by different mechanisms in response to thermal forcing.
2. The sensitivity of the jet varies between winter and summer. Latitude shift sensitivities are highest in the midlatitudes and polar regions near to the jet core, and these sensitivities move poleward in summer. Jet speed sensitivities extend more prominently into the subtropics and tropics. The winter jet latitude is, however, affected by heating in the summer hemisphere.
3. The sensitivities of the storm track closely match those of the jet, suggesting that the two shift and strengthen in near tandem in response to forcing. Understanding how the jet will respond is therefore vital in understanding how midlatitude weather will be affected.
4. The variability of the jet, as quantified in this work by the jet spread measure, is shown to broadly increase when heating in the jet core and on the equatorward flank of the jet, but decrease when forced in the polar region. Changes in spread

are consistent with changes in the baroclinic zone width and low level meridional temperature gradient, with an increase in spread due to an increase in baroclinic zone width and/or meridional temperature gradient, and a decrease due to a decrease in baroclinic zone width and/or meridional temperature gradient.

5. In the case studies examined, responses of latitude and speed scale approximately linearly with the strength and combination of forcings applied. This is especially clear for the winter cases.

The proposed mechanisms for the sensitivities determined in this chapter come from changes in the slopes of isentropic surfaces and hence in the location and strength of the strongest horizontal temperature low level gradients (as in Butler et al., 2011). Shifts of the jet are driven by changes in the latitudinal position of the maximum low level meridional temperature gradient which determines where the maximum meridional heat flux occurs. Changes in jet strength arise from changes in the low level meridional temperature gradient across the storm track, which dictates the amount of meridional heat flux.

Understanding the jet sensitivities is the first step towards understanding the effects of climate change on the midlatitudes. Given uncertainty in the distribution of thermal forcing, knowledge of the sensitivity of the dynamics to forcing location provides guidance on how uncertainty in the circulation response could be reduced. Areas where the uncertainty of the anthropogenic forcing is high and the sensitivity of the jet is high (e.g. in the tropical upper troposphere) are clearly areas that need further study to greater constrain the position and size of the expected anthropogenic forcing.

Chapter 4

Hadley cell sensitivity in an idealized GCM

In the previous chapter, we studied the sensitivity of the eddy-driven jet to localized diabatic heatings in the atmosphere. We found that the jet speed and latitude have different sensitivity distributions from each other, which also vary between summer and winter. We also presented a simple mechanistic understanding of the sensitivities by considering how the individual thermal forcings modify mean isentropic surfaces. In the cases analyzed, the jet response to forcing scaled approximately linearly with the strength of the forcing and when forcings are applied in combination. We now turn our attention to the Hadley circulation, investigating its sensitivity in the same framework. We also look at the connection between the Hadley circulation and the jet, and how the Hadley circulation plays an important role in mediating the response of the jet to heating in the opposite hemisphere.

4.1 Introduction

The tropical belt expansion is one of the robust observed (Seidel et al., 2008) and simulated (Lu et al., 2007) effects of anthropogenic climate change. This expansion could

have societal impacts due to changes in precipitation (Seidel et al., 2008). Impacts are not limited to the tropical regions but extend into the extratropics too, due to the Hadley cell and eddy-driven jet interdependence. The tropical belt expansion has been connected to extratropical circulation changes (Previdi and Liepert, 2007), with idealized studies suggesting an interactive relationship between tropics and extratropics (Becker et al., 1997; Schneider, 2004; Walker and Schneider, 2006; Mbengue and Schneider, 2013). It is therefore important to understand the broad range of circulation changes and their seasonal dependencies that may occur due to the range of forcings associated with a changing climate.

Seasonal dependencies exist in the Hadley cell (HC) and eddy-driven jet relationship (Kang and Polvani, 2011; Ceppi and Hartmann, 2013). By considering the momentum budget in the HC's upper branch, Schneider and Bordoni (2008) show the winter cell is less affected by extratropical eddies and is more thermally driven compared to the summer cell. In the summer, the HC is further poleward than winter, consistent with a smaller local Rossby number (Kang and Lu, 2012), which means eddy-driving is the dominant driver of HC variability (Caballero, 2007). Despite these seasonal dependencies, many idealized modeling studies investigating HC responses to El Niño–Southern Oscillation and global warming-like forcings are conducted under hemispherically symmetric conditions (e.g., Frierson et al., 2007; Levine and Schneider, 2011; Sun et al., 2013; Tandon et al., 2013). Idealized modeling studies in nonequinoctial conditions allow us to investigate how the seasonal response of the HC to climate forcings behaves. It has even been suggested that the solstitial cell, the single cross-equatorial cell with a rising branch in the summer hemisphere, is the dominant component of the HC (Lindzen and Hou, 1988), and the equinoctial pattern is ephemeral (Hu et al., 2007), implying results from studies run under equinoctial conditions may not hold throughout the annual cycle. It is therefore necessary to test HC and eddy-driven jet responses and their relationship to one another to verify that results from equinoctial studies apply in different seasons.

Under equinoctial conditions, Frierson et al. (2007) find that the HC expands under increases in global mean temperature. For tropically confined heating, the HCs contract, and it is only when the heating has a wider meridional extent that the HCs expand (Tandon et al., 2013). Heating with a wider meridional extent when eddy feedbacks are not present causes the HCs to contract instead of expand (Sun et al., 2013), suggesting the eddy-driven jets drive HC expansion when the heating is outside of the tropics. Using experiments with locally applied diabatic heating it is possible to test these results hold in nonequinoctial conditions. Several mechanisms for understanding HC expansion have been proposed. Lu et al. (2008) find HC expansion is driven by changes in meridional temperature gradients modifying eddy phase speeds. However, Schneider and Walker (2006) postulate HC extent is governed by the latitude at which baroclinic eddies become deep enough to reach the upper troposphere and so the eddy flux divergence of angular momentum changes sign. Ceppi et al. (2013) find a cross-hemispheric relationship between the HC and jets under equinoctial conditions, and we aim to explore the seasonality and mechanisms of the relationship using a model with a solstitial setup.

To understand the broad range of possible HC changes due to the changing climate, we conduct a systematic set of experiments applying heating in all possible locations to build up HC sensitivity plots and the seasonality of these sensitivities. Changes in the HC's poleward extent may result from the tropical belt expansion but also from intertropical convergence zone (ITCZ) shifts for fixed HC width. We examine the ITCZ position and HC width, extent and strength sensitivities to localized diabatic heating in a dry idealized model with perpetual winter and summer hemispheres. Theories in dry atmospheres are important because they provide the limits to which moist theories must converge.

The hypothesized storm track response to HC extent shifts (Mbengue and Schneider, 2017) leads us to investigate how the eddy-driven jets respond to forced HC changes. We find cross-hemispheric responses to diabatic heating; when heating is applied in one hemisphere, there is a response in the opposite hemisphere's HC branch and eddy-driven

jet. We argue that the winter cell’s dominant behavior is important for understanding the interhemispheric response mechanisms.

4.2 Methods

4.2.1 Model

Simulations are conducted using the Geophysical Fluid Dynamics Laboratory’s Flexible Modeling System dry dynamical core. The model solves the primitive equations on a sphere using a spectral transform method at T42 resolution with 37 unevenly spaced sigma levels ($\sigma = p/p_{\text{surface}}$, the model’s vertical coordinate). The model includes a quasi-equilibrium convection scheme, relaxing temperatures in an atmospheric column to a prescribed lapse rate in an energetically consistent way. A full description of the model can be found in Schneider and Walker (2006). The configuration used here is as in Chapter 3: perpetual winter (southern hemisphere of the model) and summer (northern hemisphere of the model) hemispheres with no diurnal or seasonal cycle, in which temperatures are relaxed toward a solstitial profile using Newtonian relaxation.

The model streamfunction and jets are qualitatively similar to Earth’s Southern Hemisphere (Fig. 4.1). Comparing the model’s southern hemisphere with the JJA Southern Hemisphere and the model’s northern hemisphere with the DJF Southern Hemisphere illustrates this.

We use the same simulations as used in Chapter 3. Following a 10-year spin-up period to allow the model to reach a statistically steady state, a 6-year control simulation and 306 6-year tropospheric heating experiments were run, all starting from the end of the spin-up simulation. Each simulation has a constant zonally uniform heating applied at a specific location in the latitude-sigma plane. The heating has a two dimensional Gaussian structure in this plane with a 2 K day^{-1} maximum, 7.1° latitudinal standard deviation, and 0.053 sigma units vertical standard deviation. The 306 simulations cover

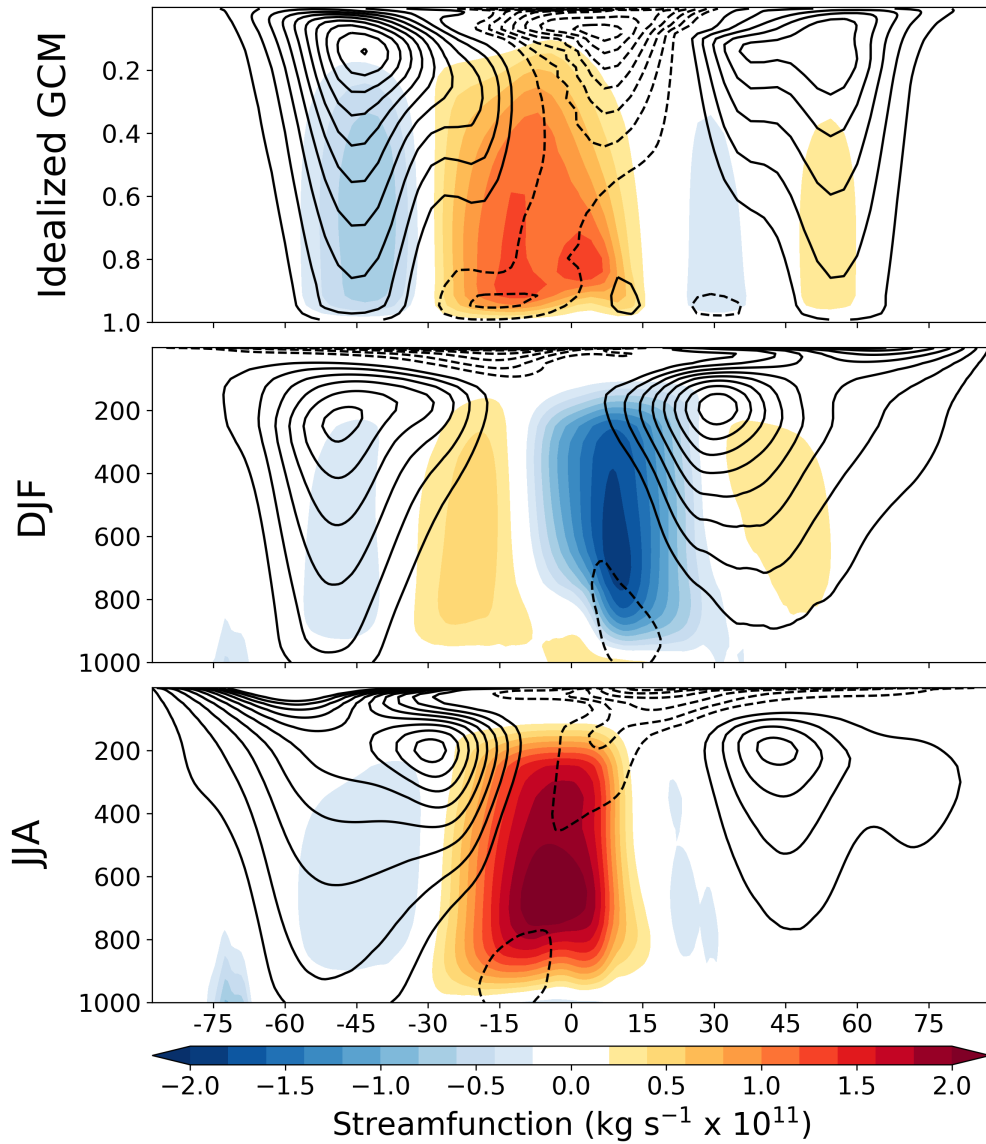


Figure 4.1: Comparison of model and reanalysis. Top panel shows the model control simulation zonal mean streamfunction (colors) and zonal mean zonal wind (contours at 5 ms^{-1} intervals). Middle and bottom panels show the zonal mean streamfunction and zonal mean zonal wind for DJF and JJA respectively from ERA-Interim (Dee et al., 2011) averaged over 1980-2014.

heating across a grid of 34 different latitudes and 9 sigma levels. The amplitude of the heating is on the same order of magnitude as the long wave cooling due to aerosols over parts of Asia (Kuhlmann and Quaas, 2010) and the latent heating produced by 8-9 mm of precipitation per day (Jin and Hoskins, 1995).

4.2.2 Indices

We define the ITCZ location as the latitude at which the zonally and vertically (σ between 0.4 and 0.8) averaged streamfunction changes sign between the winter and summer HC branches. Likewise the winter and summer HC extents are defined as the latitude at which the zonally and vertically averaged streamfunction changes sign between the respective HC and Ferrel cell branches. The strength of each HC branch is defined as the streamfunction maximum in the branch. The eddy-driven jet is defined as the latitude of the maximum zonal-mean zonal wind at $\sigma = 0.85$.

The sensitivities of each index are calculated as the difference between the time average of the monthly indices in the forced simulations and the control simulation. As in Chapter 3, we test for significance using a Student's t -test at the 1% significance level.

4.3 Results

4.3.1 Sensitivities

ITCZ shifts in response to diabatic heating are shown as a sensitivity plot in Fig. 4.2a as in Chapter 3. Each dot summarizes the response in a forced simulation with heating at that location in the latitude-sigma plane. Colors show the ITCZ response due to forcing at each location (warm colors indicate a northward ITCZ shift). Black dots mark statistically significant runs. The contours show the control streamfunction. Statistically significant ITCZ shifts arise from heating throughout the atmosphere in both hemispheres. For reference, the monthly ITCZ standard deviation in the control run is 0.51° , so many responses are large compared to the variability. ITCZ position is determined by the energy flux across the equator and the net energy input to the equatorial atmosphere (Bischoff and Schneider, 2014). Therefore, heating on either side of the ITCZ causes a shift in latitude toward the heating.

Fig. 4.2c shows the tropical belt width sensitivity (summer HC to winter HC ex-

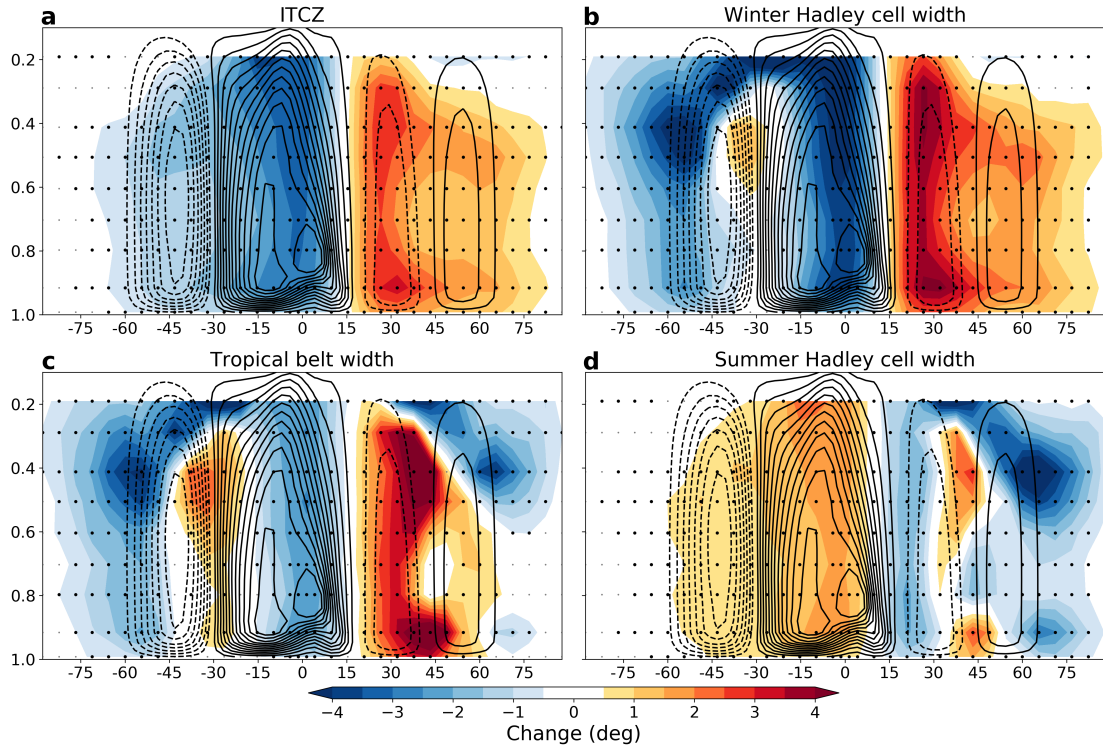


Figure 4.2: Intertropical convergence zone (ITCZ), tropical belt and Hadley cell width sensitivities to heating experiments (colors) in the latitude-sigma plane. Contours show the zonal-mean control streamfunction. Black dots mark where the difference between the perturbed and control simulations are statistically significant. For the ITCZ, positive values correspond to a shift toward 90°N.

tent). Heating confined to the winter cell in the deep tropics acts to contract the tropical belt, while subtropical upper-tropospheric heating can cause a tropical belt expansion. Heating on the poleward flank of the Ferrel cell in both hemispheres acts to contract the tropical belt. Patterns are similar to the eddy-driven jet sensitivities in Fig. 3.3, suggesting when forcing at high latitudes, it is the dynamics of extratropical eddies that affect the tropical belt width. The HC widths (Fig. 4.2b,d) display a similar behavior, with high latitude forcing contracting the HCs in the same hemisphere. In contrast, the response to deep tropical forcing appears to be dominated by ITCZ shifts.

We next consider the winter and summer HC poleward extent sensitivity (Fig. 4.3a,c). For reference, the monthly poleward HC extent standard deviation in the control run is 0.63° in winter, and 0.84° in summer. When heating is applied in the extratropical and polar regions, the HC and jet latitudinal sensitivities in that same hemisphere display

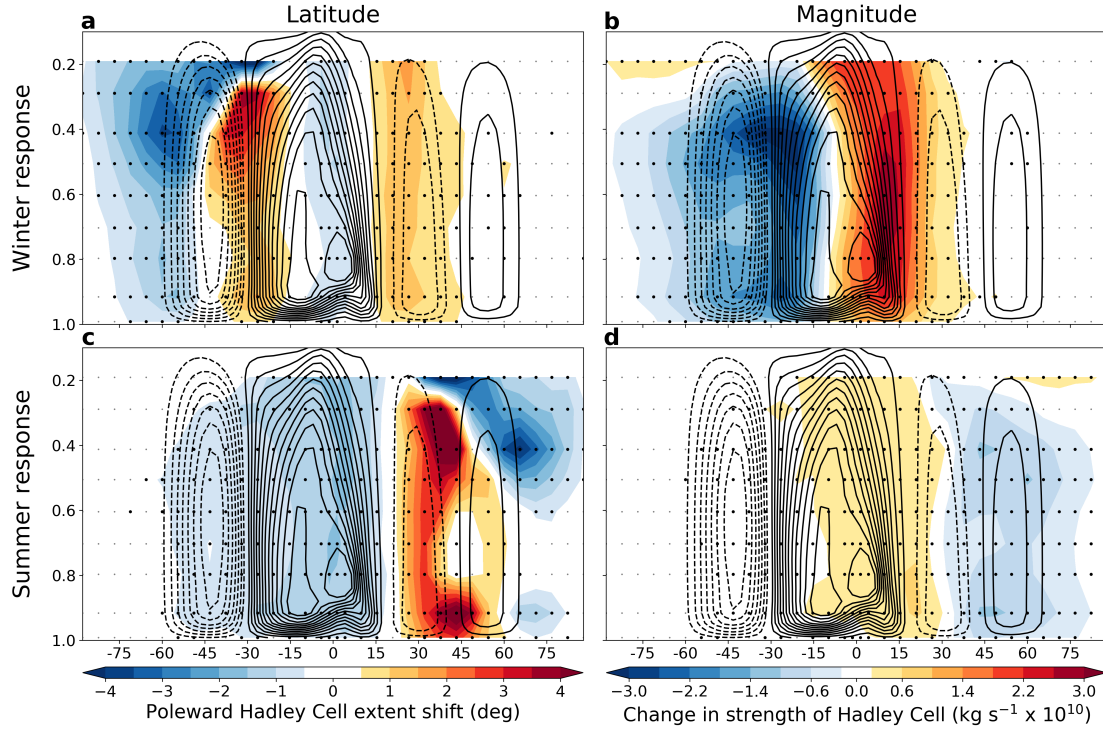


Figure 4.3: Hadley cell sensitivities to heating experiments (colors) in the latitude-sigma plane. Contours show the zonal-mean control streamfunction. Black dots mark where the difference between the perturbed and control simulations is statistically significant. Positive values correspond to poleward Hadley cell extent shift or Hadley cell strengthening in the target hemisphere (a,b: winter; c,d: summer)

strong similarities (see Fig. 3.3), but with smaller HC shifts compared to the jet. This implies that the HC extent and eddy-driven jet are shifting in tandem, with a greater eddy-driven jet shift. This is analogous to the result of Mbengue and Schneider (2013), who find the HC extent and storm track often shift in tandem with a greater storm track shift. The shifts due to heating in the midlatitudes increase with height due to static stability arguments: the higher the heating altitude, the greater the increase in static stability and thus, by simple scaling arguments (Lu et al., 2007), the greater the shift in HC extent. There is also an enhanced shift region found near the surface in summer in the extratropics (cf. Fig. 3.3).

Corroborating Lu et al. (2008), tropical heating acts to contract the HCs, the “El Niño-like” response, whereas extratropical heating produces poleward HC expansion, the global warming-like response. This effect is also seen for the HC widths (Fig.

4.2b,d), but is not seen for the summer HC extent (Fig. 4.3c), as this appears to be dominated by the ITCZ shifts.

Statistically significant cross-hemispheric responses exist, with a summer hemisphere heating (poleward of the ITCZ) causing the winter HC to expand poleward. In contrast, winter hemisphere heating causes the summer HC to contract equatorward.

HC strength changes are governed by the heating's location with respect to the cell's ascending and descending branches (Fig. 4.3b,d). Heating in the cell's ascending branch (i.e., across the ITCZ) acts to strengthen the overturning circulation in both HCs. Heating in the descending branches and the Ferrel cells acts to weaken the overturning circulation in that hemisphere. In both cases, the winter HC strength response is greater than the summer cell. For reference, the monthly HC strength standard deviation in the control run is $4.0 \times 10^9 \text{ kg s}^{-1}$ in winter, and $2.4 \times 10^9 \text{ kg s}^{-1}$ in summer.

Winter HC width and strength changes are greater than summer HC changes. This may be because the winter HC is in the regime where the angular momentum conserving model of the HC applies (a weaker cell can be wider before instability occurs) so local diabatic heating is balanced by HC changes. Weaker changes in the summer HC suggest that in summer, local diabatic heating is balanced more strongly by eddy changes; the cell is in an eddy-driven regime, with a lower local Rossby number. This HC regime behavior agrees with Schneider and Bordoni (2008), who find eddy-mediated HC regime transitions during the annual cycle.

To quantify the changes in regime behavior, we compute the local Rossby numbers in each HC, following Schneider and Bordoni (2008). This is done by taking the ratio of the local vorticity to the planetary vorticity in the upper troposphere above the HC streamfunction maximum. High Rossby numbers indicate that the cell responds more to thermal driving, whereas lower Rossby numbers suggest eddy-driving is more dominant. The ratio of the local Rossby numbers in the winter to summer cells is 1.5, supporting the hypothesis that the changes in the winter cell are more thermally driven, compared to more eddy-driven changes in the summer cell. The winter to summer ratio of maxi-

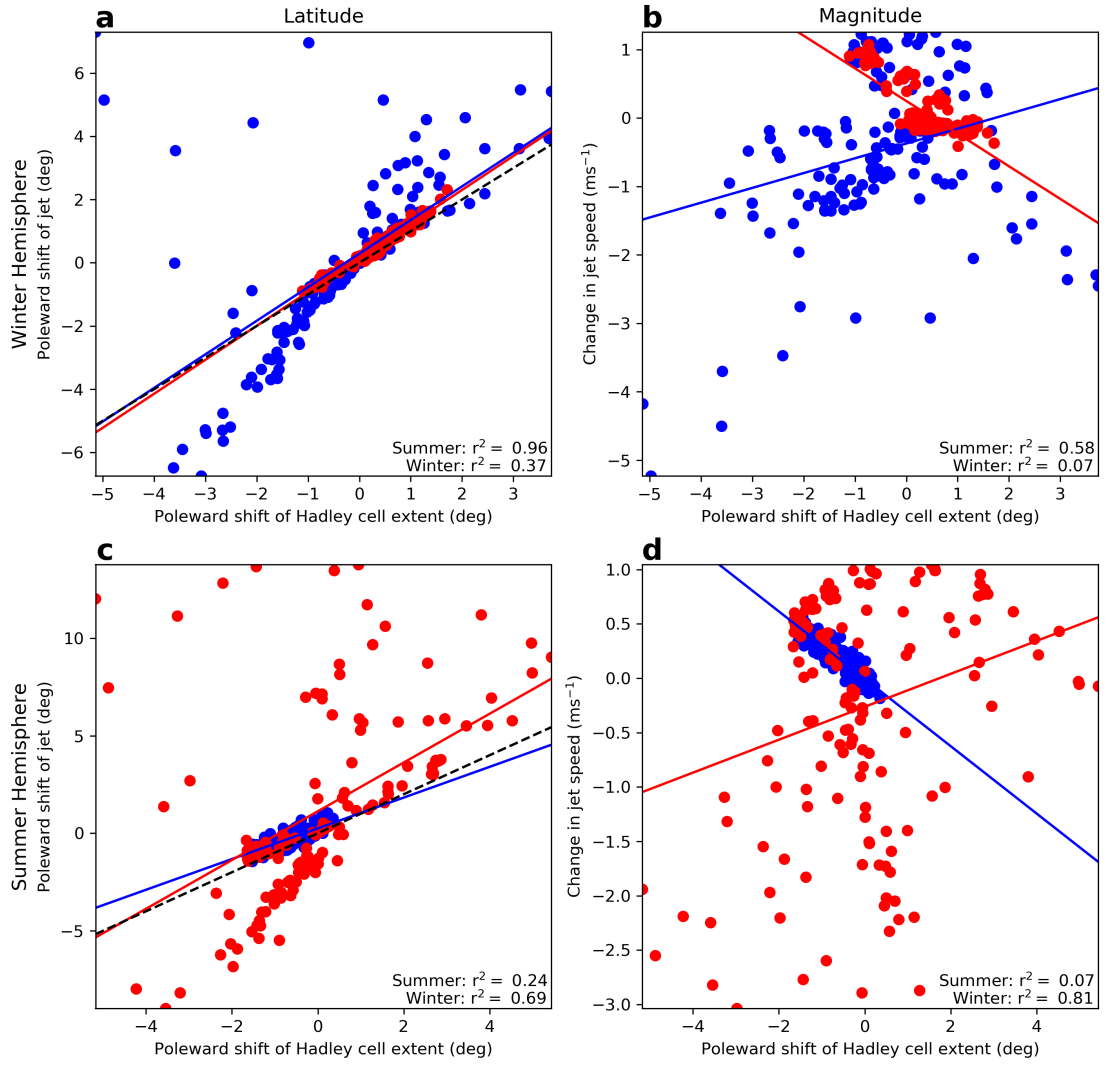


Figure 4.4: Scatter plots of changes in jet indices against changes in Hadley cell extent between the forced and control simulations. (a,b) Winter jet and Hadley cell response. (c,d) Summer jet and Hadley cell response. Blue dots show simulations where the heating is applied in the winter hemisphere, red dots where the heating is applied in the summer hemisphere. The 1:1 Hadley cell to jet shift is shown in the dashed black lines on the left-hand panels.

num HC shift divided by maximum $\overline{v'T'}$ shift in each hemisphere is 1.7, supporting the idea local diabatic heating is balanced more strongly by HC changes in winter and eddy changes in summer.

4.3.2 Cross-hemispheric connections

Both HC extents display a sensitivity to heating in the opposite hemisphere. The heating in one hemisphere also projects onto the eddy-driven jet in the other hemisphere, suggesting a cross-hemispheric jet response mediated by the HC (as in Chapter 3 and Ceppi et al., 2013). To see this, we plot the eddy-driven jet speed and latitude response against the HC extent shift in each hemisphere (Fig. 4.4). Fig. 4.4a shows heating in the summer hemisphere (red dots) shifts the winter HC's poleward extent, which is highly correlated with a poleward shift of the winter eddy-driven jet. The shifts are close to 1:1 (i.e., 1° of HC extent shift corresponds to 1° of jet shift). Fig. 4.4b shows a correlation between winter HC extent and winter eddy-driven jet speed when heating in the summer hemisphere. This correlation is less robust than the correlation between the HC extent and eddy-driven jet shift. Note here a poleward jet shift corresponds to a jet weakening. More commonly, a poleward jet shift would be associated with the jet strengthening due to increases in eastward eddy phase speeds when the jet strengthens (e.g. Chen and Held, 2007), suggesting a different mechanism in the cross-hemispheric response cases.

The summer HC extent and summer eddy-driven jet responses (Fig. 4.4c,d) are similar to the winter case, but with a stronger relation to the jet speed than latitude this time. Heating in the winter hemisphere (blue dots) causes the summer HC extent and summer jet strength to change together. The summer jet shift is correlated with the summer HC extent shifts to a lesser degree (also a near 1:1 relationship). As the jet shifts equatorward it strengthens, analogous to the poleward shifted jet weakening in the winter jet case. Note the summer HC extent and ITCZ position are correlated with an R-squared value of 0.97 (not shown), indicating the dependencies found between the summer HC extent and summer jet when heating in the winter hemisphere also apply to the ITCZ position and jet.

To investigate the mechanisms behind the HC and jet responses, we analyze two forcing simulation case studies. The first is the simulation where heating in the summer

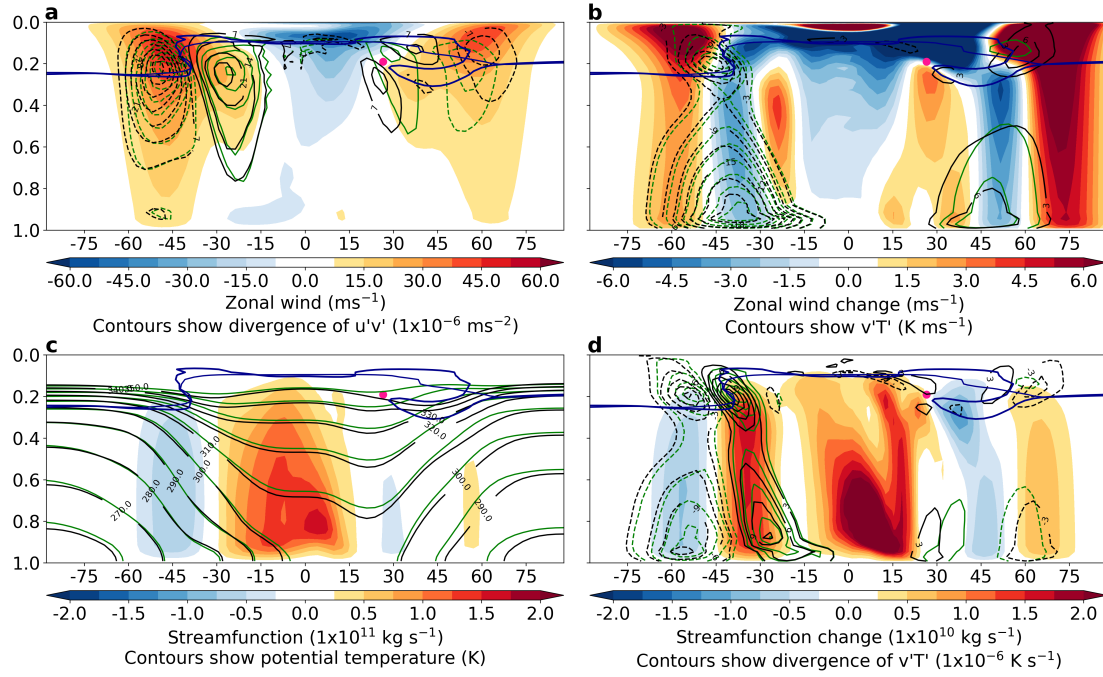


Figure 4.5: Summary plot for heating applied at $\sigma = 0.2$ and $\phi = 27^\circ$ (pink dot). (a) Simulation $\bar{u}$ (colors), $\nabla \cdot \overline{u'v'}$ (black contours for simulation, green for control). (b) Change in $\bar{u}$ (colors), $\overline{v'T'}$ (black contours for simulation, green for control). (c) Simulation streamfunction (colors), potential temperature (black contours for simulation and green for control). (d) Change in streamfunction (colors), $\nabla \cdot \overline{v'T'}$ (black contours for simulation and green for control). The thick blue line is the tropopause in the simulation, and the thin blue line is the control tropopause.

hemisphere produces the greatest poleward shift of the winter HC extent and eddy-driven jet (Fig. 4.5). Heating acts to strengthen the HC winter branch and shift both the winter Hadley and Ferrel cells poleward. There are also poleward shifts of the winter heat and momentum fluxes on the same order of magnitude as the HC, and hence eddy-driven jet. The heating is conveyed to the winter hemisphere via the HC and warms the upper troposphere near the maximum upper level winter heat flux (shown by the lowering of the isentropes near the tropopause). Heating here acts to shift the jet poleward (Fig. 3.3) by shifting the $\overline{v'T'}$ maximum poleward (by the mechanism discussed in detail in Chapter 3).

The second case study is the simulation where the heating in the winter hemisphere causes the greatest speed increase in the summer eddy-driven jet (Fig. 4.6). The eddy-driven jet strengthening is barotropic (Fig. 4.6b), implying it is caused by changes in the

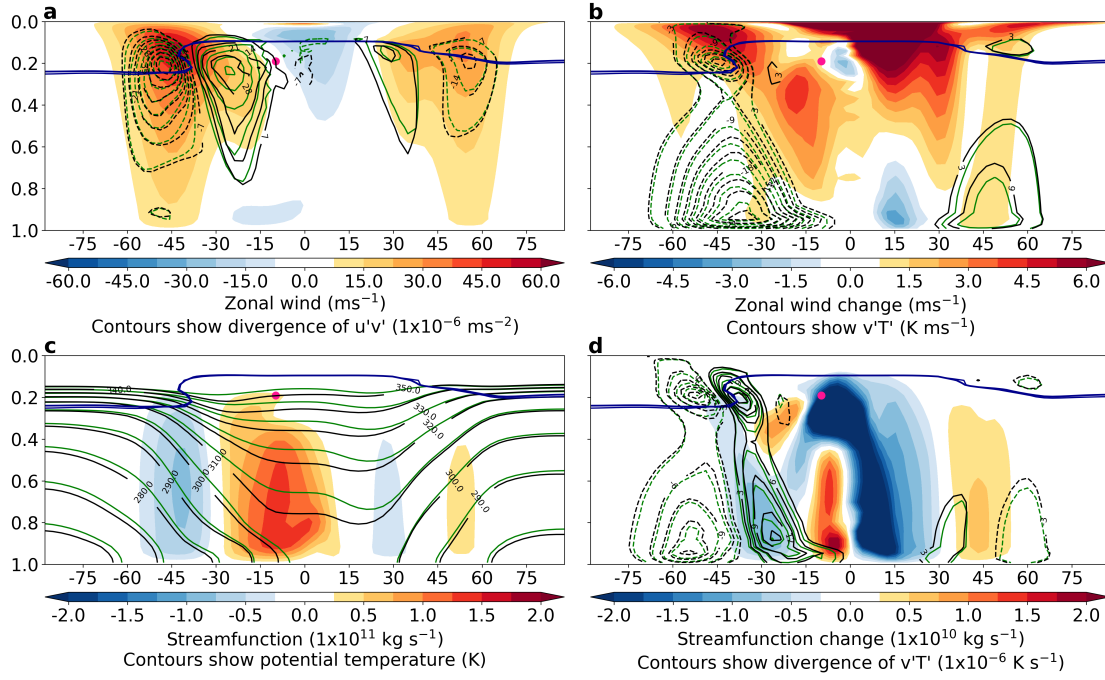


Figure 4.6: Summary plot for heating applied at $\sigma = 0.2$ and $\phi = -10^\circ$ (pink dot). (a) Simulation $\bar{u}$ (colors), $\nabla \cdot \bar{u}'v'$ (black contours for simulation, green for control). (b) Change in $\bar{u}$ (colors), $\bar{v}'T'$ (black contours for simulation, green for control). (c) Simulation streamfunction (colors), potential temperature (black contours for simulation and green for control). (d) Change in streamfunction (colors), $\nabla \cdot \bar{v}'T'$ (black contours for simulation and green for control). The thick blue line is the tropopause in the simulation, and the thin blue line is the control tropopause.

eddies. The heating strengthens summer subtropical temperature gradients (Fig. 4.6c), strengthening the subtropical jet (Fig. 4.6b). Eddies can then penetrate further into the summer tropics, demonstrated by the divergence increase on the equatorward flank of the summer eddy momentum divergence (Fig. 4.6a). The extra eddy momentum diverged from the tropics results in increased eddy momentum convergence into the jet (Fig. 4.6a), barotropically strengthening it.

Although the applied heating is near the top of the troposphere in both case studies, the response patterns in simulations where the heating is applied nearer the surface display the same features, but with smaller HC extent and jet shifts and smaller changes in jet speed (see Fig. 4.3). This suggests that the mechanisms affecting the HC and jets are amplified at higher levels, as hypothesized by Yuval and Kaspi (2016).

Given the high correlation of the ITCZ and summer HC extent, it would appear that the correlations in the summer hemisphere in response to forcing in the winter hemisphere are dominated by how the ITCZ and hence how the winter HC is responding to the heating. This suggests the winter HC is most important for understanding the cross-hemispheric responses to forcing.

4.4 Discussion and conclusions

We have shown the winter and summer HCs in a dry idealized GCM display different sensitivities to localized diabatic forcing. These seasonal sensitivities apply to HC width, extent, and strength, and cause the ITCZ shift and tropical belt width to vary depending on which hemisphere the heating is applied. For example, the HC strength is much more sensitive in winter, whilst the extent has broadly similar levels of sensitivity in both seasons. The sensitivities are broadly barotropic in nature, with some increase in magnitude as the heating is applied higher in the troposphere.

We have also shown strong correlations between the HC and eddy-driven jet when forced from the opposing hemisphere. Winter hemisphere heating forces the summer HC extent, which is highly correlated with summer jet speed changes; summer hemisphere heating forces the winter HC extent which is highly correlated with winter jet shifts. Summer HC responses are highly correlated with the ITCZ response, implying HC and jet correlations are dictated by ITCZ shifts via the dominant winter cell response.

The ITCZ shift sensitivities agree with idealized (Kang et al., 2008) and more complex modeling studies (Chiang and Bitz, 2005; Broccoli et al., 2006), with a shift in position toward the location of heating/away from cooling. The HC width, extent, and tropical belt width sensitivities are broadly consistent with a range of studies looking at HC changes in simple and complex models (Frierson et al., 2007; Previdi and Liepert, 2007; Lu et al., 2007, 2008; Sun et al., 2013; Tandon et al., 2013), showing at least the signs of responses from equinoctial studies match the solstitial case. Tandon et al.

(2013) study the “El Niño-like” heating in the equinox case, finding a contraction of the HC due to El Niño. However, El Niño peaks in NH winter, and our results corroborate the response they find but with a more relevant background state. For the eddy-driven jets and HCs, heating produces similar sensitivity patterns when the forcing is in the same hemisphere as the jets, supporting the idea that the midlatitude eddies feed back onto the HC (Becker et al., 1997; Walker and Schneider, 2006; Caballero, 2007; Kang et al., 2008; Kang and Polvani, 2011; Ceppi and Hartmann, 2013).

The correlations between HC and jet from heating in the opposing hemisphere appear to be governed by the dominant winter HC behavior. The responses see the eddy-driven jets weaken and shift poleward in tandem or strengthen and shift equatorward. This suggests that the mechanism by which the HCs force the jets is not via changes in eddy phase speeds (Chen and Held, 2007), which would see a strengthening and poleward jet shift occur in tandem. We postulate a simple mechanistic understanding of how the jets may respond to the HC changes through the modifications of the overturning circulation projecting onto the heat fluxes and subtropical jets; a full diagnosis of the mechanism is presented in Chapter 5.

Our findings suggest that it is important to study the nature of the solstitial HCs and conduct studies in an annual cycle framework, investigating the different HC seasonal responses. There are also clear implications given the uncertainty in the size and spatial extent of the anthropogenic forcing due to climate change. Areas where the uncertainty of the anthropogenic forcing is high and the HC sensitivity is high (e.g., in the tropical upper troposphere) are clearly areas needing further study to better constrain the nature of the anthropogenic forcing.

Chapter 5

Transient response to forcing

In the previous two chapters, we presented the sensitivities of the Hadley cell and eddy-driven jet to local diabatic forcing. We also looked at cross-hemispheric responses of the Hadley cell and eddy-driven jet to forcing applied in the opposite hemisphere. The eddy-driven jet sensitivities rely on eddy-mean flow interactions being set up. Arguments based on equilibrium statistics can be circular, so to gain further insight into the mechanisms that cause the responses of the Hadley cell and eddy-driven jet, a series of spin-up simulations are run and presented in this chapter. These aim to elucidate how the response is set up when the forcing is switched on, a technique used successfully by others in the field (e.g. Deser et al., 2007; Strong and Magnusdottir, 2010; Chen et al., 2013; Sun et al., 2013; Lu et al., 2014).

5.1 Methods

We make use of the GFDL Flexible Modeling System, configured as described in Chapter 3. Following the 10-year spin-up simulation used previously, we run the model for a further 2000 days, taking the state at each 10 day interval to give 200 initial condition states. For a forcing experiment with heating applied at a particular location in the latitude-sigma plane, we can then run a 200 member spin-up ensemble from the 200 un-

forced initial condition states. The forcing is applied at day 0, and then using the initial condition state, the model is run for 180 days. The model outputs u , v and T at 6-hourly timesteps.

We take the 2000 day run used to generate the initial condition states as a control run. The atmospheric response to forcing for u , v and T is defined as the difference between the mean of the 200 member ensemble at a particular timestep and the mean of the control ensemble over the entire 2000 day period. Transient eddy quantities are computed for each timestep using the following formula:

$$u'v' = \frac{1}{200} \sum_{i=1}^{200} [(u_i - \bar{u}_i)(v_i - \bar{v}_i) \cos \phi] - \langle u'v'_{\text{control}} \rangle, \quad (5.1)$$

where the subscript i corresponds to the i th ensemble member, the overbars indicate the zonal mean at that timestep, ϕ is the gridpoint latitude and $\langle u'v'_{\text{control}} \rangle$ is the control mean eddy quantity as output by the model (which includes the cosine weighting), and then meaned over the full control run period. Using these variables, and other derived quantities such as potential temperature and the streamfunction, we can investigate how the response to a locally imposed forcing evolves over time and how the steady state response is set up.

5.2 Results

Fig. 5.1 shows the transient response to a heating applied at $\sigma = 0.9$ and $\phi = 49^\circ$. This heating location is in a very similar location to the first case study discussed in Chapter 3, and results in a poleward shifted but weakened summer eddy-driven jet. The variables plotted in the Hovmöller plots have been filtered using a 20-day low-pass Lanczos filter with 51 weights (Duchon, 1979) to remove high frequency variability, and then normalized by the maximum magnitude of their transient response. Note that even with a 200 member ensemble, the noise due to internal variability remains noticeable in the

ensemble mean because of the variability of the midlatitude winds. Hence we focus on the forced ensemble mean features. The initial response to the heating (0-10 days) manifests as an increase in $v'T'$ on the poleward flank of the climatological $v'T'$ maximum at low levels. These changes in heat fluxes occur as the diabatic heating directly modifies the local potential temperature isentropes (as shown in Chapter 3). The next stage of the response is a decrease in the eddy momentum convergence, initially on the equatorward side of the maximum convergence in the midlatitudes (10-20 days). Following this, the $v'T'$ response becomes a poleward shift and weakening of the climatological $v'T'$ maximum, and a weakening in the core of the eddy momentum flux convergence, with strengthening of the convergence at higher latitudes (set up around day 30). The zonal wind then shifts, initially at higher levels, but subsequently at lower levels (full response set up by day 60). The greatest change in the zonal wind is an increase on the poleward side of the eddy-driven jet, but as the weakening occurs closer to the jet core, the overall response is a weakening of the jet. As a consequence of the eddy-driven jet shifting polewards, there is anomalous eddy momentum flux convergence in the subtropics (from day 20). This causes the Hadley cell to expand and weaken (as $f\bar{v} \sim \partial_y \overline{u'v'}$ in the subtropics), weakening the subtropical jet (from day 60).

The Hovmöller plots point to the importance of the eddy adjustment for changes in the eddy-driven jet. To gain further insight into the mechanisms behind the jet changes, we look at the quasi-geostrophic Eliassen-Palm (EP) flux diagnostics. Starting with the conventional transformed Eulerian-mean momentum equation (Edmon et al., 1980; Andrews et al., 1987),

$$\frac{\partial \bar{u}}{\partial t} = f\bar{v}^* + \frac{1}{a \cos \phi} \nabla \cdot \mathbf{F} + \bar{X}, \quad (5.2)$$

where f is the Coriolis parameter, a is the radius of the Earth, $\bar{v}^*$ is the residual meridional velocity, $\bar{X}$ is the friction near the surface, and $\nabla \cdot \mathbf{F}$ is the EP divergence, which

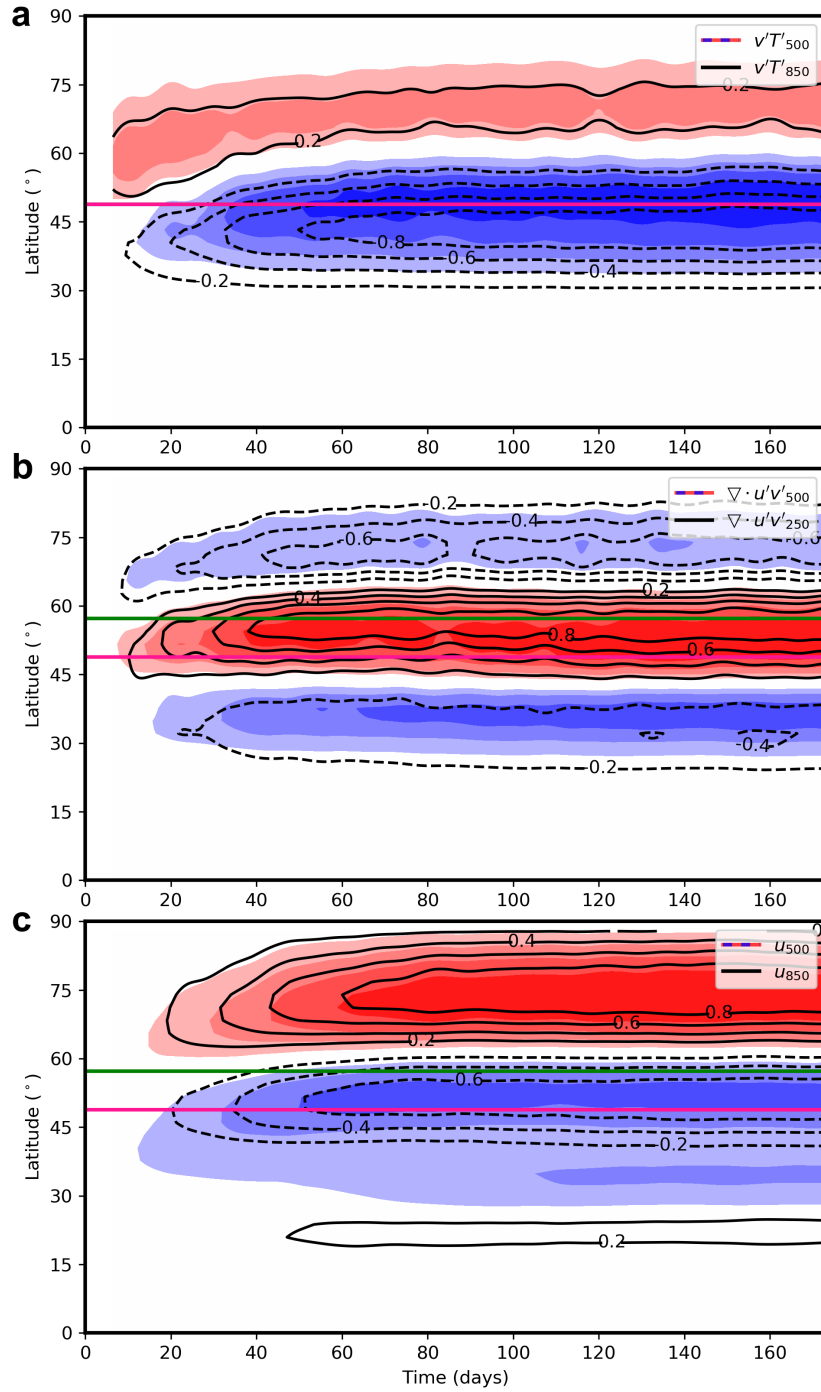


Figure 5.1: Hovmöller plots for heating applied at $\sigma = 0.9$ and $\phi = 49^\circ$. Normalised transient responses in (a) $v'T'$, (b) $\nabla \cdot u'v'$, (c) u . Contour intervals for the colours and contours are 0.2. The pink line indicates the heating latitude. The green line indicates the latitude of the corresponding index's maximum in the control run [e.g. in (c) the u maximum at $\sigma = 0.85$ in the control]. Note in (a), the latitude of the maximum $v'T'$ coincides with the latitude of the heating.

has the form

$$\frac{1}{a \cos \phi} \nabla \cdot \mathbf{F} = -\frac{1}{a \cos^2 \phi} \frac{\partial (\overline{u'v'} \cos^2 \phi)}{\partial \phi} + f \left(\frac{\overline{v'\theta'}}{\partial \Theta / \partial p} \right)_p. \quad (5.3)$$

Here, Θ is the global mean potential temperature θ at each pressure level. Assuming that the eddies are geostrophic and ignoring the divergence from the gradient in f , it can be shown that

$$\overline{v'q'} = \frac{1}{a \cos \phi} \nabla \cdot \mathbf{F}, \quad (5.4)$$

where $\overline{v'q'}$ is the zonal mean eddy potential vorticity flux, and q is the quasigeostrophic potential vorticity,

$$q = f + \zeta + f \frac{\partial}{\partial p} \left(\frac{\theta - \Theta}{\partial \Theta / \partial p} \right). \quad (5.5)$$

Fig. 5.2a shows the summer hemisphere climatological EP flux vectors (arrows) and divergence (shading), along with the momentum convergence (contours) (Fig. 5.7a shows both hemispheres). The vertical component of the EP flux arises from the meridional heat flux and acts to reduce the intensity of upper level midlatitude westerlies, transferring angular momentum from the upper troposphere to the surface where it is dissipated by surface friction. The horizontal component of the EP flux acts to extract angular momentum from the subtropics, decelerating the subtropical jet, and deposit it in the midlatitudes, accelerating the eddy-driven jet.

When the heating is switched on, the initial response sees the eddy source region strengthening on the poleward flank of the $v'T'$ maximum. This manifests as anomalous near-surface convergence of the EP flux (Fig. 5.2b), associated with the aforementioned increase in near-surface $v'T'$, leading to an increase in upwards wave flux activity. These changes are driven by enhanced baroclinic instability due to the decrease in static stability and low-level meridional temperature gradient (as shown in Chapter 3). The next stage of the response (Fig. 5.2c) sees the eddy source region weaken on the equatorward side with anomalous convergence and downward EP flux on the equatorward flank of

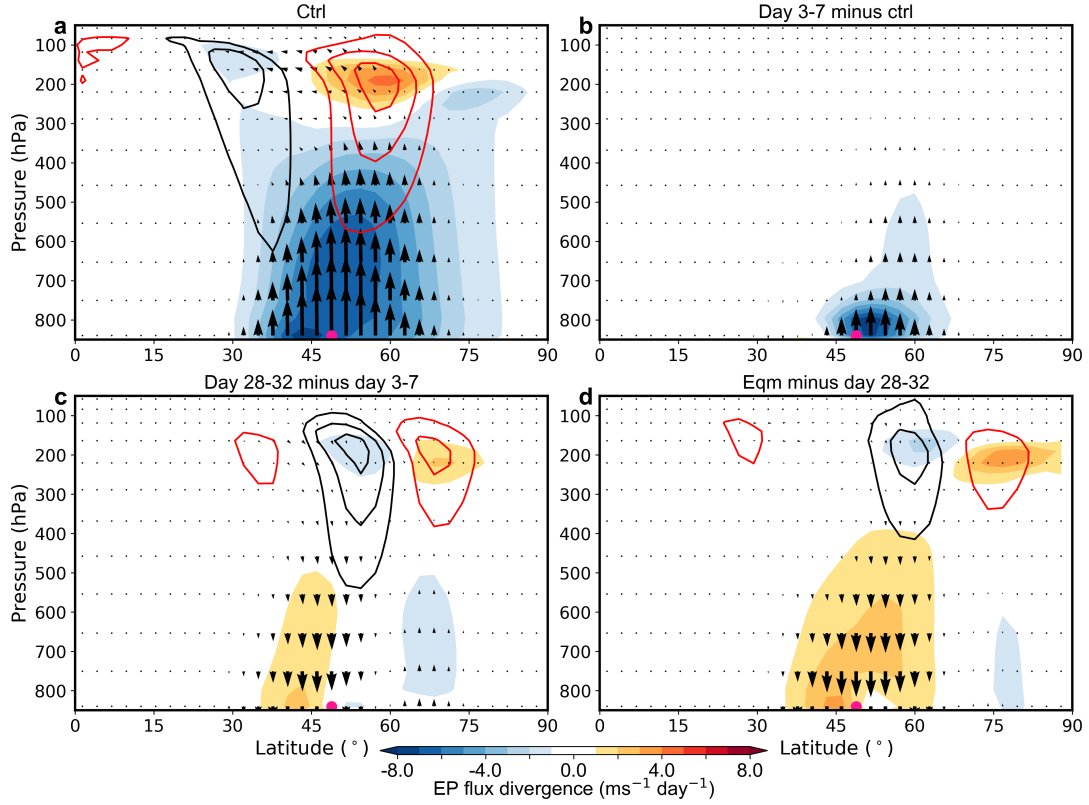


Figure 5.2: The EP cross section for heating applied at $\sigma = 0.9$ and $\phi = 49^\circ$ (latitude marked by pink circle). (a) the climatology, (b) day 3-7 average minus climatology, (c) day 28-32 average minus day 3-7 average, and (d) equilibrium response minus day 28-32 average. The shading denotes the eddy EP flux divergence ($\text{ms}^{-1} \text{day}^{-1}$) and the arrows denote the EP vectors [the arrow unit length in (a) represents an EP flux vector twice as large as in (b)-(d)]. The contours denote the eddy momentum flux convergence [red for convergence and black for divergence; interval: $0.5 \text{ ms}^{-1} \text{day}^{-1}$ for (a) and $0.25 \text{ ms}^{-1} \text{day}^{-1}$ for (b)-(d)].

the near surface $v'T'$ maximum, and anomalous divergence and upwards EP flux on the poleward flank of the $v'T'$ maximum. This manifests as a poleward shift in the maximum of $v'T'$. Correspondingly, there are changes in the meridional eddy momentum flux convergence; a decrease in convergence in the core of the eddy-driven jet, and an increase in convergence on the poleward flank of the jet associated with anomalous upper level horizontal eddy propagation. This initiates the shift in the zonal winds. As the atmosphere reaches equilibrium, the anomalous downwards EP flux relative to days 28-32 strengthens further and expands polewards (Fig. 5.2d). This leads to the equilibrated zonal wind response; a weakening and poleward shift of the eddy-driven jet.

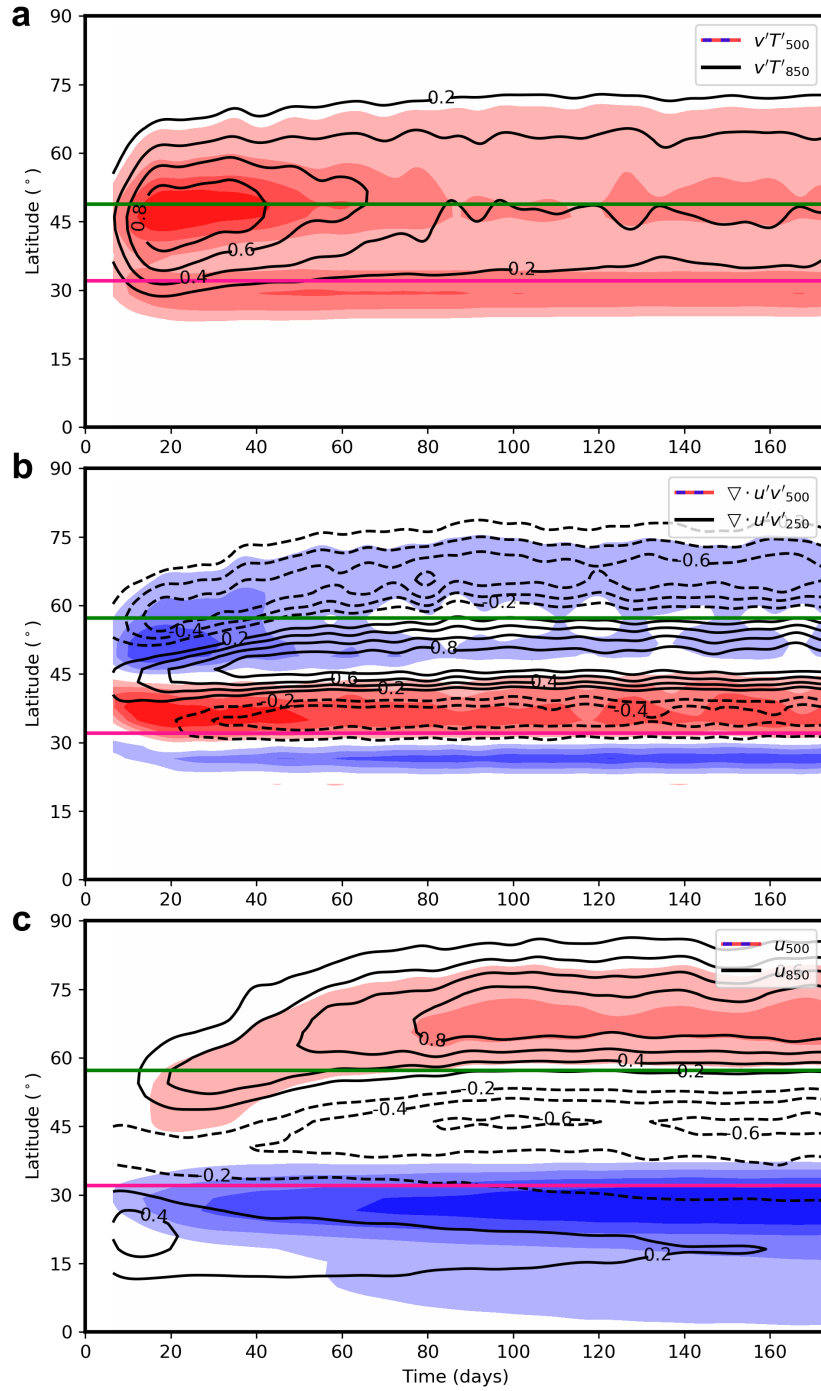


Figure 5.3: Hovmöller plots for heating applied at $\sigma = 0.6$ and $\phi = 32^\circ$. Normalised transient responses in (a) $v'T'$, (b) $\nabla \cdot u'v'$, (c) u . Contour intervals for the colours and contours are 0.2. The pink line indicates the heating latitude. The green line indicates the latitude of the corresponding index's maximum in the control run [e.g. in (c) the u maximum at $\sigma = 0.85$ in the control].

We now examine the other three case studies from Chapter 3. The second case study has heating applied at $\sigma = 0.6$ and $\phi = 32^\circ$, which sees a strengthening and poleward shift of the eddy-driven jet. The initial response (days 0-20) to the heating is an immediate increase in the maximum of $v'T'$ and an increase in the subtropical low level zonal winds (Fig. 5.3). Both these responses decay as the atmosphere equilibrates. There is also an immediate increase in the 500 hPa eddy momentum flux divergence, local to the heating perturbation. However, the response of the 250 hPa eddy momentum flux divergence behaves differently, with a weakening of the eddy momentum flux divergence just polewards of the heating at the upper levels (day 20). Next, the upper level meridional eddy momentum convergence starts to shift poleward (days 20-40). Finally, the lower-level zonal winds begin to shift poleward and strengthen, whilst the subtropical jet weakens (full response set up by day 80). Even though the eddy momentum fluxes respond very rapidly, the fact that the responses differ at 500 hPa and 250 hPa suggests that the 500 hPa responses occur as a local response to the imposed heating. The local 500 hPa eddy momentum flux response may be less important in setting up the final wind response which does not reach equilibrium until after the 250 hPa eddy momentum fluxes respond and begin to reach equilibrium.

A similar chain of causality occurs in the third case study, which has heating at $\sigma = 0.9$ and $\phi = -66^\circ$ (a similar location to the heating in case study 3 of Chapter 3, albeit lower in the troposphere). The heating causes a weakening and equatorward shift of the winter hemisphere eddy-driven jet. The initial response (days 0-15) to heating is again seen in $v'T'$ (Fig. 5.4), with a weakening on the poleward flank of its hemispheric maximum (note the heat flux maximum is negative in this hemisphere). The response is then conveyed to the eddy momentum fluxes, where the convergence weakens on the poleward flank of the eddy-driven jet and the region of maximum convergence shifts equatorwards (days 15-30). The zonal winds respond to these changes in convergence by weakening on the poleward flank of the jet as less momentum is converged into the jet (day 20 onwards). Finally, the jet shifts equatorwards with the increase in convergence

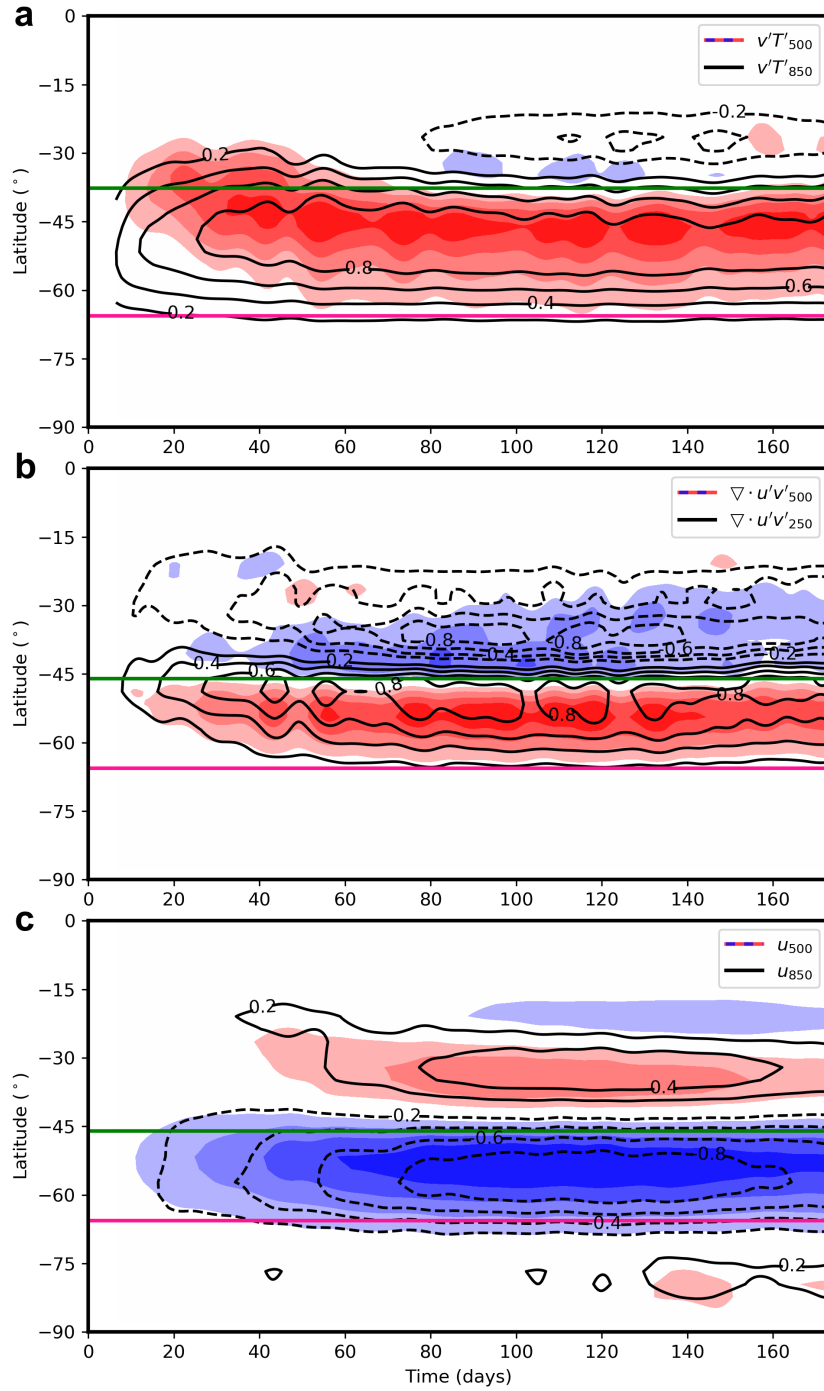


Figure 5.4: Hovmöller plots for heating applied at $\sigma = 0.9$ and $\phi = -66^\circ$. Normalised transient responses in (a) $v'T'$, (b) $\nabla \cdot u'v'$, (c) u . Contour intervals for the colours and contours are 0.2. The pink line indicates the heating latitude. The green line indicates the latitude of the corresponding index's maximum in the control run [e.g. in (c) the u maximum at $\sigma = 0.85$ in the control].

of momentum on its equatorward flank (final response set up day 80).

The final case study has heating at $\sigma = 0.4$ and $\phi = -32^\circ$, the same location as case four in Chapter 3. Heating here causes a weakening and poleward shift of the winter eddy-driven jet. The heating, being high in the troposphere and close to the change in sign of the meridional eddy momentum flux divergence, causes a local effect on $\nabla \cdot u'v'$ (Fig. 5.5). Nevertheless, there is still an almost immediate response in $v'T'$, with a weakening on its equatorward flank, and a slight strengthening further poleward (days 5-20). The upper level eddy momentum flux convergence weakens on the equatorward flank of the eddy-driven jet and then the maximum convergence slowly shifts polewards (days 10-40). This causes the eddy-driven jet to weaken and then shift poleward in line with the changes in momentum flux into the jet (full response set by day 90).

In all four of these cases examined, the overarching mechanism starts with the changes in $v'T'$ and baroclinic stirring, indicating a change in eddy growth. Anomalous eddy activity propagates upwards, leading to changes in the eddy momentum fluxes, which in turn affect the zonal wind and the eddy-driven jet. Therefore, the final equilibrated response is different from the initial response to the forcing. This mechanism explains how the changes in zonal winds respond to heating in the same hemisphere. As postulated in Chapter 3, the mechanism causing changes in the opposing hemisphere is mediated via the Hadley circulation, and we now explore spin-up simulations of the two Hadley cell case studies presented in Chapter 4.

The first of these cases has heating at $\sigma = 0.2$ and $\phi = 27^\circ$, which produces the greatest poleward shift of the winter Hadley cell and eddy-driven jet of all the heating locations in the summer hemisphere. Fig. 5.6 shows the initial response manifests as a change in the zonal wind in the same hemisphere as the heating. The Hadley circulation strengthens as shown by the tropical 850 hPa easterly zonal wind increase around days 20-30, and conveys the heating to the winter hemisphere (the greatest cross-hemisphere heating is located near the tropopause around -40° , see isentropes in Fig. 4.5c). This occurs after the summer jet responds but before the winter jet responds. Here, the same

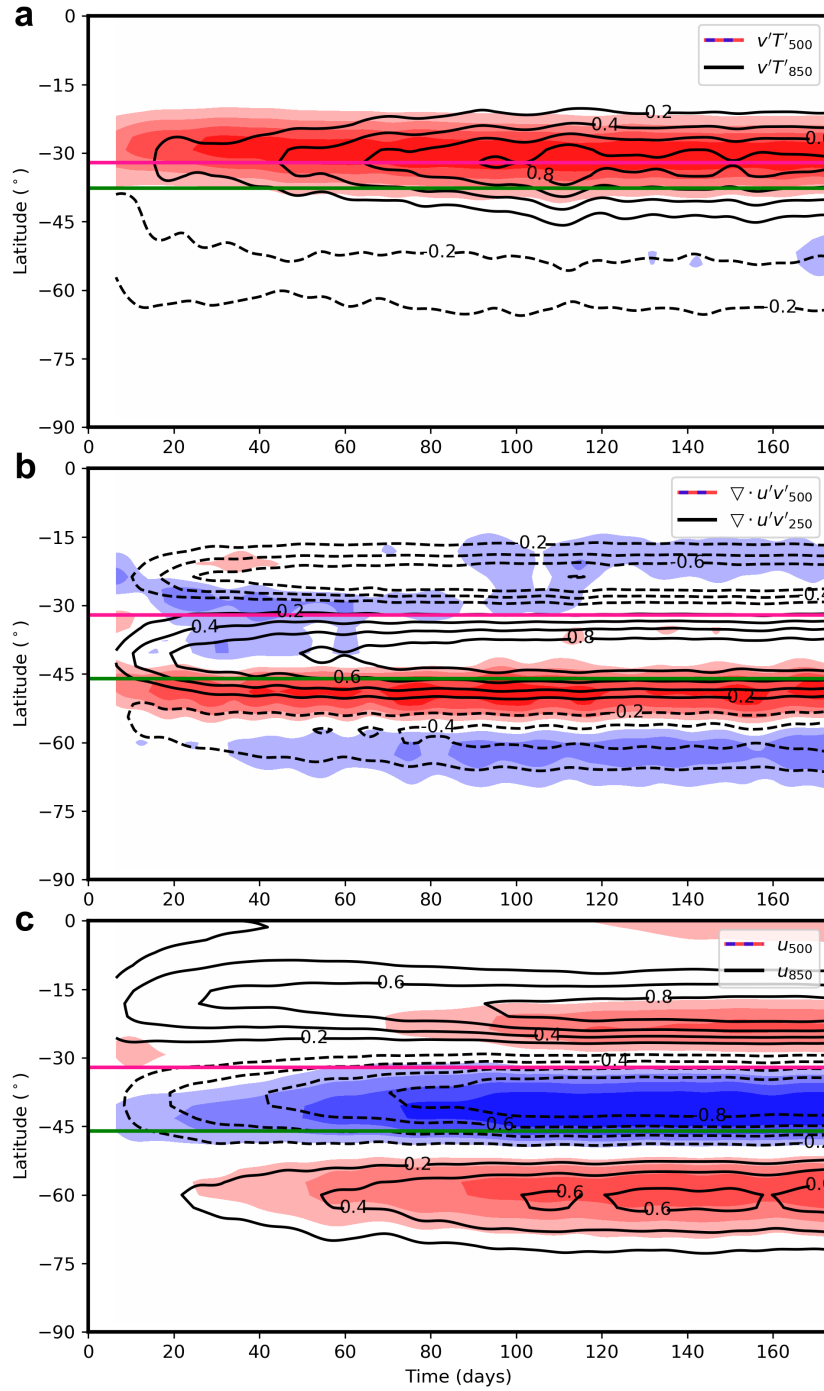


Figure 5.5: Hovmöller plots for heating applied at $\sigma = 0.4$ and $\phi = -32^\circ$. Normalised transient responses in (a) $v'T'$, (b) $\nabla \cdot u'v'$, (c) u . Contour intervals for the colours and contours are 0.2. The pink line indicates the heating latitude. The green line indicates the latitude of the corresponding index's maximum in the control run [e.g. in (c) the u maximum at $\sigma = 0.85$ in the control].

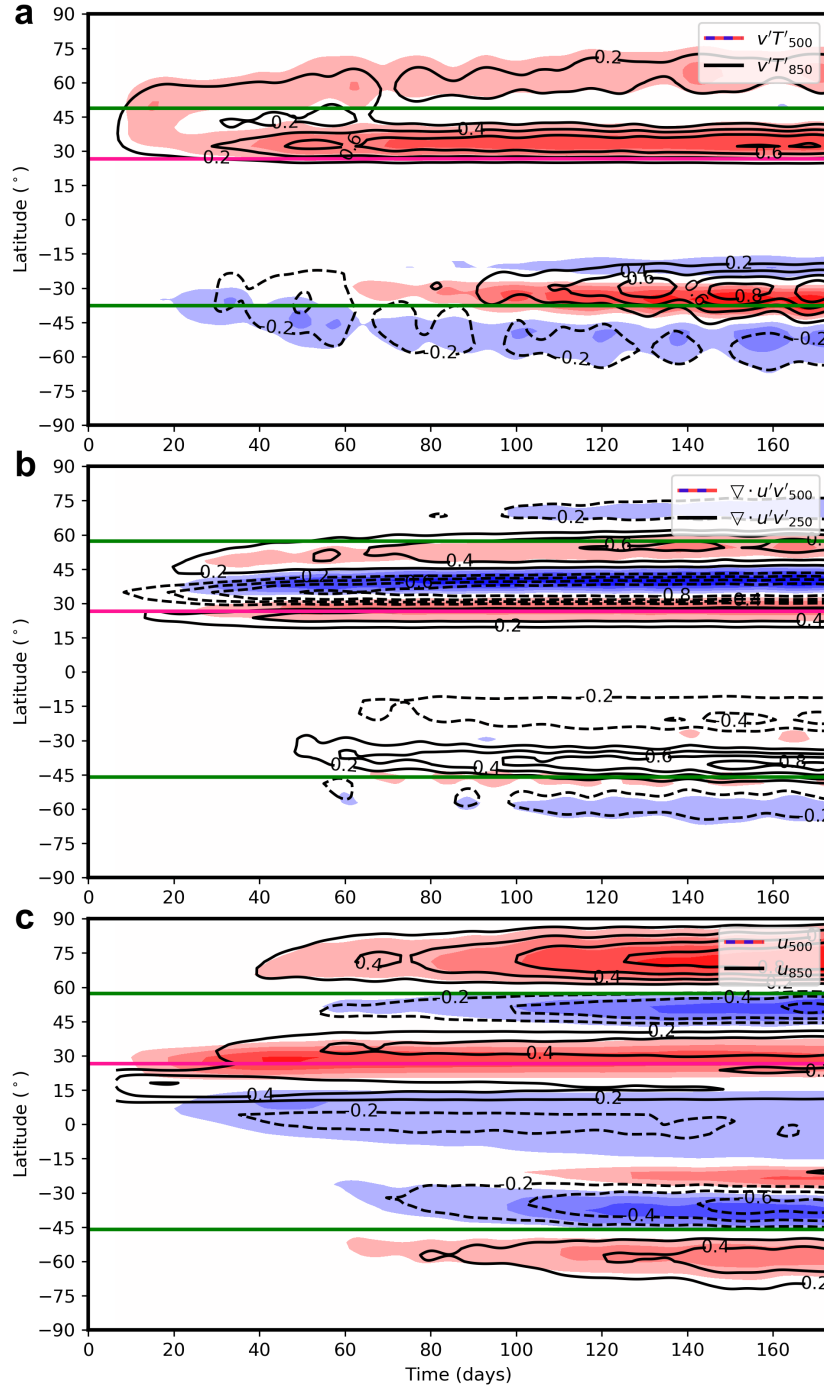


Figure 5.6: Hovmöller plots for heating applied at $\sigma = 0.2$ and $\phi = 27^\circ$. Normalised transient responses in (a) $v'T'$, (b) $\nabla \cdot u'v'$, (c) u . Contour intervals for the colours and contours are 0.2. The pink line indicates the heating latitude. The green line indicates the latitude of the corresponding index's maximum in the control run [e.g. in (c) the u maximum at $\sigma = 0.85$ in the control].

chain of causality in the previous case studies is observed, with a strengthening of $v'T'$ in the winter (day 20) hemisphere causing the eddy momentum convergence to shift poleward and consequently shift the eddy-driven jet poleward (around day 60). The shift takes around 80 days to first appear in the low-level part of the eddy-driven jet, much longer than the same hemisphere cases, as the heating takes time to reach the opposite hemisphere. The idea that the Hadley cell is responsible for conveying the heating to the winter hemisphere is supported by considering the typical timescale in the ascending branch of the Hadley cell. Taking a typical vertical velocity in the ascending branch of the Hadley cell of 0.05 Pa s^{-1} (e.g. Chen et al., 2014) gives a timescale on the order of 20 days. This is similar to the time it takes for the winter hemisphere to start responding to the summer heating, suggesting that the heating is transported directly to the winter hemisphere by the Hadley cell.

The EP fluxes allow further insight into how this response is set up (Fig. 5.7). Very little response is observed in Fig. 5.7b; it takes until around 50 days for the initial response to occur in the winter hemisphere (Fig. 5.7c). However, the zonal wind changes in the tropics (Fig. 5.6c) respond by day 10, supporting the idea that the winter Hadley circulation is primarily thermally driven, consistent with its larger local Rossby number. Once the heat has been transported to the winter hemisphere by the Hadley cell, the EP flux convergence and vertical EP flux vectors in the winter hemisphere respond (Fig. 5.7c). This supports the idea that the heat fluxes are affected before the eddy momentum fluxes. Fig. 5.7d shows that the equilibrated atmosphere has anomalous EP flux convergence on the equatorward side of the jet, and anomalous divergence on the poleward flank, acting to shift the jet poleward.

The second cross-hemispheric case study (shown in Fig. 5.8) has heating at $\sigma = 0.2$ and $\phi = -10^\circ$, which produces the greatest speed increase of the summer eddy-driven jet of all the heating locations in the winter hemisphere. Again, the Hadley circulation plays a key role in mediating the response of the jet. The heating strengthens the subtropical temperature gradients in the summer hemisphere, which causes a strengthening

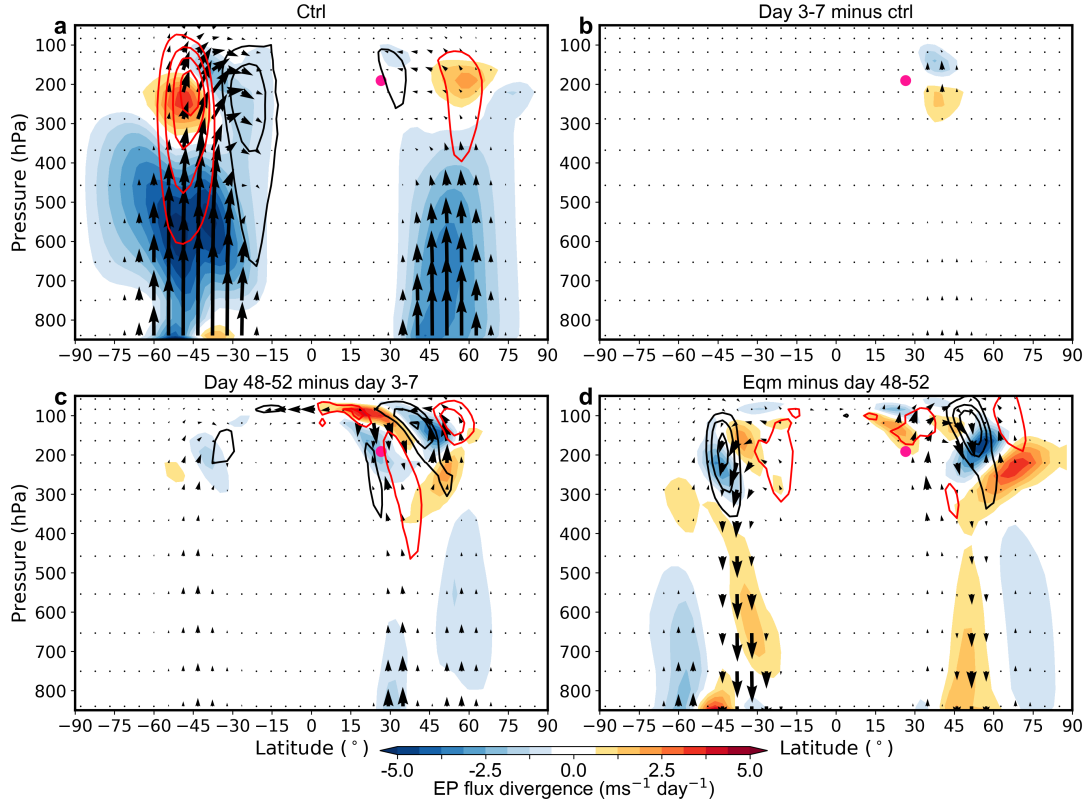


Figure 5.7: The EP cross section for heating applied at $\sigma = 0.2$ and $\phi = 27^\circ$ (latitude marked by pink circle). (a) the climatology, (b) day 3-7 average minus climatology, (c) day 48-52 average minus day 3-7 average, and (d) equilibrium response minus day 48-52 average. The shading denotes the eddy EP flux divergence ($\text{ms}^{-1} \text{day}^{-1}$), climatological divergences are divided by two to fit the colour scale. The arrows denote the EP vectors [the arrow unit length in (a) represents an EP flux vector twice as large as in (b)-(d)]. The contours denote the eddy momentum flux convergence [red for convergence and black for divergence; interval: $1.0 \text{ ms}^{-1} \text{day}^{-1}$ for (a) and $0.5 \text{ ms}^{-1} \text{day}^{-1}$ for (b)-(d)].

in the subtropical jet. In this case study, the changes in $v'T'$ in the summer hemisphere appear to be less important for the jet speed increase, but do play a role in shifting the jet equatorward (around 60 days). Instead, there is first an increase in eddy momentum flux convergence on the equatorward flank of the jet (days 30-40), which leads to more momentum being converged into the jet, thus a stronger jet.

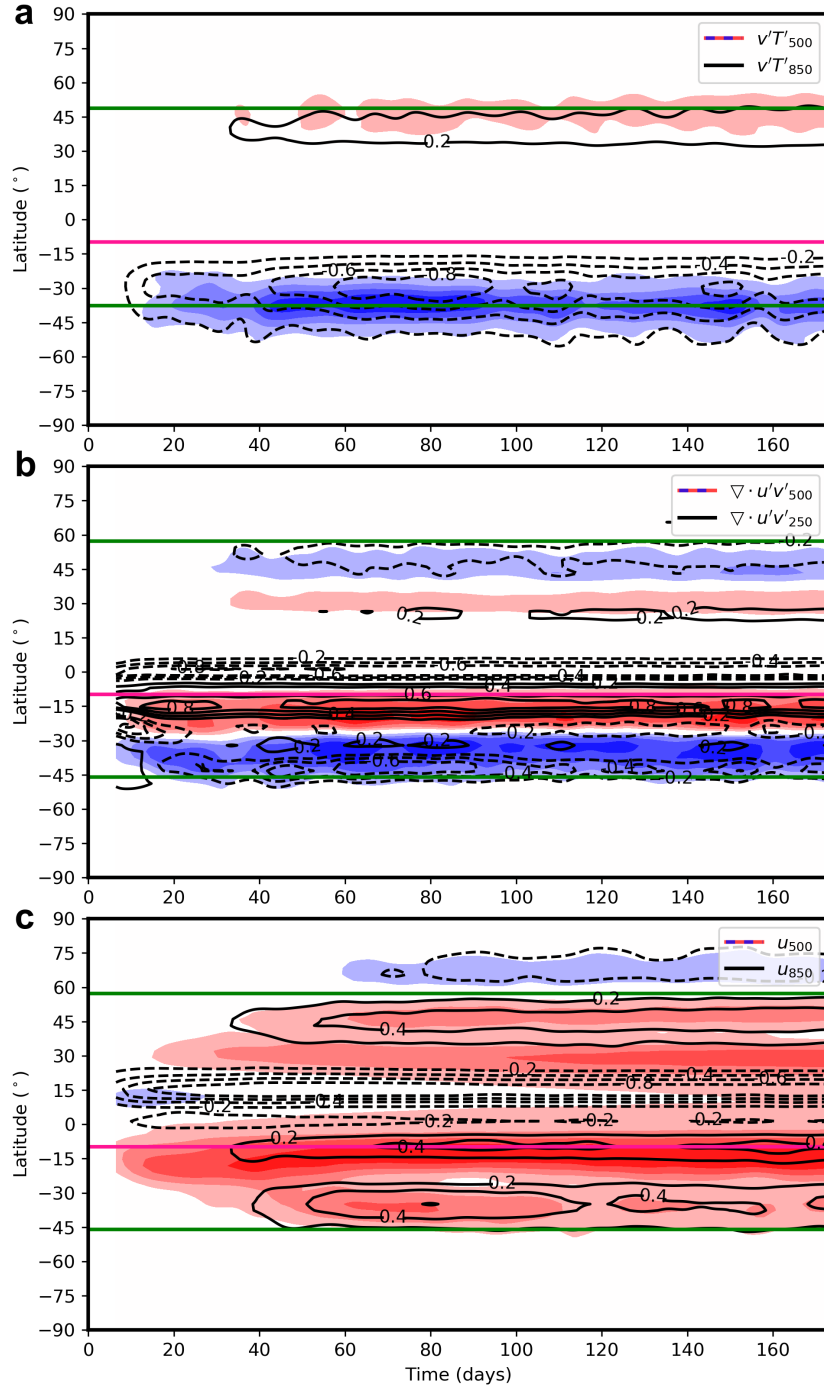


Figure 5.8: Hovmöller plots for heating applied at $\sigma = 0.2$ and $\phi = -10^\circ$. Normalised transient responses in (a) $v'T'$, (b) $\nabla \cdot u'v'$, (c) u . Contour intervals for the colours and contours are 0.2. The pink line indicates the heating latitude. The green line indicates the latitude of the corresponding index's maximum in the control run [e.g. in (c) the u maximum at $\sigma = 0.85$ in the control].

5.3 Conclusions

This chapter has explored, through the use of spin-up ensembles, the mechanisms by which the eddy-driven jet responds to local diabatic heatings. Consistent with previous work (Ring and Plumb, 2007), the role of the eddies is vitally important: without eddy-mean flow interactions, the zonal wind responses can be opposite in sign, and are typically much less barotropic. Our findings suggest that the predominant mechanism by which the jet responds is as follows: changes in heat fluxes and low level baroclinic stirring modify the eddy momentum fluxes which in turn cause the jet to respond as regions of momentum convergence and divergence shift and change in magnitude. The use of heating perturbations rather than mechanical perturbations (cf. Ring and Plumb, 2008) highlights the importance of the changes in heat fluxes in setting up the zonal wind response, as the heating projects directly on to temperature gradients and hence heat fluxes.

There are a multitude of theories to explain changes in the eddy-driven jet due to forcing. Our results point towards the theory that changes in low-level baroclinicity are responsible for the changes in the jet, as in agreement with previous work (Yin, 2005; Brayshaw et al., 2008; Frierson, 2008; Lu et al., 2010; Butler et al., 2011; Mbengue and Schneider, 2017). With regards to the wave propagation changes theory proposed by Chen and Held (2007), is that under their theory, a poleward shift of the jet would be accompanied by a strengthening of the jet due to eddy phase speed arguments. They propose that increases in eddy phase speeds prevent eddies propagating as far equatorward, thus converging their momentum into the jet further poleward, strengthening and shifting the jet poleward. However, it can be seen from the case studies presented here, a poleward shift of the jet is not always associated with a strengthening. That said, our results do not conflict with other theories. For example, in the Rossby wave mechanism put forward by Lorenz (2014), it is possible for the jet to shift poleward and weaken on its poleward flank. A zonal wind increase on the poleward flank of the jet causes a wider

range of poleward propagating waves to be reflected before the poleward flank of the jet. This leads to increased equatorward propagating waves directly opposing the poleward shift which can cause the jet to weaken (Lorenz, 2014).

The mechanisms put forward here are all diagnosed using a dry idealized GCM. In the real atmosphere there are much more complex processes and dynamics at play (e.g. Ceppi and Shepherd, 2017), with clouds and other moist processes potentially contributing to the heating that the dry dynamics are responding to. However, 80% of the storm track shifts seen in CMIP5 can be reproduced by dry dynamics (Mbengue and Schneider, 2013), suggesting that most of the important processes and mechanisms are captured in dry GCMs. Further work using more complex GCMs with moist processes could confirm the proposed chain of causality here. Our forcings are only applied in a zonally symmetric manner, which is useful when considering zonal mean climate changes such as Arctic amplification or upper-tropospheric tropical warming. However, in other cases, such as changes to the Indian monsoon where forcings are longitudinally localized, other responses and mechanisms may be important. To consider these types of forcing, in the next chapter we use a more complex GCM to investigate how more local forcings, applied as sea surface temperature anomalies, cause the atmosphere to respond.

Chapter 6

Circulation sensitivity to SSTs in an atmosphere-only GCM

In the previous three chapters, we investigated the circulation response to idealized forcing perturbations in a simple GCM. We now turn our attention to the sensitivity of the circulation in a more complex, atmosphere-only, GCM, HadAM3P. Instead of applying idealized forcing perturbations as diabatic heating in the atmosphere, we will apply perturbations to the sea surface temperatures. We will compare and contrast our results with the idealized findings, and explore potential uses for the circulation sensitivities in forecasting and attribution.

6.1 Introduction

The North Atlantic Oscillation (NAO) is a teleconnection pattern that dominates the climate variability over the North Atlantic (Wallace and Gutzler, 1981; Hurrell, 1995). Partly resulting from the internal variability of midlatitude dynamics, it is also externally driven (e.g. Hurrell et al., 2013). Sea surface temperatures (SSTs) in many different locations have been shown to be important external NAO drivers (e.g. Rodwell et al., 1999; Hoerling et al., 2001; Czaja and Frankignoul, 2002; Peng et al., 2002; Bader and Latif,

2003; Wang et al., 2004; Kucharski et al., 2006). However, due to internal variability and signal-to-noise issues, one requires either long observational datasets or large climate ensembles to say with confidence where and to what extent SSTs are responsible for forcing the NAO.

There have been numerous studies investigating the winter NAO response to SST forcing. The recent warming trend in the Indian Ocean has been shown to force a positive NAO (Hoerling et al., 2001, 2014; Bader and Latif, 2003, 2005). Whilst there were concerns about spurious results arising from missing air-sea coupling leading to incorrect sea level pressure (SLP) in these studies (Copsey et al., 2006), Colfescu et al. (2013) showed that the incorrect SLP may in fact arise from model bias, thus not ruling out the Indian Ocean-NAO link. Further to this, Fletcher and Cassou (2015) also demonstrate the link using a coupled model. The mechanism postulated is the generation of a Rossby wave from the SST warming which constructively interferes with a climatological stationary wave, thus forcing a positive NAO response (Fletcher and Kushner, 2011). However, the simulated NAO trend amplitude underestimates the observed trend, suggesting that other drivers may also be important (King et al., 2010). The Atlantic tripole SST pattern has also been shown to force the NAO (Rodwell et al., 1999; Czaja and Frankignoul, 2002; Peng et al., 2002; Wang et al., 2004), although the tripole pattern in the SSTs can also be considered as a combined SST/atmospheric response (Cassou et al., 2007). Further to this, the Pacific has also been linked to forcing the NAO (Kucharski et al., 2006; Fletcher and Kushner, 2011). These numerous studies employ a range of methodologies to find the links between SSTs and the winter NAO, which makes it hard to compare and identify the SST forcing importance in each location. In contrast, there are only a handful of studies looking at the summer NAO response to SSTs, which indicate some connection to the Atlantic Multidecadal Oscillation (Folland et al., 2009; Sutton and Dong, 2012). However, given that the summer NAO is the dominant pattern of variability in the North Atlantic (Folland et al., 2009; Bladé et al., 2012), a systematic study of the summer NAO sensitivity to SSTs would provide useful information for

predicting and attributing particular summer NAO phases.

Skilfully predicting the winter NAO has, until recently, proved elusive (Scaife et al., 2014), due to climate models showing little response to slowly varying climate system components such as the ocean (Kushnir et al., 2006). Recently, high predictability of the winter NAO has been observed using the Met Office Global Seasonal forecast System 5 (Scaife et al., 2014). A portion of this skill is shown to come from North Atlantic SSTs (Smith et al., 2016). Using a global sensitivity map of the NAO to SSTs, it is possible to reconstruct the winter NAO time series (Li and Forest, 2014) and investigate if there is any skill in reconstructing and forecasting the NAO and if so where it comes from. To date, no skill has been reported for predicting the summer NAO in seasonal forecast systems (Dunstone et al., 2018).

Early studies computed the sensitivity of atmospheric variables to SSTs using a Green’s function approach via the systematic application of patches across the ocean domains (Barsugli and Sardeshmukh, 2002; Barsugli et al., 2006). However, later work (Li et al., 2012; Li and Forest, 2014), uses a random perturbation method (RPM), whereby random SST fields are generated with anomalies across the ocean domain. The two methods give consistent sensitivity information and similar reconstruction of the regional response (Li et al., 2012). However, the RPM method is about twelve times more computationally efficient than the patch method due mainly to the larger area-integrated SST forcing amplitude.

Here we build on the work of Li and Forest (2014), using a large climate ensemble to investigate the winter and summer NAO sensitivities to SST forcing. We also investigate the eddy-driven jet sensitivity to SSTs as the jet is intimately linked to the NAO. We then use these sensitivities to reconstruct NAO and jet index time series, and explore where the skill in these reconstructions comes from. Section 6.2 describes the reanalysis data used, how the ensemble simulations are designed, and how the linear statistical-dynamic method works. In Section 6.3 we present our findings, along with a discussion of the results in the context of previous studies, before concluding in Section 6.4.

6.2 Methods

6.2.1 Reanalysis datasets

Throughout this chapter, we make use of three reanalysis datasets to compare to and validate our model results. These are: the NOAA-CIRES 20th Century Reanalysis, hereafter NOAA-20C (Compo et al., 2011), the ERA-20C Reanalysis, hereafter ERA-20C (Poli et al., 2016), and the NCEP-DOE Reanalysis 2, hereafter NCEP2 (Kanamitsu et al., 2002). NAO time series for each reanalysis are computed by projecting each individual season onto the first EOF of the reanalysis SLP for that particular season (e.g. the ERA-20C winter NAO time series is computed by projecting the December-January-February (DJF) mean SLP for each year onto the first EOF of the DJF ERA-20C SLP). The NAO EOF is computed over the region 20°N-80°N, 90°W-40°E. Throughout the chapter, we refer to winter as the DJF period, and summer as the June-July-August (JJA) period.

6.2.2 RPM ensemble

We make the assumption that the large-scale atmosphere responds to SST forcing with a linear component. Following Li et al. (2012), the relationship between forcing and response can then be represented as a continuous Green's function:

$$\Delta R(\mathbf{x}) \approx \int G(\mathbf{x}, \mathbf{x}') \Delta SST(\mathbf{x}') dA, \quad (6.1)$$

where ΔR is the atmospheric response anomaly, $\Delta SST(\mathbf{x}')$ is the related SST anomaly located at $\mathbf{x}'$ with area dA , and G is a linear operator. This linear operator, the global teleconnection operator (GTO), is a set of sensitivities describing the atmospheric response to SST forcing at each gridpoint.

To allow us to compute the GTO, we use a large ensemble of perturbed SST runs. We use the RPM to generate an ensemble of SST perturbation fields as described in Li et al.

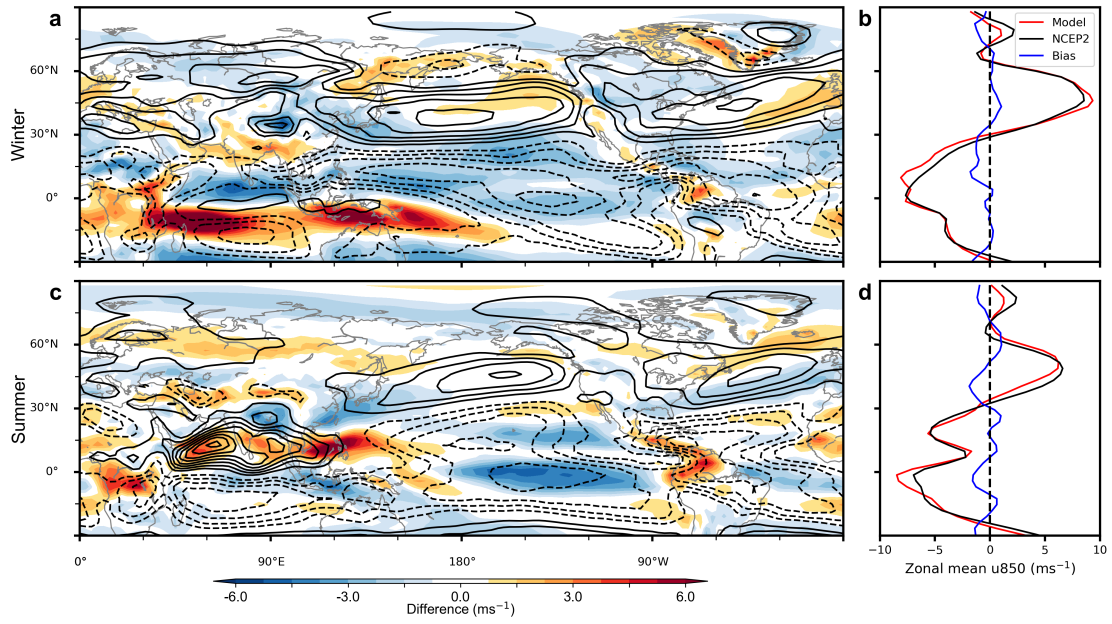


Figure 6.1: Mean 850 hPa zonal wind biases in HadAM3P in (a, b) DJF and (c, d) JJA. The contours (at 2 ms^{-1} intervals) show the mean jet climatology from the NCEP2 reanalysis (1979-2017), whilst the colours show the difference between the ensemble mean of the control simulations and the reanalysis. The zonal means in panels (b) and (d) are computed over 60°W - 0° .

(2012). Following Li et al. (2012), each random perturbation field is created as a 16×16 matrix, with each point having a temperature value randomly drawn from a uniform temperature distribution between -2 K and 2 K . The matrix is then bilinearly interpolated onto a grid with N96 resolution (145 latitude points by 192 longitude points). To ensure local maxima and minima do not just occur on the original 16×16 grid, we shift each initial coordinate randomly by 0-8 gridpoints zonally and 0-5 gridpoints meridionally.

We generate 5000 SST perturbation fields and add them to the 1980-1999 climatological monthly SSTs in the merged Hadley-NOAA/OI SST dataset (Hurrell et al., 2008) to create a 5000 member ensemble. Perturbation fields are applied between 60°N - 60°S , and to ensure a smooth decrease at the boundaries, we weight the perturbations between 50° - 60° (both north and south) with linearly decreasing weights from 0.9 to 0.1.

We use the HadAM3P model to run the ensemble. HadAM3P is an atmosphere only, medium resolution (run here at N96 resolution, or $1.25^\circ \times 1.875^\circ$), GCM developed by the UK Met Office. It is based upon the atmospheric component of HadCM3 (Pope et al.,

2000; Gordon et al., 2000). We use an improved version of HadAM3P which has a more sophisticated land-surface scheme (Guillod et al., 2017). We run HadAM3P using the large-ensemble capability provided by the climateprediction.net volunteer computing network (Allen, 1999; Guillod et al., 2017), where members of the public run initial condition ensemble GCM simulations. The HadAM3P representations of many midlatitude Northern Hemisphere winter circulation features, including the tri-modal Atlantic eddy-driven jet structure, the Atlantic storm track and Atlantic blocking, have been shown to be better than the majority of models in CMIP5 and associated atmosphere-only models (Mitchell et al., 2017b). The 850 hPa zonal wind biases are shown in Fig. 6.1, showing that in winter and summer the model North Atlantic jet biases are small. Biases are also small in the other midlatitude regions, suggesting the model should simulate teleconnections well too due to a low waveguide bias. The ability of HadAM3P to capture these dynamical phenomena well makes it a good model choice for assessing the midlatitude circulation sensitivity to SSTs.

For each ensemble member we run a control simulation and forced simulation. The control simulation is run for 25 months (December 1998 to December 2000), and uses the 1980-1999 Hadley-NOAA/OI climatological mean monthly SSTs, and observed greenhouse gas forcings from 1998-2000. The forced simulation is exactly the same, but with the SST perturbation applied, giving 5000 control and forced ensemble member pairings. Each pairing also has an initial condition perturbation via perturbations to the potential temperature (which is the same for both runs in the pair), for full details see Massey et al. (2015). We discard the first year of each simulation, giving us one winter and one summer season for each ensemble member. We define each ensemble member response as the difference between each forced and control ensemble member pair.

6.2.3 The GTO and linear reconstructions

Following Li et al. (2012), we compute the GTO for a particular atmospheric response variable by employing a linear regression between the seasonal atmospheric anomaly and the SST anomalies:

$$R = G\Delta SST + e, \quad (6.2)$$

where e is the error related to the nonlinearity and the internal noise. Using the result derived in the appendix of Li et al. (2012), the smoothed GTO estimate is given by:

$$G(\mathbf{x}_i) = \frac{Cov(\Delta SST_n(\mathbf{x}_i), R_n)}{\frac{T_{\max}^2}{3} L_x L_y} + e, \quad (6.3)$$

where G is computed at each gridpoint $\mathbf{x}_i$. $\Delta SST_n(\mathbf{x}_i)$ is a vector of the 5000 SST anomalies at the gridpoint $\mathbf{x}_i$, R_n is a vector of the 5000 atmospheric anomaly indices, $T_{\max} = 2$ K, the maximum SST perturbation, and $L_x = a\lambda_0 \cos(\phi_i)$ and $L_y = a\phi_0$ are the RPM lattice local length scales (a is the Earth's radius, ϕ_0 and λ_0 are the latitude and longitude gridpoint spacings in radians, and ϕ_i is the latitude of the gridpoint for which G is being computed for).

Once the GTO has been obtained, it can be used to reconstruct the response of an atmospheric variable to a particular SST field. This is achieved by projecting the SST field onto the GTO:

$$\Delta R = \gamma \sum_{\mathbf{x}_i} G(\mathbf{x}_i) \Delta T_{\text{SST}}(\mathbf{x}_i) L_x L_y, \quad (6.4)$$

where ΔT_{SST} refers to the SST anomaly of the particular SST field one wishes to reconstruct the atmospheric response to. The coefficient γ depends on the length scales of both the forcing and response for a given target response based on the dynamics (Li et al., 2012). Although the scaling by the γ coefficient does not affect correlation results between reconstructed and reanalysis time series, it is insightful to be able to investigate the size of the signal of the reconstructed response. In order to scale the reconstructed

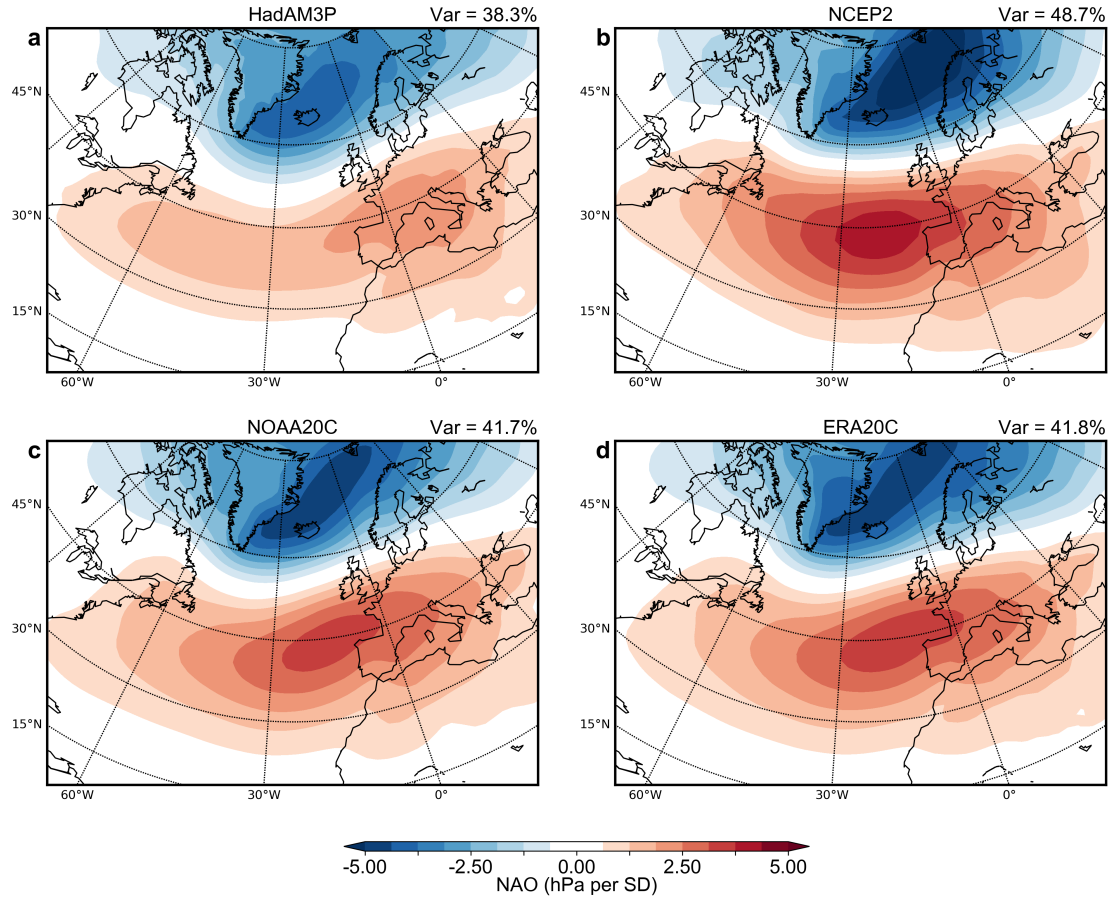


Figure 6.2: SLP regressed onto the 1st principal component time series of DJF SLP over 20°N-80°N, 90°W-40°E. (a) HadAM3P, (b) NCEP2 (1979-2017), (c) NOAA-20C (1871-2012), (d) ERA-20C (1900-2010). Numbers indicate the percentage of the variance the EOF explains.

response to have physical meaning, we perform a simple linear regression of the reconstructed time series onto the NOAA-20C reanalysis time series for each variable. The reconstructed time series is then scaled by the regression coefficient, i.e.:

$$R_{\text{scaled}} = \alpha R_{\text{reconstructed}}, \quad (6.5)$$

where α is the regression coefficient determined by the regression $R_{\text{reanalysis}} = \alpha R_{\text{reconstructed}}$.

In this chapter we investigate the GTO for the NAO and North Atlantic eddy-driven jet. For each ensemble member pair, we define response NAO and jet indices. The response NAO for each member is defined by the difference of the projections of the

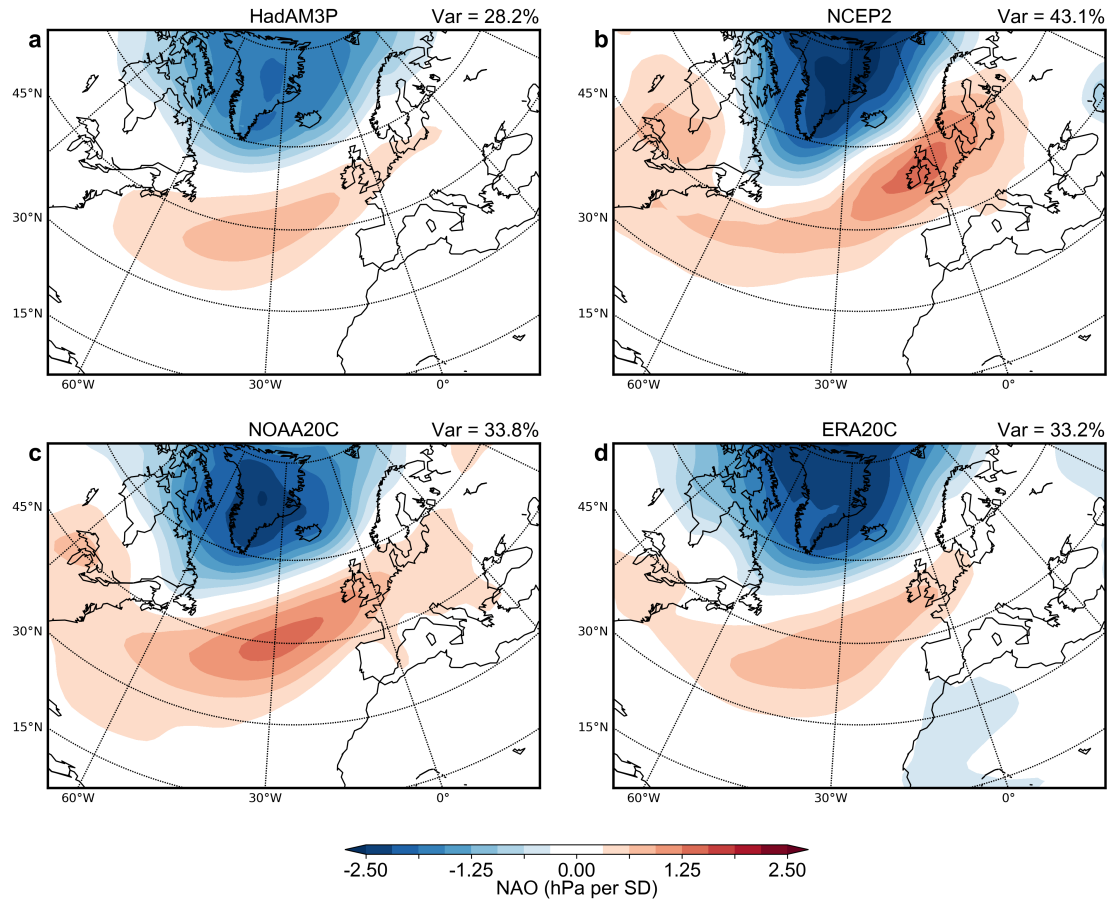


Figure 6.3: SLP regressed onto the 1st principal component time series of JJA SLP over 20°N-80°N, 90°W-40°E. (a) HadAM3P, (b) NCEP2 (1979-2017), (c) NOAA-20C (1871-2012), (d) ERA-20C (1900-2010). Numbers indicate the percentage of the variance the EOF explains.

forced and control seasonal mean SLP fields onto the 1st model EOF of SLP over 20°N-80°N, 90°W-40°E. The model and reanalysis EOFs are shown in Fig. 6.2 and Fig. 6.3 for winter and summer respectively. The pattern correlations between any two given EOFs in a particular season are all greater than 0.9, and using the model EOF or a reanalysis EOF does not significantly affect the calculation of the GTO (not shown). The jet speed and latitude responses are taken as the difference between the forced and control indices as defined using the jet indices in Woollings et al. (2010), defined over 15°N-75°N, 60°W-0°. We use daily data at 850 hPa to compute the jet indices, and then compute the mean index for each season before taking the difference to compute the response.

To reconstruct long time series of the NAO and jet indices from the GTO, we use the

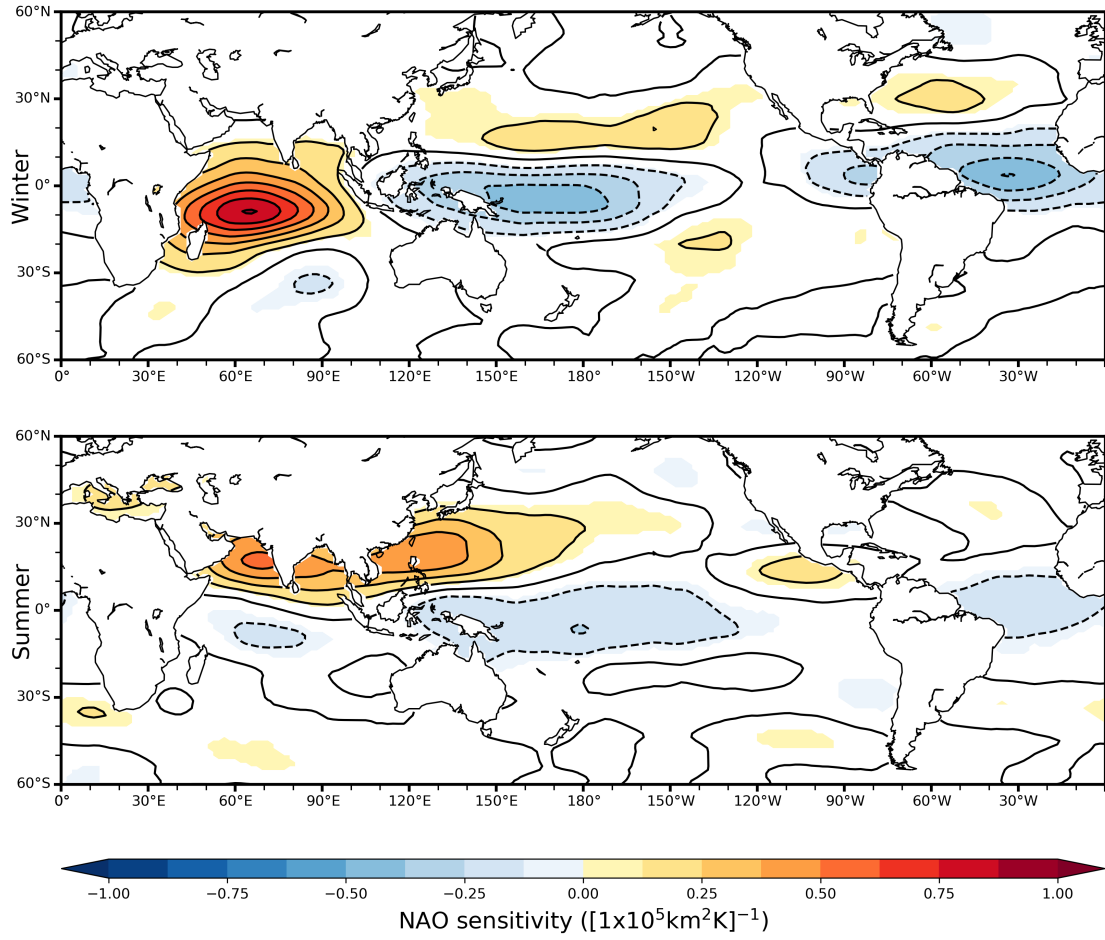


Figure 6.4: Sensitivity of the NAO index to SST anomalies given by the $G(x')$ in Eq. 6.1. Shaded regions indicate statistical significance at the 5% level using a two-tailed *t*-test.

HadISST dataset (Rayner et al., 2003). We project the monthly SSTs from the dataset onto the GTO and then take the mean across each season to define a reconstructed index for a particular year (e.g. to compute the summer NAO for a particular year, we project June, July and August SSTs separately onto the summer NAO GTO, then take the mean of the three projections to get the reconstructed NAO for that year).

6.3 Results and Discussion

6.3.1 The North Atlantic Oscillation

Fig. 6.4 shows the GTO for the NAO in winter and summer. The maps display the NAO sensitivity to anomalous SSTs. For example, the winter NAO exhibits a positive response to anomalous warming over the Indian Ocean. The winter GTO bears strong resemblance to the winter GTOs computed for the CAM5 and GFDL models, shown in Fig. 3 of Li and Forest (2014). It is worth noting that Li and Forest (2014) use 200 member ensembles to compute their GTOs, and so there are fewer regions of statistical significance compared to the HadAM3P GTO. The strong region of winter NAO sensitivity to Indian Ocean SSTs corroborates previous studies linking the warming trend in Indian Ocean SSTs and the positive NAO trend from 1950-2000 (e.g. Hoerling et al., 2001, 2004; Bader and Latif, 2003, 2005).

The winter NAO also appears sensitive to an SST dipole in the tropical Pacific, and a tripole pattern in the North Atlantic. Kucharski et al. (2006) link a Pacific dipole pattern to the winter NAO, however their dipole pattern consists of warm SSTs in the tropics and cooler SSTs in the northern subtropics. There are several reasons why a discrepancy may arise between our result and their work. The GTO only picks out the linear sensitivity to SSTs; the GTO is constructed so that it captures the response to an anomalous SST pattern throughout the year and so does not provide any temporal information about when the NAO is sensitive to the SSTs (i.e. it may be the October SSTs in the Pacific that are important for forcing the winter NAO. This will be investigated later in the chapter). Kucharski et al. (2006) find the link between the low frequency variability in the Pacific and the NAO.

The North Atlantic tripole pattern has also been studied and found to be associated with the winter NAO (e.g. Rodwell et al., 1999; Czaja and Frankignoul, 2002; Peng et al., 2002; Wang et al., 2004). The GTO picks out the same sign of tripole pattern (cold-

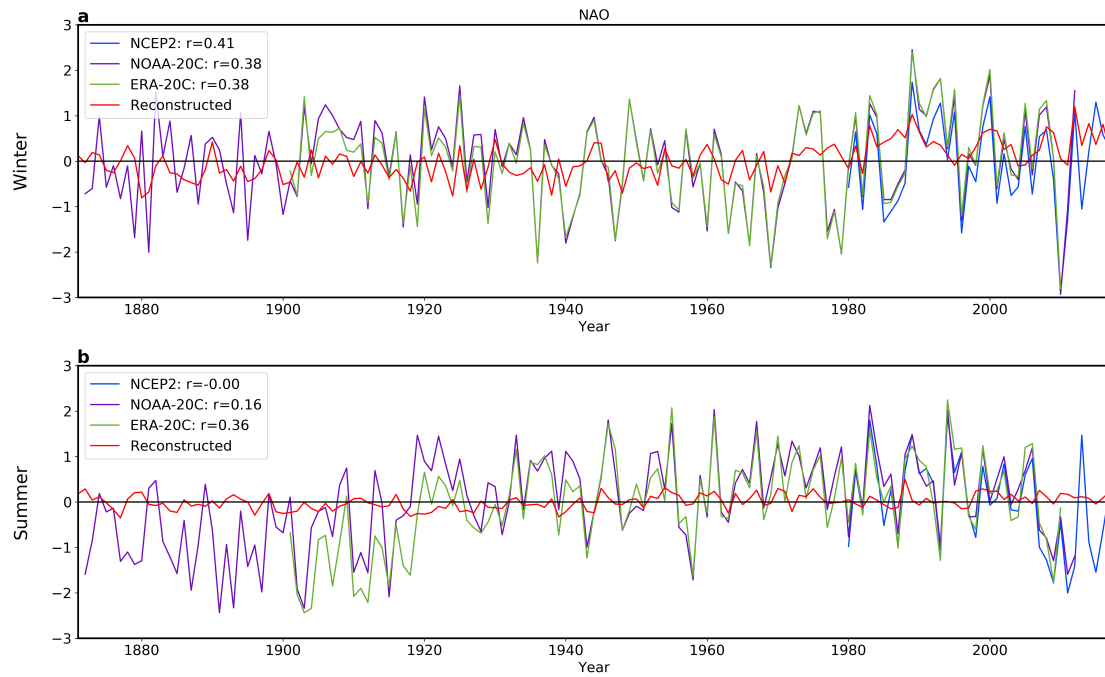


Figure 6.5: Time series of the NAO in (a) winter and (b) summer. The reconstructed NAO time series is calculated using Eq. 6.4, and then scaled using Eq. 6.5. For winter, the year refers to the year in which January and February fall.

warm-cold) as in the literature that is associated with a positive NAO. On the question of causality between the SST anomalies and the NAO, the GTO method takes prescribed SST patterns and looks at their impact on the NAO. The sensitivity map given by the GTO therefore only includes information of the model response to SSTs, and does not take into account any influence from the atmosphere onto the SSTs (discussed later).

The summer NAO also appears to be sensitive to SSTs, in similar regions to the winter NAO. The northern Indian Ocean and subtropical west Pacific are associated with a positive summer NAO, whilst the southern Indian and tropical western Pacific are associated with a negative summer NAO. The tripole pattern is also visible, except only the tropical Atlantic displays a statistically significant sensitivity region. The weaker, yet still statistically significant, sensitivity map for the summer NAO suggests that there may be some skill in reconstructing the summer NAO, something which has previously not been reported in any forecast systems (Dunstone et al., 2018).

Using Eq. 6.4 and Eq. 6.5, we reconstruct winter and summer NAO time series

from the HadISST dataset. Shown in Fig. 6.5 are the reanalysis and reconstructed time series, along with the correlation coefficients between each reanalysis time series and the reconstructed time series. The reconstructed winter NAO is significantly (at the 5% level using a two-tailed t -test) correlated with all three reanalysis products. By construction, due to the scaling method applied in Eq. 6.5, the reconstructed time series standard deviation is equal to the correlation with the NOAA-20C time series. The reconstructed time series captures the trend towards more positive NAO values later in the time series, suggesting this trend is forced by SSTs. This will be explored later in the chapter. Only the ERA-20C time series is correlated significantly with the reconstructed summer NAO. Given the high correlation between the three reanalysis time series, this suggests that the teleconnection strength between the SSTs and the summer NAO varies temporally. The reconstructed summer NAO also fails to capture the trend in recent years to a more negative summer NAO, suggesting this trend is not being forced, at least linearly, by changes in SSTs.

Given the apparent varying reconstruction skill in summer, we proceed to calculate the correlations between the reanalysis and reconstructed NAO time series using a 31-yr moving window (shown in Fig. 6.6). We assess statistical significance at the 5% level using a two-tailed t -test. Autocorrelation is accounted for following Wilks (2011) by using the lag-1 autocorrelation of the full NOAA-20C and reconstructed NAO time series to determine the effective sample size of the 31-yr period. In winter, the skill is fairly consistently around 0.4, with a statistically significant correlation between the reconstructed and reanalysis time series across much of the period. Interestingly, we do not observe the same level of drop off in skill as seen by Weisheimer et al. (2017) in the 1950-70s, who use atmosphere only hindcast experiments to investigate the skill in NAO forecasts. This could be due to inter-model differences in the teleconnection strength between SSTs and the NAO, or other processes simulated poorly in the atmosphere-only model which do not contribute in the GTO method. The winter NAO skill using the GTO method is also consistently equal to or higher than that in Weisheimer et al. (2017). In

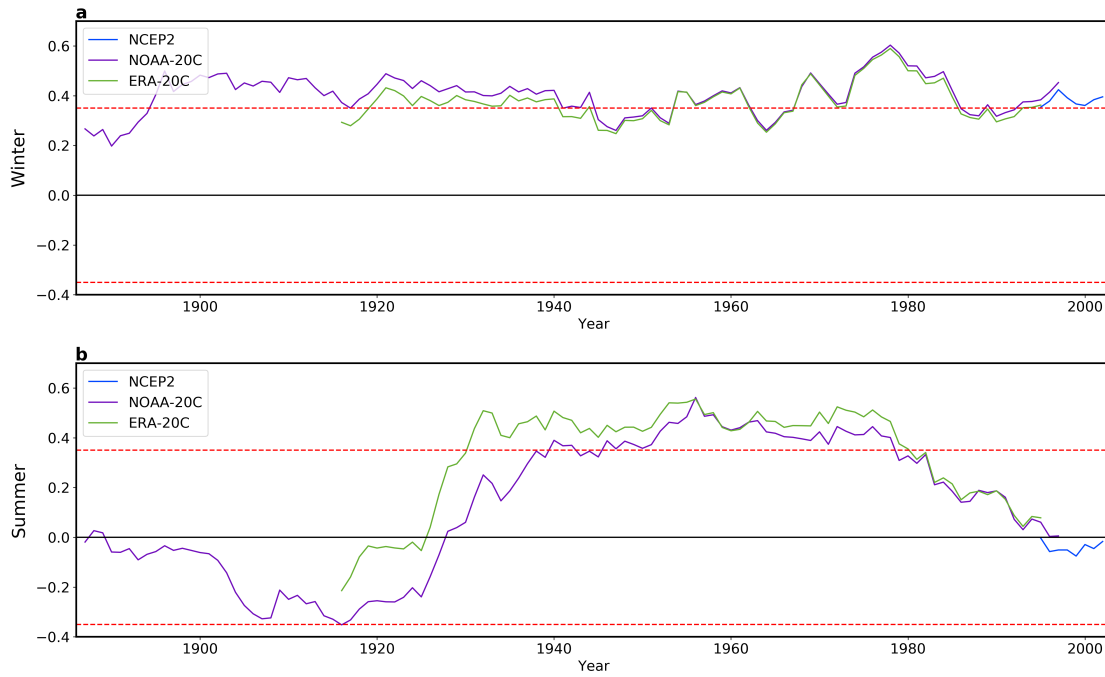


Figure 6.6: 31-yr moving correlation of the reconstructed NAO with the NAO times series from the reanalyses in (a) winter, (b) summer. the red dashed lines indicate statistical significance at the 5% level using a two-tailed t -test. The year refers to the middle year of the 31-yr moving window.

summer, as mentioned earlier, the skill varies across the period. Before 1915, the skill is very poor, but then rises through to 1925, remaining high until 1995, before declining again. The reason for this time varying skill will be explored later in the chapter. In this high skill period, the correlation is statistically significant and consistently between 0.4 and 0.6. We believe that this high level of skill in reconstructing the summer NAO has not been shown in any previous studies.

We now return to the issue of atmosphere-ocean coupling. Barsugli and Battisti (1998) investigate its effect on midlatitude variability, finding that the coupling in the midlatitudes is important for correctly simulating the low-frequency atmospheric thermal variance. Bretherton and Battisti (2000) extend the analysis further, showing that the useful predictability associated with midlatitude SST anomalies may be limited to around 6 months. The problem lies in the fact that the observed midlatitude SSTs have some component forced from the atmosphere, and so by imposing observed SSTs in

a model, some information about the correct atmospheric state is already being prescribed. The mechanism acting in the Barsugli and Battisti (1998) model is reduced thermal damping, which acts as a positive feedback giving a positive correlation between the atmosphere and the local SSTs. Our analysis shows that the NAO is sensitive to remote SSTs, often in the tropics, and this mechanism does not apply to remote linkages. If we reconstruct the NAO using only SSTs southwards of 30°N , we obtain a reconstruction which has a correlation coefficient with the NOAA-20C NAO of 0.36, compared to 0.30 correlation when using SSTs north of 30°N . This suggests that there is a significant element of predictability arising from the tropical SSTs, which is not impacted by the lack of atmosphere-ocean coupling in the model. The coupling effect is also more important on longer timescales (Bretherton and Battisti, 2000), whereas the skill in our reconstructions comes from higher frequency, interannual SST forcing.

6.3.1.1 Winter

In this section we investigate the winter sensitivities, reconstructions and forcing mechanisms in more detail. To gain an understanding of how SSTs force the NAO, there are two components to understand. The first, demonstrated by the GTO, is where and to what degree is the NAO sensitive to SSTs. The second component is to understand where SSTs are varying in these regions of sensitivity. We decompose the global oceans into three regions: the Atlantic (90°W - 30°E), the Indian Ocean (30°E - 110°E), and the Pacific (110°E - 90°W). We then proceed to reconstruct the NAO using SSTs from each ocean in turn (shown in Fig. 6.7 for winter). In winter, the Atlantic reconstruction captures the largest portion of the variance, with the Pacific also contributing. Using this analysis, it is possible to see if the NAO in a particular year has some response that may be attributed to SST forcing, and also to look at which basin may have been important. For example, in 2010 the anomalously negative observed NAO appears to have been forced in some part by the Atlantic and Pacific SSTs. Interestingly, on the fast timescales, the Indian Ocean is not correlated with the reanalysis, even though the winter NAO has a

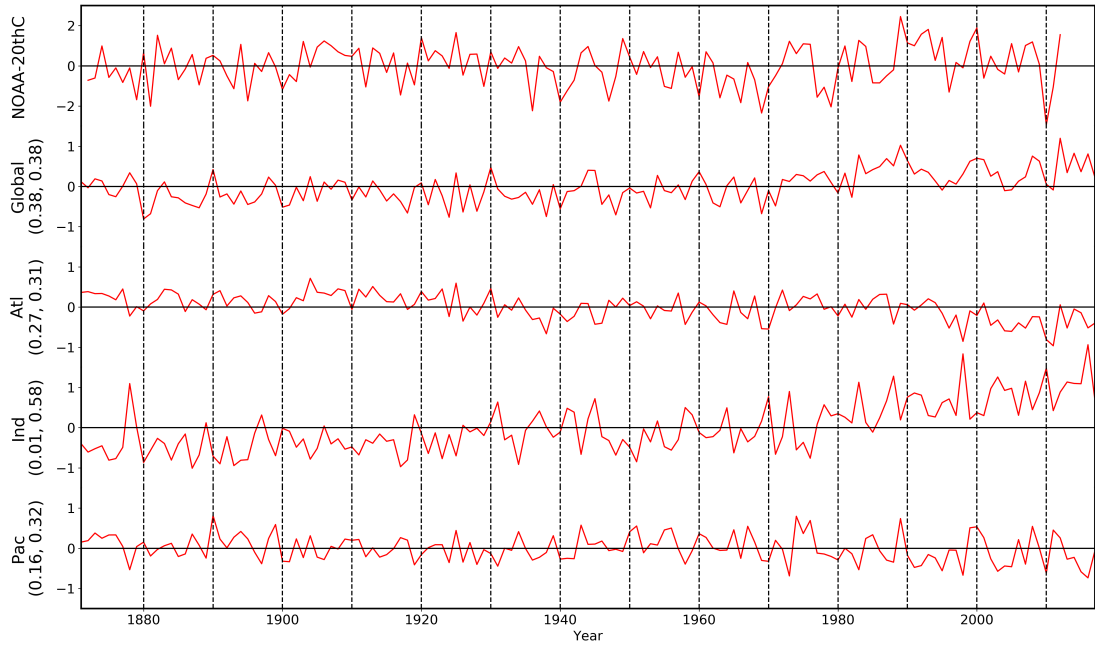


Figure 6.7: NOAA-20C winter NAO time series and reconstructed NAO time series using different ocean basin winter SSTs. Global uses all SSTs, Atlantic uses 90°W-30°E, Indian uses 30°E-110°E, Pacific uses 110°E-90°W. The year refers to the year in which January and February fall. The numbers in brackets indicate the correlation of the time series with the NOAA-20C NAO, and the standard deviation of the time series.

large region of sensitivity to SSTs in the Indian Ocean, and the high standard deviation in the Indian basin reconstruction suggests that there is variability in the Indian Ocean SSTs. However, the trend to more positive NAO years in the later part of the time period does appear to be linked to the Indian Ocean. However, we cannot confirm this as the reconstruction has no skill coming from the Indian Ocean.

The large sensitivity to the Indian Ocean in Fig. 6.4 suggests that the model NAO is influenced by the Indian Ocean on interannual timescales. However, the comparison of the reconstructed NAO from the Indian Ocean SSTs with observations (Fig. 6.7) does not show this teleconnection occurring on these timescales. One possible reason for this is that in reality, on these timescales, atmospheric circulation associated with the El Niño–Southern Oscillation (ENSO) can force the Indian Ocean SSTs, rather than the Indian Ocean SSTs forcing the atmosphere (Klein et al., 1999). However, Fletcher and Cassou (2015) do observe in coupled model simulations that the Indian Ocean can

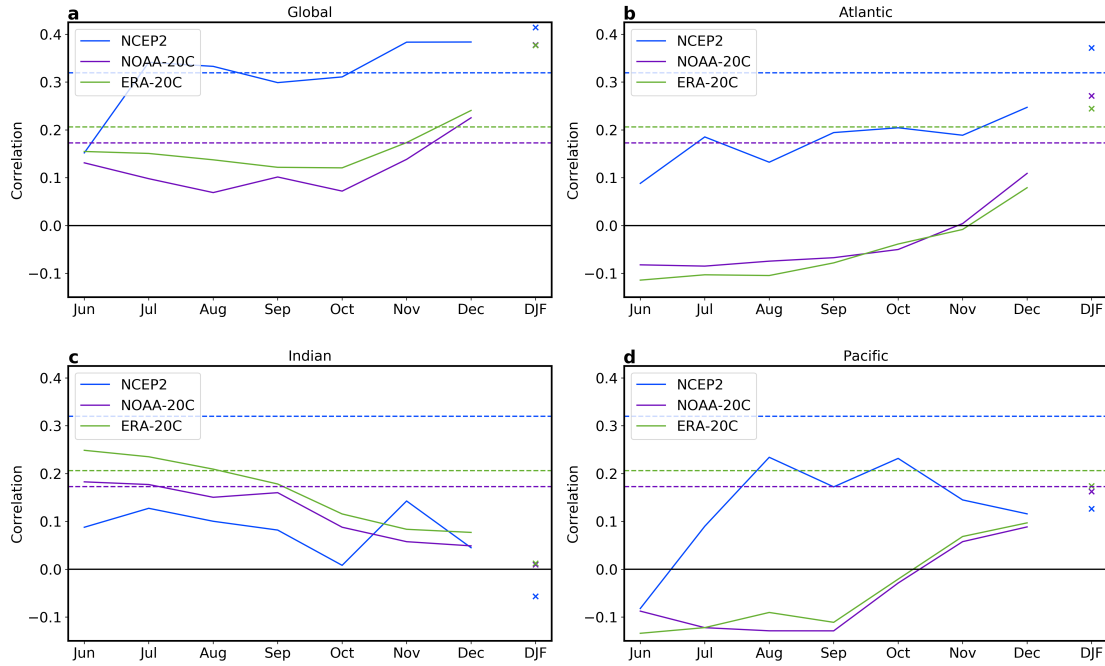


Figure 6.8: Reconstructed time series correlated with reanalysis time series using SSTs (a) globally, (b) from the Atlantic basin, 90°W - 30°E , (c) from the Indian basin, 30°E - 110°E , (d) from the Pacific basin, 110°E - 90°W . Along the abscissa are the individual months used to reconstruct the succeeding winter (DJF) NAO. The crosses mark the correlation of the reconstruction using the DJF SSTs with the reanalysis time series. The dashed lines mark the correlation required for the reconstruction and reanalysis to be correlated with a statistical significance at the 5% level, computed using a two-tailed t -test.

force the NAO, once isolated from ENSO variability. Further work is clearly needed to investigate the complex interactions between ENSO, tropical SSTs and teleconnections that force the NAO to confirm if these interactions are indeed the reason why we do not observe any skill in reconstructing the NAO from the Indian Ocean SSTs.

To investigate forecasting skill using preceding monthly SSTs, we reconstruct the winter NAO using SSTs in turn from individual months in the preceding year, up to the December in which the winter begins. We then correlate each reconstruction with the reanalysis time series. We do this for global SSTs and for each of the three basins. Fig. 6.8 shows the results of these lagged reconstructions for all the months and basins. Autocorrelation is accounted for in the significance test by computing the effective sample size as before, this time using the minimum effective sample size across all basins and

reconstruction month time series. As expected, the global reconstruction improves as the month used gets closer to the winter being reconstructed. We do not observe any skill arising from the previous winter SSTs (not shown) (cf. Dunstone et al., 2016). For the later period (NCEP2 reanalysis), the correlation is significant when using the November SSTs, indicating that there is some skill in forecasting the winter NAO. Decomposing this skill further into tropical versus midlatitude components gives a 0.40 correlation between the reconstruction using tropical November SSTs and the NCEP2 NAO versus a 0.01 correlation from using the midlatitude SSTs. This further validates the model, and the skill of preceding SSTs in forecasting the NAO shows that we are not just seeing a manifestation of false predictability arising from a lack of atmosphere-ocean coupling (cf. Bretherton and Battisti, 2000) as the skill is coming from remote tropical forcing and from SSTs in the month before the NAO is being reconstructed.

The skill also increases towards the winter when using SSTs from the Atlantic basin, suggesting a fairly direct effect of the SSTs on the circulation such as due to changing temperature gradients or modifying eddy behaviour. In contrast, the main skill in the Indian Ocean comes from SSTs earlier in the year, with some significant skill using June and July SSTs to forecast the winter NAO in ERA-20C and NOAA-20C. We note that these long lead timescales should be treated with caution; whilst it appears the Indian Ocean SSTs may contribute to the trend in the NAO and this decadal signal may be what is being picked up as significant, the Indian Ocean SSTs do not seem to add any forecast skill for forecasting on interannual timescales. It may be that the Indian Ocean SSTs are impacting SSTs in other regions which then influence the NAO, however an investigation of this effect is beyond the scope of this thesis. The Pacific SSTs in the later period seem to provide some skill in late summer/early autumn, but this is not statistically significant.

Despite the apparent lack of skill in the reconstruction and forecasting arising from the Indian Ocean, it is still insightful to consider how it forces the NAO in the model. The spatial separation and lag between the Indian Ocean SSTs forcing the winter NAO, and to a lesser extent the Pacific SSTs forcing the winter NAO, suggest that the forcing mech-

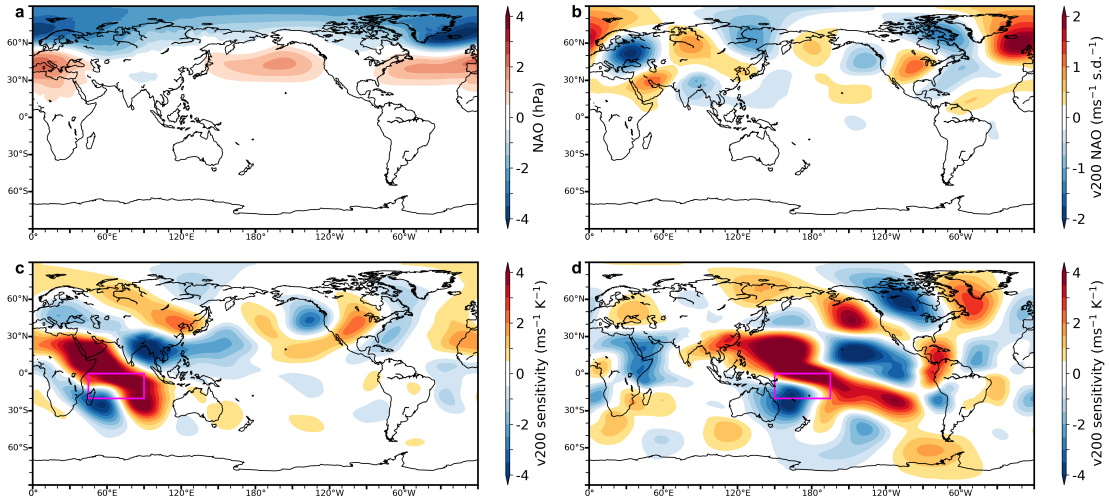


Figure 6.9: (a) Control ensemble member winter SLP regressed onto principal components associated with the 1st EOF of winter SLP in HadAM3P (defined over the region 20°N - 80°N , 90°W - 40°E). (b) 200 hPa meridional wind from the HadAM3P control ensemble members regressed onto the ensemble member principal components of the 1st EOF. Regression of the 200 hPa meridional wind responses onto the SST perturbations averaged over the pink box computed for (c) Indian Ocean SST box, (d) Pacific Ocean SST box.

anism may manifest via Rossby waves. To investigate this we look for patterns in the 200 hPa meridional wind response in winter to Indian and Pacific SSTs. Fig. 6.9a shows the control ensemble members SLP regressed onto the set of control ensemble member principal components associated with the 1st EOF of SLP in winter in HadAM3P (defined over the region 20°N - 80°N , 90°W - 40°E). Fig. 6.9b is the 200 hPa meridional wind at each gridpoint across the control ensemble members regressed onto the set of control ensemble member principal components of the model's 1st EOF. It shows that the winter NAO has a circumglobal wave train associated with it, which has a cyclonic region over Greenland, as expected for a positive NAO. This pattern is very similar to the pattern obtained by performing the same regression for the observed meridional wind and NAO (see Fig. 6.10). To investigate the circulation response to SSTs, we form an SST anomaly index given by the anomaly between a forced and control ensemble member pair, averaged over a particular region. We then regress the 200 hPa meridional wind anomalies onto the SSTs anomalies. Fig. 6.9c shows the meridional wind response to

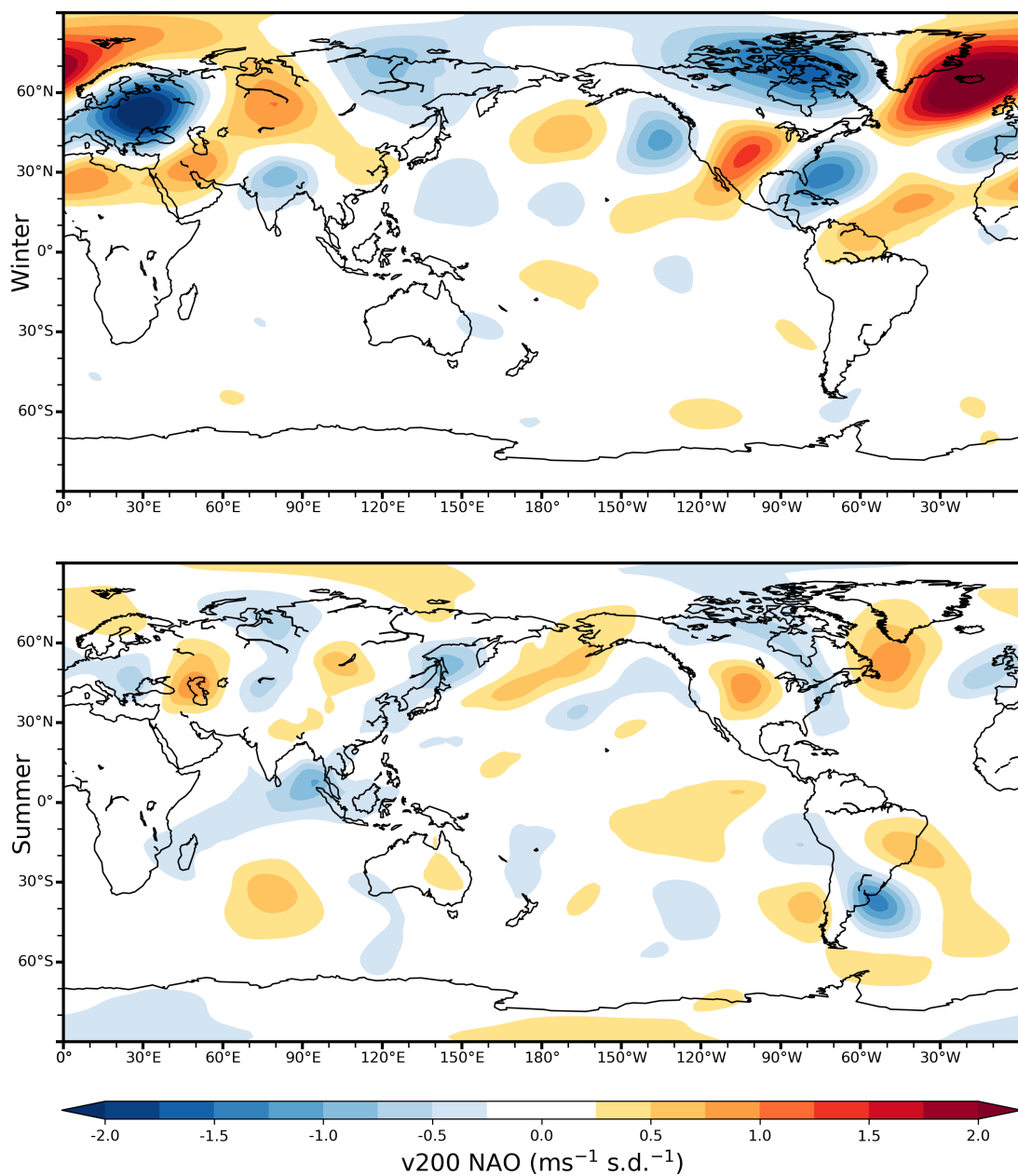


Figure 6.10: The seasonal mean NOAA-20C 200 hPa meridional wind regressed onto the NOAA-20C NAO in (a) DJF and (b) JJA.

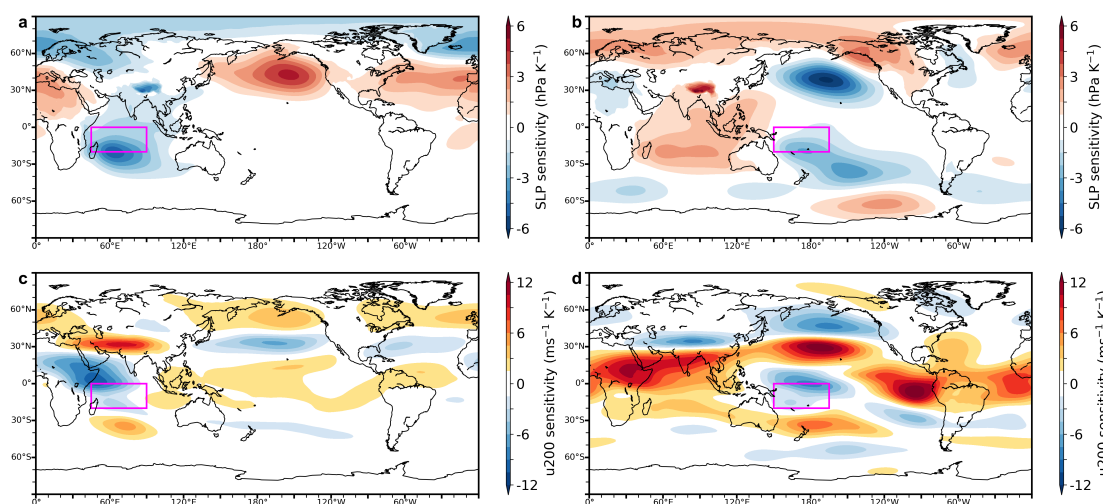


Figure 6.11: Regression of the SLP responses onto the SST perturbations averaged over the pink box computed for (a) Indian Ocean SST box, (b) Pacific Ocean SST box. Regression of the 200 hPa zonal wind responses onto the SST perturbations averaged over the pink box computed for (c) Indian Ocean SST box, (d) Pacific Ocean SST box.

SSTs over the region of high winter NAO sensitivity in the Indian Ocean (pink box). The local response exhibits the classical quadrupole structure (e.g. Jin and Hoskins, 1995), however, there is also a midlatitude wave response which propagates eastward into the Atlantic basin. The wavenumber and phase of this wave match the wave associated with the model EOF, particularly over the western Pacific, North America and into the Atlantic region. This provides evidence for a Rossby wavelike forcing of the winter NAO from the Indian Ocean. The Pacific response (Fig. 6.9d) also features a wavetrain propagating over North America and into the Atlantic, which has an anticyclonic region over Iceland, consistent with a negative NAO. The opposite sign of wave response over the North Atlantic and projection onto the EOF, at least in the Indian heating case, is consistent with Fletcher and Kushner (2011), who ascribe opposite cases of linear interference of a forced Rossby wave with the climatological stationary wave to Indian and Pacific SST forcing. This linear interference leads to a positive NAO for Indian Ocean forcing and a negative for Pacific Ocean forcing as observed in this chapter for Indian and Pacific forcing respectively.

We also calculate the SLP and 200 hPa zonal wind responses to Indian and Pacific

SSTs (Fig. 6.11). Consistent with previous studies (e.g. Hoskins and Karoly, 1981), the local response to heating is a negative pressure anomaly. The heating also projects remotely onto the Aleutian Low and the NAO via the Rossby wave as discussed. In the Indian Ocean case, the heating is associated with a pressure pattern over the North Atlantic that strongly resembles a positive NAO, whilst the Pacific pattern resembles the NAO to a lesser extent. The effect on the Aleutian low is to create a ridge in the case of Indian Ocean heating and a trough in the case of Pacific Ocean heating. This is in agreement with Drouard et al. (2013) who find a Pacific ridge can act as a precursor to a positive NAO, and a trough can act as a precursor to a negative NAO. The effect of the Pacific circulation state on the NAO is also demonstrated by Franzke et al. (2004), who find that an equatorward Pacific jet can lead to a positive NAO and a poleward jet can lead to a negative NAO. However, here we see the opposite: in the Indian heating case the Pacific jet shifts poleward, whilst in the Pacific heating case the Pacific jet shifts equatorwards. This suggests that the effect on the NAO from the tropical SSTs does not manifest via the breaking wave mechanism in Franzke et al. (2004).

6.3.1.2 Summer

We now return to investigate the summer NAO in greater detail, in particular the contributions of the different basins and the reason behind the time varying skill in reconstructing the summer NAO. As before, we divide the globe into three ocean basins, reconstructing the summer NAO using the concurrent summer SSTs from each basin in turn (shown in Fig. 6.12). The correlations and standard deviations are computed over the time period where the reconstruction is skilful (1925-1994). This time, it is clear that the majority of the skill in the reconstruction comes from the Pacific, where the skill is nearly the same as the global reconstruction, and the standard deviation is much higher than the other two basin reconstructions. The basin lagged predictors for summer (Fig. 6.13) show that directly preceding monthly SSTs are important in the Pacific, allowing the summer NAO to be forecast with some level of statistically significant skill

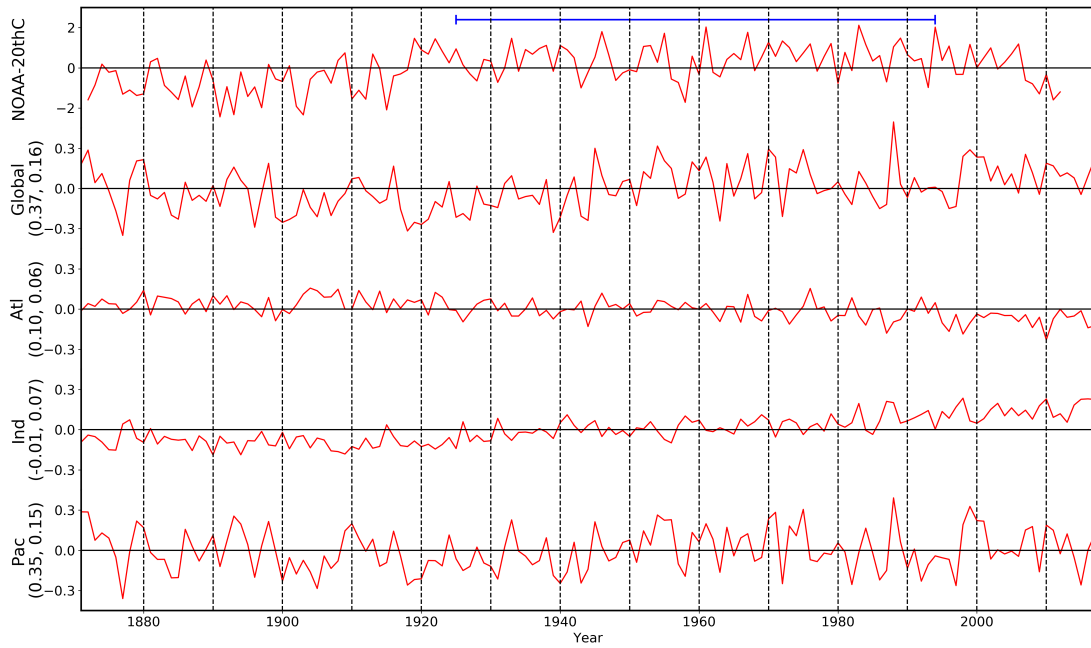


Figure 6.12: NOAA-20C summer NAO time series and reconstructed NAO time series using different ocean basin summer SSTs. Global uses all SSTs, Atlantic uses 90°W-30°E, Indian uses 30°E-110°E, Pacific uses 110°E-90°W. The numbers in brackets indicate the correlation of the time series with the NOAA-20C NAO, and the standard deviation of the time series, both computed over the period 1925-1994 (shown by the blue line).

(as opposed to winter where the Pacific SSTs slightly earlier in the year were important) using the global or just Pacific SSTs from April or May. Decomposing this skill further into tropical versus midlatitude components gives a 0.29 correlation between the reconstruction using tropical May SSTs and the ERA-20C NAO versus a 0.24 correlation from using the midlatitude SSTs. This further validates our model as it exhibits real predictability, which is not an artefact of imposed SSTs and the lack of atmosphere-ocean coupling effect discussed earlier, as a significant portion of the skill is coming from the tropical SSTs. As in winter, the Indian SSTs appear to add some skill, but we suggest this may be due to the trend in Indian Ocean SSTs being important in summer too.

Referring back to Fig. 6.4, we observe that there are three significant regions in the Pacific that appear to force the summer NAO. We isolate each region individually and reconstruct the summer NAO just using SSTs from that region. The most skilful

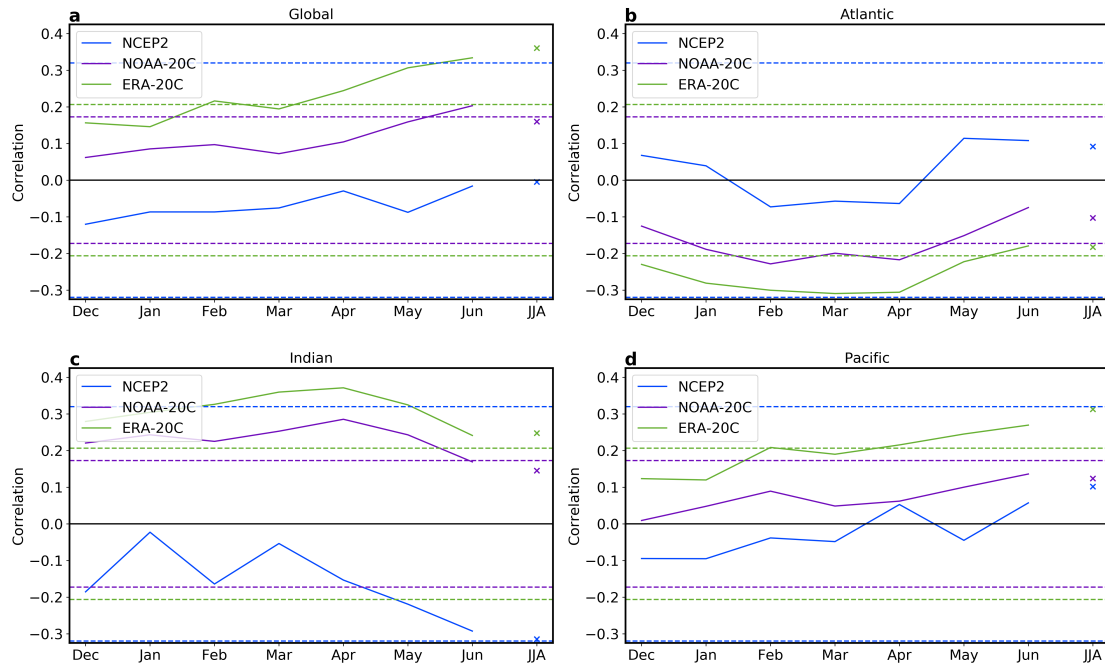


Figure 6.13: Reconstructed time series correlated with reanalysis time series using SSTs (a) globally, (b) from the Atlantic basin, 90°W - 30°E , (c) from the Indian basin, 30°E - 110°E , (d) from the Pacific basin, 110°E - 90°W . For each reanalysis, the correlation is computed using the whole period available from that reanalysis. Along the abscissa are the individual months used to reconstruct the succeeding summer (JJA) NAO. The crosses mark the correlation of the reconstruction using the JJA SSTs with the reanalysis time series over their full time periods. The dashed lines mark the correlation required for the reconstruction and reanalysis to be correlated with a statistical significance at the 5% level, computed using a two-tailed t -test.

reconstruction comes from the positive sensitivity Pacific region (30°N - 10°N , 100°E - 150°E) which has a correlation of 0.22 with the NOAA-20C summer NAO over 1925-1994. We now compute the summer SST standard deviation averaged over this region, and correlate them with the 31-yr moving correlation between the NOAA-20C summer NAO and reconstructed time series (using the global SSTs). Fig. 6.14 shows that the two quantities are very highly correlated (0.90), suggesting that the variability in this region is key for a skilful summer NAO reconstruction. If the variability is low, then there is not a large enough signal from the SSTs to skilfully reconstruct the summer NAO. An interesting facet to the summer NAO reconstruction is its relation to the changing spatial pattern over the time period (see Fig. 6.15 and Bladé et al., 2012). In periods where

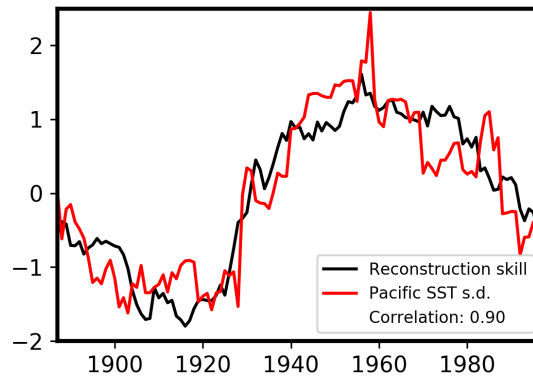


Figure 6.14: Time series of the reconstruction skill, which is the standardized 31-yr moving correlation between the NOAA-20C summer NAO and the reconstructed NAO, and the standardized time series of the 31-yr moving standard deviation of the detrended SSTs in the positive sensitivity Pacific region: 30°N-10°N, 100°E-150°E. The year refers to the middle year of the 31-yr moving window.

the reconstruction is skilful, the positive centre of the summer NAO in observations is located further north and east over the UK, whereas in the earlier period, and to a lesser extent in the later period, the centre is located over the Atlantic Ocean. Moving 51-yr EOFs from the observations (not shown) show that the positive centre of action moves smoothly east and north until the 1970s, when it begins to shift back towards the Atlantic. The latitudinal and longitudinal extents of the positive centre of action are both highly correlated (>0.80) with the Pacific SST standard deviation, and the summer NAO reconstruction skill. Inspection of the model EOF used to compute the GTO (Fig. 6.3a) shows that the EOF is similar to the NOAA-20C and ERA-20C EOFs, and not the NCEP2 EOF, even though the reconstruction performs better in periods when the EOF pattern looks like the NCEP2 EOF. This precludes the possibility that the reconstruction skill depends on the EOF used to compute the GTO. As an aside, we tested the statistical significance of the changing summer NAO pattern by using a Monte Carlo method to construct EOFs from randomly drawn samples of 70 years from NOAA-20C. The pattern correlation coefficient between the whole time series EOF and the second half of the time series EOF fall in the bottom 0.2% of 1000 randomly constructed EOFs, demonstrating that the EOF pattern is indeed statistically significantly different in the later time period.

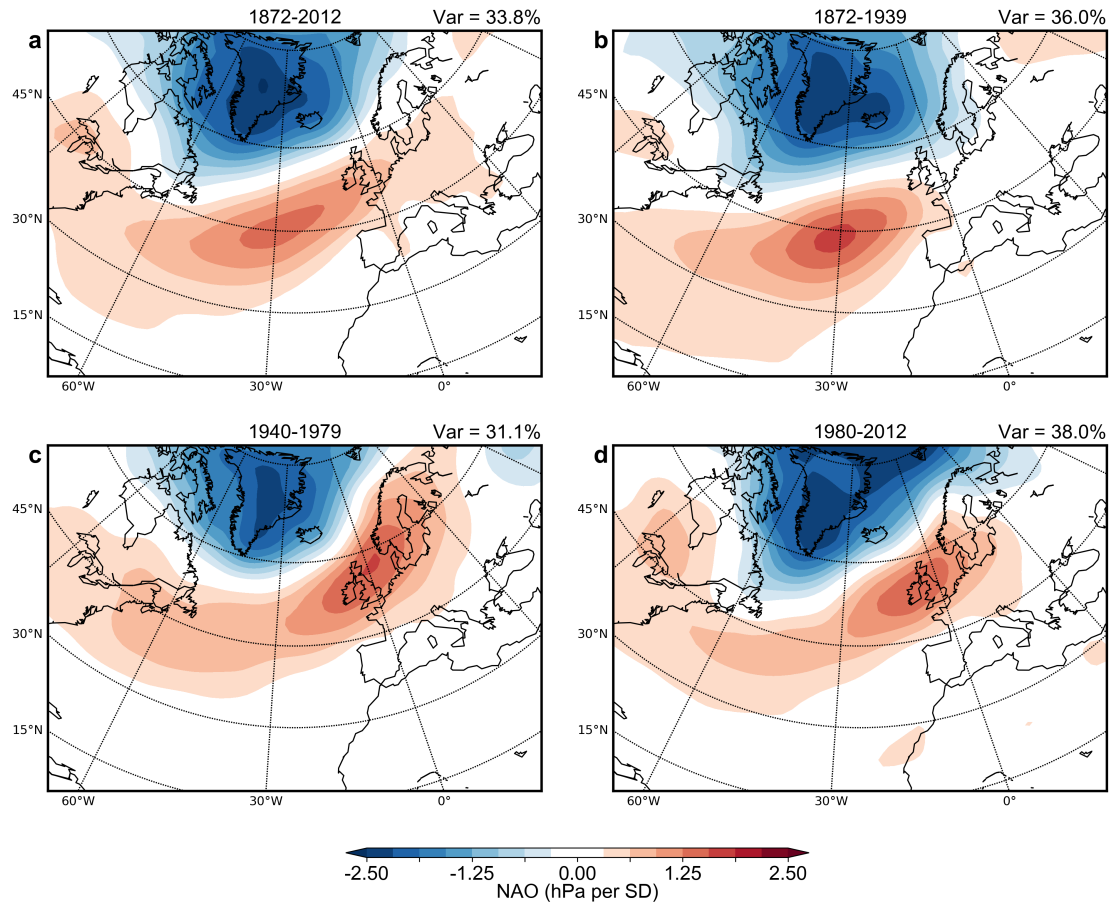


Figure 6.15: SLP regressed onto the 1st principal component time series of JJA SLP over 20°N-80°N, 90°W-40°E from NOAA-20C. (a) 1872-2012, (b) 1872-1939, (c) 1940-1979, (d) 1980-2012. Numbers indicate the percentage of the variance the EOF explains.

This changing nature of the summer NAO could be evidence that the summer NAO, like the winter NAO, is in fact a continuum of patterns (Johnson et al., 2008).

To investigate the mechanism by which the positive sensitivity Pacific region forces the summer NAO, we regress the circulation onto the SSTs in the region. Fig. 6.16a shows the control ensemble members SLP regressed onto the set of control ensemble member principal components associated with the 1st EOF of SLP in summer in HadAM3P (defined over the region 20°N-80°N, 90°W-40°E). Fig. 6.16b is the 200 hPa meridional wind at each gridpoint across the control ensemble members regressed onto the set of control ensemble member principal components of the model EOF. As in winter, there is a circumglobal wave pattern associated with the NAO, with a cyclonic region

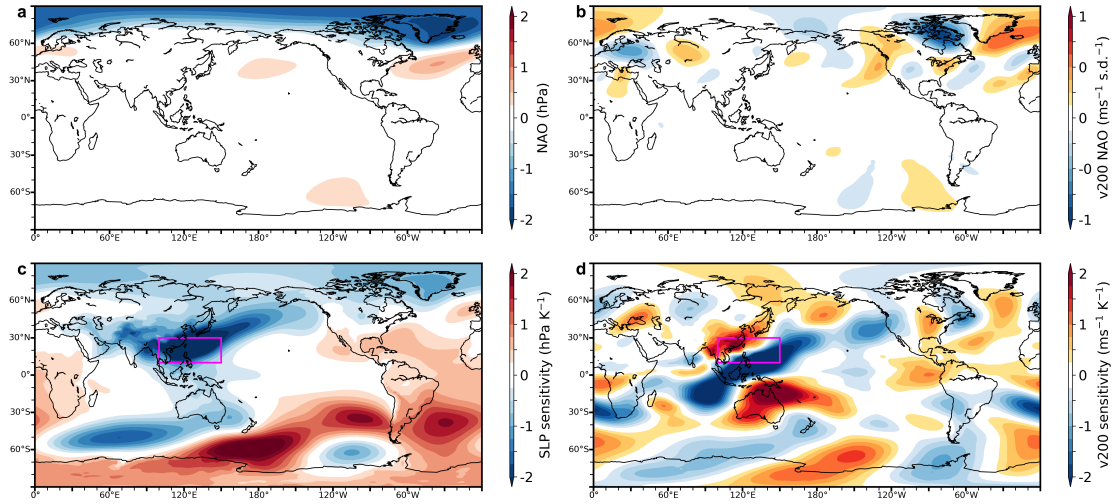


Figure 6.16: (a) Control ensemble member summer SLP regressed onto principal components associated with the 1st EOF of summer SLP in HadAM3P (defined over the region 20°N-80°N, 90°W-40°E). (b) 200 hPa meridional wind from the HadAM3P control ensemble members regressed onto the ensemble member principal components of the 1st EOF. Regression of (c) SLP and (d) 200 hPa meridional wind responses onto the SST perturbations averaged over the pink box.

over Greenland, as expected for a positive NAO. However, the model pattern does not match the equivalent pattern computed from regressing the observed meridional wind onto the observed NAO (Fig. 6.10). Fig. 6.16c shows the SLP response to SSTs over the Pacific region. Interestingly, the pressure response over the North Atlantic looks like the summer NAO in the 1940-1979 period in the reanalysis (Fig. 6.15c), rather than the model summer NAO. The 200 hPa meridional wind response (Fig. 6.16d) does not appear to project onto the model meridional wind associated with the 1st EOF, but instead contains a wave with an anticyclonic centre just west of the British Isles, which is associated with the 1940-1979 summer NAO pattern. This provides further explanation as to why the link between the SST variability in this Pacific region and the summer NAO reconstruction skill in the 1940-1979 period is high. Also, the meridional wind response in Fig. 6.16d has a wave over North America and the North Atlantic which does resemble the wave pattern obtained from the regression of the observed meridional wind onto the observed NAO (Fig. 6.10), suggesting that this mechanism does occur in the real world. Regressions of the circulation onto other high sensitivity regions in the

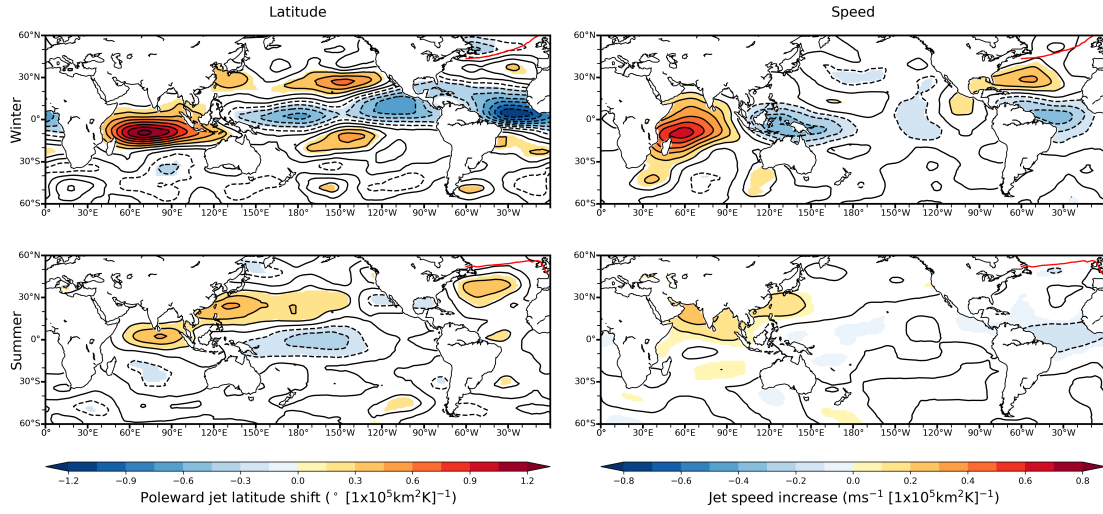


Figure 6.17: Sensitivity of the North Atlantic jet indices to SST anomalies given by the $G(x')$ in Eq. 6.1. Shaded regions indicate statistical significance at the 5% level using a two-tailed t -test. The red line indicates the position of the climatological North Atlantic storm track (taken as the maximum of the meridional component of the eddy heat flux).

Indian and Pacific basins (not shown) also show the NAO response in the model arising through a cyclone/anticyclone in the same region (i.e. not projecting onto the meridional wind associated with the model EOF). This suggests that the SST forcings in the model are not proficient at forcing the model's own EOF, but can force a summer NAO response that resembles that in the observed 1940-1979 period, explaining the skill in this period.

6.3.2 The eddy-driven jet

We now focus on the North Atlantic eddy-driven jet, and compare results to the NAO findings. The GTOs for jet latitude and speed are shown in Fig. 6.17. The sensitivities bear resemblance to those for the NAO, and the winter jet latitude plot exhibits more widespread areas of significant sensitivity than the winter NAO. All four GTOs indicate a large degree of sensitivity to Indian Ocean SSTs. In winter, the North Atlantic tripole pattern is also clear for the jet latitude, but the jet speed only appears sensitive to the two southerly parts of the tripole. The sensitivities for the jet latitude in the Atlantic and Pacific agree with the sensitivities found in Chapter 3 for diabatic heating in a dry idealized

Table 6.1: Table of reconstructed jet latitude and speed skill as given by the correlation between the reconstructed time series and the reanalysis time series. Coefficients in bold are statistically significant at the 5% level using a two-tailed t -test.

		NCEP2	NOAA-20C	ERA-20C
<i>Winter</i>	Latitude	0.26	0.34	0.33
	Speed	0.23	0.32	0.34
<i>Summer</i>	Latitude	0.06	0.18	0.14
	Speed	0.22	0.18	0.24

model. Heating on the equator leads to an equatorward jet shift, whilst heating in the subtropics up to the southern flank of the jet shifts the jet poleward. This is reminiscent of the result in Lu et al. (2008): the typical El Niño response due to narrow SST increases confined to the equator is an equatorward jet shift, as opposed to the anticipated poleward shift under global warming due to broader heating extending out of the tropics. Poleward of the jet, heating acts to shift the jet equatorward. In agreement with Chapter 3, there is also a cross-hemispheric sensitivity observed in the Pacific and Atlantic, with heating around 20°S forcing a poleward shift of the eddy-driven jet. The mechanism for this shift, discussed in Chapter 4, could be mediated via Hadley cell changes. A full investigation of the mechanisms in HadAM3P is not undertaken for this thesis. The mechanism by which the Indian Ocean forces the North Atlantic circulation is via the propagation of Rossby waves (as discussed earlier), and therefore the sensitivities in this region differ to the other two ocean basins. The speed sensitivities, however, do not agree with Chapter 3, where we found that tropical heating up to the latitude of the storm track strengthens the jet, whereas here, SST increases in the tropics act to weaken the jet.

We reconstruct time series of the jet latitude and speed indices and compare them to the reanalysis time series. The correlation coefficients between the reconstructions and the different reanalyses are shown in Table 6.1. In winter, the speed and latitude

reconstructions are significantly correlated with the NOAA-20C and ERA-20C time series, however the correlation drops in the later period, indicating some element of non-stationarity in the teleconnection between the SSTs and the jet speed and latitude that is not present in the NAO. As before, the GTO reconstruction performs better than the set of hindcast experiments used in Weisheimer et al. (2017) for both winter jet speed and latitude (Parker et al., 2019). For summer, the correlation coefficients are lower, in keeping with the smaller sensitivities of the two indices in the GTOs compared to winter. The highest skill in summer again occurs during the 1925-1994 period for both the speed and latitude (not shown), with the latitude reconstruction skill dropping off in the later period, but the speed reconstruction skill not.

The skill in reconstructing the NAO over the longer reanalysis periods is greater than for the individual jet indices, suggesting that the skill in the NAO may be coming from both jet latitude and speed. Following Parker et al. (2019), we perform a multiple linear regression to determine the relationship between the NAO and the jet latitude and speed in the reconstruction, with the aim of attributing the predictable NAO signal to jet latitude, speed, or both. In winter, the regression coefficients for the reconstruction are 0.60 for jet speed and 0.56 for jet latitude, combining to explain 67% of the NAO variability over the reconstruction time period. In summer, the regression coefficients for the reconstruction are 0.28 for jet speed and 0.78 for jet latitude, combining to explain 69% of the NAO variability. The coefficients for NOAA-20C are 0.32 and 0.81 for winter jet speed and latitude respectively (explaining 76% of the variability) and 0.34 and 0.69 for summer jet speed and latitude (explaining 59% of the variability). Whilst the summer coefficients are fairly similar between the reconstruction and the reanalysis, it seems like the winter NAO in the model is over dependent on the jet speed. This could be due to the winter jet being too zonal and too strong in the model (see mean jet biases in Fig. 6.1). The jet also has a higher ratio of jet speed standard deviation to the jet latitude standard deviation (0.46 in the model versus 0.37 in NOAA-20C) suggesting that the model has a greater component of pulsing variability than shifting variability compared

Table 6.2: Table of changing NAO skill for various terms in the NAO multilinear regression. The *Slow* and *Fast* columns show the correlation between the NOAA-20C and reconstructed time series when that element of the reconstructed NAO calculation is omitted. The correlation without omitting any elements is 0.43 in winter and 0.21 in summer.

		<i>Slow</i>	<i>Fast</i>
<i>Winter</i>	Latitude	0.39	0.28
	Speed	0.46	0.42
<i>Summer</i>	Latitude	0.25	0.15
	Speed	0.20	0.16

to the reanalysis, consistent with the jet being further poleward in the model further poleward in the model climatology than in NOAA-20C (48.2° versus 47.2° in winter). This is consistent with Barnes and Hartmann (2011) who observe that the leading mode of zonal-wind variability transitions from a shift to a pulse as the eddy-driven jet moves poleward. In summer, there is no difference between the standard deviations ratios, even though the model jet is further poleward (50.6° versus 48.8°).

To ascertain the time scales on which the jet features are providing skill to the NAO, we perform a multiple linear regression following Parker et al. (2019). We use the jet latitude and speed on both fast and slow time scales as predictors for the NAO. We then remove each feature in turn to determine their contribution to the NAO skill. The slow components are computed using an 11-yr low pass Lanczos filter with 11 weights (Duchon, 1979), and the fast components using an 11-yr high pass Lanczos filter. Table 6.2 shows the correlations obtained between the NOAA-20C time series and regression constructed time series as each component is individually removed. In winter, it is only the fast component of the latitude that appreciably reduces the skill (the same is true for 31-yr filter windows, not shown). This agrees with Woollings et al. (2014) in the sense that on fast timescales the NAO is dominated by variations in jet latitude, however, the relationship they find on longer timescales between the jet speed and NAO is not cap-

tured in this method. In summer, both the fast timescale speed and latitude appear to be important, and the removal of the slow timescale latitude actually slightly increases the reconstruction skill. From this simple linear approach it appears that it is the faster timescales that give predictability to the NAO, suggesting again that the decadal Indian Ocean influence may be less important than previous studies have suggested.

6.4 Conclusions

In this chapter, we have used a linear statistical-dynamic method applied to a large ensemble of randomly perturbed SSTs to investigate the North Atlantic circulation sensitivity to SSTs. We first showed the regions where the winter and summer NAO is sensitive to changes in SSTs, identifying several key regions. The Atlantic tripole pattern is present in both winter and summer sensitivity maps, in particular the tropical Atlantic node of the tripole. Over the Indian Ocean there is a large SST monopole in winter and dipole in summer that the NAO is sensitive to, whilst in the West Pacific there is a dipole in both seasons. Using the GTO, we demonstrated skill in reconstructing both the winter and summer NAO, although there is some non-stationarity in the skill in reconstructing the summer NAO that appears to arise from the Pacific SST variability. In winter, the reconstruction performs better than a recent ensemble of atmosphere-only hindcast experiments and the reconstruction skill does not display the same drop off in skill in the 1950-70s as in the hindcasts (Weisheimer et al., 2017).

When using concurrent SSTs to reconstruct the NAO, the Atlantic SSTs and to a lesser extent the Pacific SSTs, are important in providing the skill in the reconstruction. We demonstrated that this linear reconstruction method can be applied using SSTs before the winter season, allowing a skilful winter NAO forecast to be made. In summer, the Pacific SSTs are the most important in providing skill, and we demonstrated using this method that skilful summer NAO forecasts can be made, which to this date have not been reported elsewhere (Dunstone et al., 2018).

We also computed the North Atlantic eddy-driven jet latitude and speed sensitivities to SSTs, which display broadly similar patterns to the NAO, although some differences do exist between all three sensitivity maps. We demonstrated similar levels of skill in reconstructing time series of the jet latitude and speed, showing that in winter, the speed and latitude contribute roughly equally to the variability in the NAO reconstruction, whereas in summer the latitude contributes most of the variability in the reconstruction. Investigating the importance of fast and slow varying timescales in the jet latitude and speed, it seems that the fast timescale jet latitude is important for reconstructing the winter NAO, whilst both fast jet latitude and speed are important for reconstructing the summer NAO.

The sensitivities we observe for the NAO and jet are in broad agreement with previous literature, identifying the Indian Ocean region (as seen in e.g. Hoerling et al., 2001, 2004; Bader and Latif, 2003, 2005; Fletcher and Kushner, 2011) and Atlantic tripole (as seen in e.g. Rodwell et al., 1999; Czaja and Frankignoul, 2002; Peng et al., 2002; Wang et al., 2004) as important sensitivity regions. In agreement with Fletcher and Kushner (2011), the Indian Ocean sensitivity appears to arise from, in winter at least, the forcing of Rossby waves which constructively interfere with the climatological stationary wave, thus forcing a positive NAO response. To a lesser extent, the Pacific Ocean SSTs produce a Rossby wave which destructively interferes with the climatological wave, thus forcing a negative NAO. The jet latitude sensitivities in the Pacific and Atlantic basins agree with idealized sensitivity studies (Lu et al., 2008, and Chapter 3 of this thesis), and contain the equatorward jet shift seen for El Niño-like forcing and the poleward shift for broader tropical heating as in the global warming case.

Here we have only considered the impact of tropical and midlatitude SSTs on the NAO and jet, when there are numerous other potentially important drivers such as sea ice (e.g. Pedersen et al., 2016), snow (e.g. Henderson et al., 2018), the sun (e.g. Gray et al., 2016), the stratosphere (e.g. Scaife et al., 2005) and atmosphere-ocean coupling effects (e.g. Czaja et al., 2013). The fact therefore, that we see some predictive skill in both win-

ter and summer in reconstructing the NAO using a purely linear method demonstrates that SSTs are an important driver of North Atlantic atmospheric circulation. There may also be other non-linear SST effects that the RPM does not capture, and these effects are not quantified in this thesis. Given the importance of the NAO and jet on European, North American and North African weather and climate (e.g. Visbeck et al., 2001; Hurrell and Deser, 2009; Cattiaux et al., 2010; Bladé et al., 2012; Schaller et al., 2016), it is vitally important that we can forecast the NAO skilfully. The ability of the RPM to skilfully reconstruct the NAO from SSTs therefore provides a highly useful tool for attribution and forecasting, also providing direct attribution of where the signals are coming from in the system.

Chapter 7

Direct CO₂ radiative forcing versus SST warming

Having explored the sensitivity of the atmospheric circulation in a simple and a more complex GCM, we now seek to understand how different types of forcing can affect the atmospheric circulation differently. In this chapter, we will consider the effects of both radiative forcing and anomalous sea surface temperatures on the Northern Hemisphere circulation. We will examine how the sensitivity of the climate response to these two types of forcings is important in determining the realised response to climate change.

7.1 Introduction

It is well established that there are differing, and sometimes opposing, effects of radiative forcing versus sea surface temperature (SST) change on the atmospheric circulation (e.g. Mitchell, 1983; He and Soden, 2015; Shaw and Voigt, 2015; Chadwick et al., 2017). Land warming is primarily driven by the SST warming of the oceans, an indirect effect of anthropogenic emissions, rather than the direct response to increasing greenhouse gases (GHGs) over land (Compo and Sardeshmukh, 2009). However, global mean temperature increase driven by the SST warming is not the sole driver of changes in temperature and

precipitation patterns (Folland et al., 1998; Sexton et al., 2003; Bony et al., 2013; Dong and Sutton, 2015; Richardson et al., 2016), extremes (Kamae et al., 2014a; Seneviratne et al., 2014; Dong et al., 2016), or changes in midlatitude circulation (Ceppi et al., 2018).

Shaw and Voigt (2015) investigate the large scale circulation responses to direct CO₂ radiative forcing versus SST warming in the Coupled Model Intercomparison Project phase 5 (CMIP5) ensemble, finding that CO₂ forcing leads to positive land-sea equivalent potential temperature (θ_e) contrasts, and SST forcing leads to negative land-sea θ_e contrasts. In the CO₂ case, the positive land-sea θ_e contrasts lead to low-level convergence around Asia, a stronger Asian monsoon cyclone and Pacific anticyclone, and a poleward shift of the Pacific jet stream. The negative land-sea θ_e contrasts in the SST case lead to the opposite effects on the circulation. However, Shaw and Voigt (2015) do not look specifically at changes in midlatitude stationary waves in response to the two types of forcing, which may have an important impact on future changes in regional Northern Hemisphere (NH) summer temperatures and heatwave hazard due to the direct effect of CO₂ (investigated further in Chapter 8).

A key factor in the opposing circulation responses comes from the differing land-sea contrast response to radiative forcing versus SST warming (Kamae et al., 2014b; Shaw and Voigt, 2015). The land-sea contrast arises from the non-linearity of the Clausius-Clapeyron relationship between saturation specific humidity and temperature (Joshi et al., 2008). The dynamical precipitation response to direct CO₂ forcing has been associated with the land-sea contrast, leading to an increase in precipitation over tropical land, and a decrease over the oceans due to a weakening of the overturning circulation (Bony et al., 2013). Upper tropospheric heating stabilises the atmosphere which leads to a decrease in convection (He and Soden, 2015). However, in response to the heating the land warms and destabilises the atmosphere, leading to an increase in convection over land (Bony et al., 2013).

Changes in convection can cause changes in the upper-level divergence, which can force Rossby waves that can propagate from the tropics into the extratropics (Sardesh-

mukh and Hoskins, 1988). Specifically, changes in the Walker circulation have been shown to affect upper tropospheric divergence and produce a stationary wave with wavenumber 5 over the NH midlatitudes in winter (Haarsma and Selten, 2012) which strongly resembles the circumglobal wave guide pattern described by Branstator (2002). The NH summer circulation also has a wave 5 circumglobal wavetrain associated with natural variability (Ding and Wang, 2005), which has a forced component arising from the tropics (Ding and Wang, 2007; Yasui and Watanabe, 2010).

Schubert et al. (2011) suggest that stationary Rossby waves can play an important role in causing NH summer heat extremes and Screen and Simmonds (2014) show the link between extreme weather event frequency in the NH and amplified stationary planetary waves. Studies using reanalysis (Cassou et al., 2005) and model data (Lee et al., 2017) point towards tropical forcing in exciting Rossby waves that cause extremes. One particular theory of how summer stationary waves cause heat extremes is quasiresonant amplification (Petoukhov et al., 2013). However, in this chapter we test the hypothesis that simple changes in tropical forcing can trigger stationary Rossby waves that can affect heat extremes. Previous studies identify the tropical Atlantic (Cassou et al., 2005), tropical Pacific and Caribbean (Wulff et al., 2017; O'Reilly et al., 2018), Indian monsoon region (Joseph and Srinivasan, 1999; Lin, 2009) and equatorial Africa (Yasui and Watanabe, 2010) as important Rossby wave forcing regions. We aim to diagnose the important forcing locations for the stationary waves seen in response to the SST and CO₂ forcings.

Here we investigate the responses of precipitation and NH stationary waves to these two types of forcings. Using three general circulation models (GCMs) and the CMIP5 ensemble, we develop and test a theory on the midlatitude circulation response to SST warming and direct CO₂ forcing. We hypothesize that the two different forcings produce opposite changes in divergence (and corresponding opposite changes in precipitation) in the tropics, leading to stationary waves in antiphase. These opposite divergence responses arise from the opposite responses of land-sea θ_e contrasts and the differing ther-

modynamic effects of radiative forcing versus SST increases. We then test our hypothesis that the anomalous forcings produce the stationary wave using a simple barotropic model to reproduce the GCM wave response and identify key forcing locations.

Section 7.2 outlines the methods and models used in the chapter. Section 7.3 presents the GCM results leading us to perform the barotropic model experiments which we analyse in Section 7.4. We then investigate the forcing mechanisms in Section 7.5. Section 7.6 discusses the relevance of the work and the implications of our findings, before we present our conclusions in Section 7.7.

7.2 Methods

7.2.1 GCM data

In this chapter, we use results from large ensembles of three general circulation models (GCMs): HadAM3P, MIROC5 (Watanabe et al., 2010) and CAM4 (Gent et al., 2011). These are run at $1.25^\circ \times 1.875^\circ$, $1.4^\circ \times 1.4^\circ$ and $1.9^\circ \times 2.5^\circ$ resolutions respectively. The models are run as Atmospheric Model Intercomparison Project (AMIP) style ensembles (Taylor et al., 2012) with prescribed SSTs and sea ice. All the ensembles and their setups are summarised in Table 7.1.

7.2.1.1 1.5°C experiments

The first set of experiments, the 1.5°C experiments, have two ensembles for each model, which use the Half a degree Additional warming, Prognosis and Projected Impacts (HAPPI; www.happimip.org) design (Mitchell et al., 2017a). These two ensembles use SSTs and other boundary conditions to simulate conditions in a world 1.5°C warmer than pre-industrial (1861-1880) conditions. Here we use the HAPPI Tier 1 experimental design (Mitchell et al., 2017a), which uses the multi-model mean SST patterns; thus, all ensembles and models are run with an identical time varying 1.5°C SST pattern. The

Table 7.1: Summary of ensembles and their setups.

<i>Experiment</i>	<i>Ensemble</i>	<i>Model</i>	<i>Resolution</i>	<i>SST</i>	<i>CO₂[ppm]</i>
1.5°C	LCO ₂	HadAM3P	1.25°×1.875°	1.5°C	395.8
	HCO ₂	HadAM3P	1.25°×1.875°	1.5°C	550.0
	LCO ₂	MIROC5	1.4°×1.4°	1.5°C	395.8
	HCO ₂	MIROC5	1.4°×1.4°	1.5°C	550.0
	LCO ₂	CAM4	1.9°×2.5°	1.5°C	379.0
	HCO ₂	CAM4	1.9°×2.5°	1.5°C	423.1
Hadley	NAT	HadAM3P	1.25°×1.875°	Natural	284.4
	GHG	HadAM3P	1.25°×1.875°	Natural	390.4
	SST	HadAM3P	1.25°×1.875°	Historical	284.4
CMIP5	AMIP	see text	various	Historical	Historical
	AMIP4xCO ₂	see text	various	Historical	4×Historical
	AMIP4K	see text	various	Historical+4K	Historical

full details are discussed in Mitchell et al. (2017a).

The two ensembles differ only in CO₂ concentration, but otherwise use the RCP2.6 scenario forcings from the last decade of the 21st century. In HadAM3P and MIROC5, CO₂ is fixed at 395.8 ppm (hereafter the LCO₂ ensemble) and 550.0 ppm (HCO₂ ensemble), representing the lower and higher likely CO₂ concentrations averaged over 2091–2100 in adaptive pathways that succeed in achieving warming below 1.5°C in 2100 for the assessed ranges of climate response uncertainty (Millar et al., 2017). For CAM4, we use experiments with the CO₂ fixed at 379.0 ppm and 423.1 ppm; hence, differences between the CAM4 ensembles still occur due to the difference in CO₂ concentration, but should not be quantitatively compared to the other two models. All CO₂ concentrations are prescribed as a global mean atmospheric concentration and the CO₂ is well-mixed. All ensemble members are run for 11 years, with the first year discarded to allow the atmosphere to spin up. For HadAM3P there are 98 and 119 members for the LCO₂ and HCO₂ ensembles respectively. Both MIROC5 ensembles have 50 members, and both CAM4 ensembles have 501 members. Justifications of the ensembles setups and further details will be given in Chapter 8 where it is more relevant to do so. The main point for this chapter is that these simulations provide large ensembles from three models in which the CO₂ concentration is varied but the SSTs are held fixed.

7.2.1.2 Hadley experiments

We run a further three ensembles just using HadAM3P, the Hadley experiments. These ensembles, named NAT, GHG and SST, are designed to isolate the response of the atmosphere to SST forcings and GHG forcings associated with global warming separately. The NAT ensemble has natural SSTs and GHGs which simulates a world with no anthropogenic climate change. The NAT SSTs are calculated by removing the modelled anthropogenic warming from observed SSTs. The anthropogenic warming is calculated as in Haustein et al. (2016): 11 GCM simulations from the CMIP5 archive are averaged and the monthly climatologies (1996-2005) of the *HistoricalNat* simulations are subtracted from the *Historical* simulations. The mean of these 11 patterns is used as the SST pattern for the NAT ensemble and pre-industrial atmospheric GHG concentrations (IPCC 1900 forcing) are used to force the NAT ensemble. For further details, see Haustein et al. (2016) and Schaller et al. (2016). The GHG ensemble uses the NAT SSTs, with RCP4.5 GHGs (from 2005-2015). The SST ensemble uses the observed 2005-2015 SSTs from the OSTIA SST dataset (Donlon et al., 2012) and pre-industrial GHGs. The CO₂ concentration of the NAT and SST ensembles is 284.4 ppm, the average CO₂ concentration of the GHG ensemble is 390.4 ppm. Sea ice for the NAT and GHG ensembles is taken from the maximum well-observed sea ice (DJF 1986/1987), for the SST ensemble the 2005-2015 observed OSTIA sea ice is used. Volcanic forcings are set to zero in all three ensembles, and the solar forcing is taken from 2005-2015. All three ensembles are run for 11 years, with the first year discarded. There are 124, 118 and 113 ensemble members for the NAT, GHG and SST ensembles respectively.

7.2.1.3 CMIP5 experiments

We also make use of AMIP data from the AMIP control, AMIP4K and AMIP4xCO₂ simulations (Taylor et al., 2012) from the following CMIP5 models: bcc-csm1-1, CanAM4, CNRM-CM5, FGOALS-g2, HadGEM2-A, IPSL-CM5A-LR, IPSL-CM5B-LR, MPI-

ESM-LR and MPI-ESM-MR. All simulations are forced with historical SSTs and greenhouse gases and are run over the 30-year period 1979-2008.

7.2.2 Barotropic model

In this chapter we employ the idealised barotropic model from O'Reilly et al. (2018). Following Hoskins and Ambrizzi (1993), the model integrates the barotropic vorticity equation on a sphere:

$$\left(\frac{\partial}{\partial t} + \mathbf{u}_\Psi \cdot \nabla \right) \zeta = F - \lambda \zeta - \mu \nabla^4 \zeta, \quad (7.1)$$

where $\mathbf{u}_\Psi$ is the rotational velocity field, ζ is the absolute vorticity (the sum of planetary and relative vorticity components), F is a constant forcing term, λ is a linear damping rate with a timescale of 10 days, and $\mu = 2.4 \times 10^{16} \text{ m}^4 \text{ s}^{-1}$ is a biharmonic diffusion coefficient. The equation is solved using spectral harmonics with a triangular truncation at wavenumber 42 (i.e. T42). The forcing is decomposed into two terms, $F = \bar{F} + F'$, where $\bar{F}$ is selected to maintain the basic state exactly, which is taken to be the climatological vorticity field at 200 hPa from the relevant dataset.

The model output from HadAM3P, used as the background state in the barotropic model, includes winds on 850, 500 and 200 hPa pressure surfaces. We choose the 200 hPa winds to calculate the climatological vorticity field for the barotropic model; the anomalous tropical forcing (via divergent outflow) is strongest in the upper troposphere (e.g. Krishnamurti et al., 2013); and vorticity gradients, upon which the signal propagates out of the tropics, are also significantly stronger in the upper troposphere. It is necessary that the basic state includes these strong wave guides to achieve a reasonable response, both in amplitude and pattern (O'Reilly et al., 2018).

The anomalous forcing, F' , is selected to mimic the forcing of the barotropic vorticity equation by divergent outflow in the tropics. Following Sardeshmukh and Hoskins

(1988), the linearised Rossby wave source (RWS) anomaly S' is given by

$$S' = -\mathbf{u}'_{\chi} \cdot \nabla \bar{\zeta} - \bar{\mathbf{u}}_{\chi} \cdot \nabla \zeta' - \bar{\zeta} D' - \zeta' \bar{D}, \quad (7.2)$$

where $\mathbf{u}_{\chi}$ is the divergent velocity field and $D = \nabla \cdot \mathbf{u}_{\chi}$ is the divergence. Here the overbars and the primes denote climatological and anomaly fields respectively. In this chapter, we force the model with a RWS anomaly due only to anomalous divergence and anomalous transports of mean vorticity. We ignore terms due to anomalous vorticity, as these terms largely correspond to the circulation response we are assessing. The forcing term therefore reduces to:

$$S' = -\mathbf{u}'_{\chi} \cdot \nabla \bar{\zeta} - \bar{\zeta} D' = F'. \quad (7.3)$$

The model is initialised from the basic state and integrated forward for 25 days. The response becomes quasi-stationary after about two weeks, so the stationary response to forcing is presented as the average over the period 16-20 days.

7.2.3 Equivalent potential temperature

To analyse differences in the response to SST versus CO_2 forcing, we compute the equivalent potential temperature, θ_e , the temperature a parcel of air would reach if brought adiabatically to a standard reference pressure (taken to be 1000 hPa) and all the water vapour in the parcel were to condense, releasing its latent heat (Ambaum, 2010). Following Bolton (1980), θ_e can be calculated as follows:

$$\theta_e = \theta_L \exp \left[\left(\frac{3036}{T_L} - 1.78 \right) r (1 + 0.448r) \right]. \quad (7.4)$$

θ_L is the dry potential temperature (in Kelvin) at the lifted condensation level,

$$\theta_L = T \left(\frac{p_0}{p - e} \right)^{\kappa_d} \left(\frac{T}{T_L} \right)^{0.28r}. \quad (7.5)$$

T_L is the approximate temperature (in Kelvin) at the lifted condensation level,

$$T_L = \frac{1}{\frac{1}{T_d - 56} + \frac{\ln(T/T_d)}{800}} + 56. \quad (7.6)$$

Here, T is the temperature (in Kelvin) of the air at pressure p (in hPa), p_0 is a standard reference pressure (taken to be 1000 hPa) and $\kappa_d = R_d/c_{pd} = 0.2854$ is the ratio of the specific gas constant to the specific heat of dry air at constant pressure. e is the water vapour pressure in hPa, calculated from relative humidity by $RH = \frac{e}{e_s} \times 100\%$, where the saturated water pressure $e_s(T) = 6.1094 \exp \left[\frac{17.625(T-273.15)}{T-30.11} \right]$ (Alduchov and Eskridge, 1996). $r = 0.622 \frac{e}{p-e}$ is the mixing ratio of water vapour per mass (kg kg^{-1}). T_d is the dew point temperature which can be calculated from the relative humidity $RH = e_s(T_d)/e_s(T)$ (Ambaum, 2010).

7.3 GCM results

The June-July-August (JJA) difference in precipitation (colours) and 200 hPa meridional wind (contours) between ensemble mean HCO₂ and LCO₂ ensembles (CO₂ response) are shown in Fig. 7.1. The pattern of tropical precipitation change in the CO₂ response for the three models is very similar: a wetting over land and a drying over the oceans. This is associated with a poleward shift of the intertropical convergence zone (ITCZ) over the Sahel region of Western Africa.

We also observe a circumglobal wave pattern over the NH between 30°N-70°N. Whilst the strength of the wave varies spatially in the different models, a wavenumber 5 stationary wave with wave centres in the same locations is observed in all three CO₂ responses. Due to large ensemble sizes and long runs, the responses of the tropical

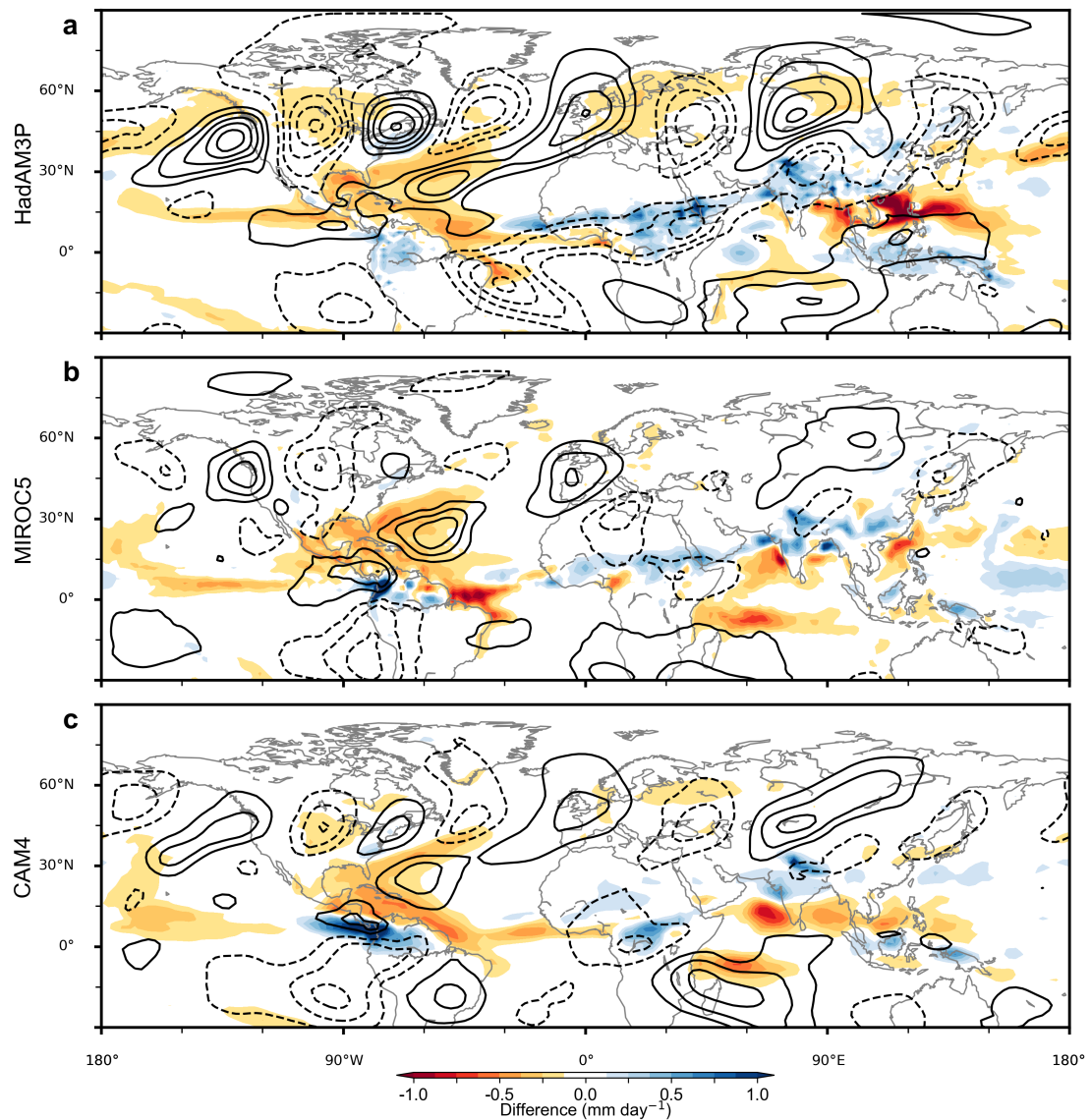


Figure 7.1: Responses of the different models to CO_2 forcing. (a) HadAM3P CO_2 , (b) MIROC5 CO_2 and (c) CAM4 CO_2 . Contours show the JJA 200 hPa meridional wind response (interval: 0.2 ms^{-1} with the zero contour suppressed), colours show the JJA precipitation response. The CAM4 wind and precipitation responses have been multiplied subjectively by 2.5 to be plotted on the same scale as the other two model responses.

precipitation and meridional wind are both statistically significant (shown for the quantities plotted in Fig. 7.1 in Fig. 7.2 and Fig. 7.3). To quantify the agreement between the model responses, we compute the anomaly correlation coefficients between the responses (shown in Table 7.2 and Table 7.3). As the correlation coefficients show, the model responses bear strong quantitative similarities to one another.

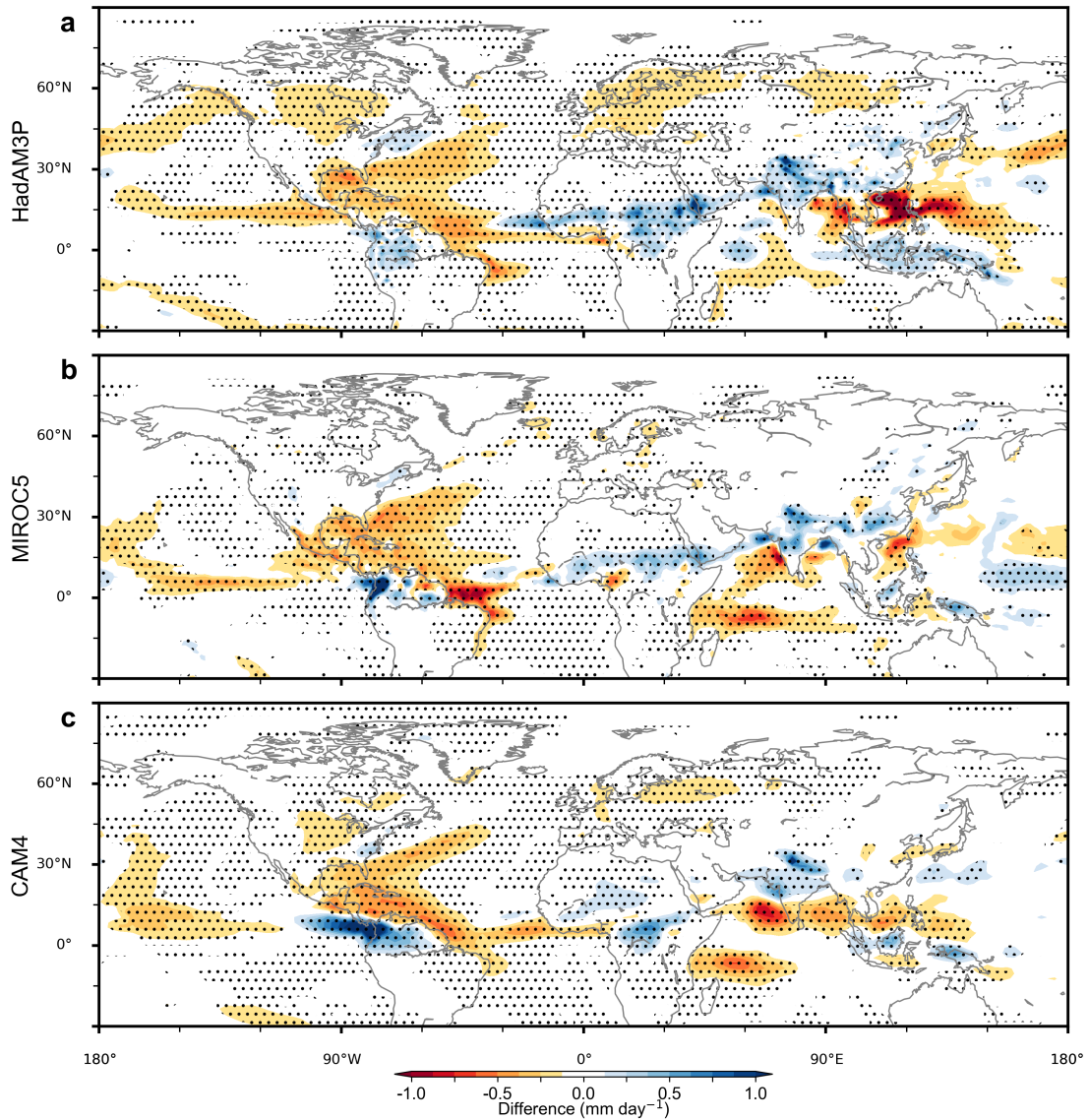


Figure 7.2: Significance of responses of the different models to CO₂ forcing. (a) HadAM3P CO₂, (b) MIROC5 CO₂ and (c) CAM4 CO₂. Colours show the JJA precipitation response. The CAM4 response has been multiplied subjectively by 2.5 to be plotted on the same scale as the other two model responses. Stippling shows significance between the means of the two sets of 10-year means for each ensemble member which is calculated using a two-tailed *t*-test at the 5% level.

Very similar stationary wave and precipitation responses to CO₂ forcing are evident and robust in the difference between the multi-model-mean of the AMIP4xCO₂ and AMIP control ensembles (AMIP4xCO₂ response) (Fig. 7.4a, individual model responses in Fig. 7.5). The responses to a uniform SST increase are shown in Fig. 7.4b (individual model responses in Fig. 7.6), the difference between the multi-model mean

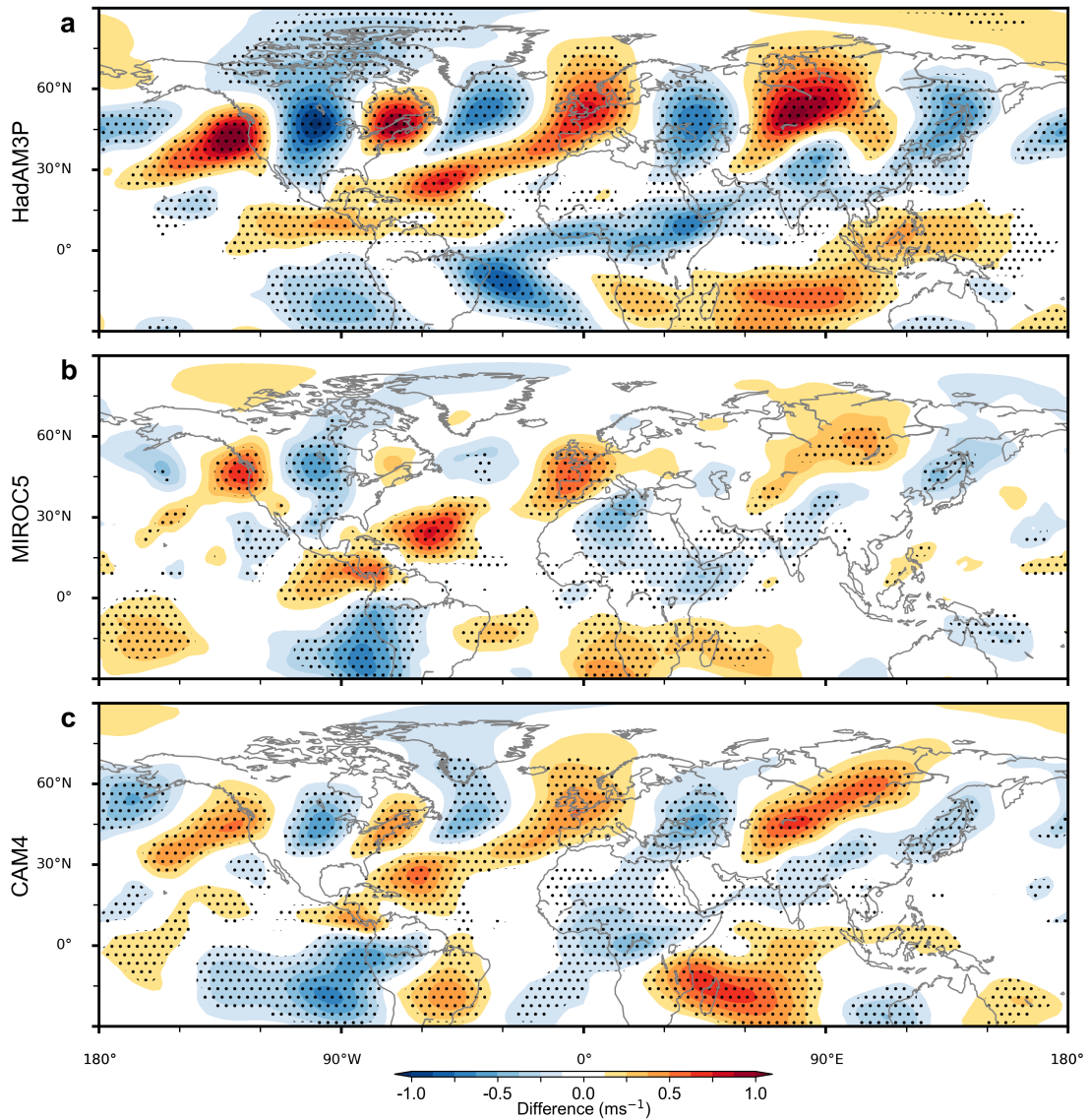


Figure 7.3: Significance of responses of the different models to CO_2 forcing. (a) HadAM3P CO_2 , (b) MIROC5 CO_2 and (c) CAM4 CO_2 . Colours show the JJA 200 hPa meridional wind response. The CAM4 response has been multiplied subjectively by 2.5 to be plotted on the same scale as the other two model responses. Stippling shows significance between the means of the two sets of 10-year means for each ensemble member which is calculated using a two-tailed t -test at the 5% level.

AMIP4K and AMIP control ensembles (AMIP4K response). Here, there is an opposite response in the precipitation over the tropical Atlantic and over the tropical region of Africa compared to the response to CO_2 . The stationary wave is still present, but with the wave centres in antiphase to the CO_2 and AMIP4x CO_2 responses.

A similar picture emerges when we consider the difference between the GHG and

Table 7.2: Anomaly correlation coefficients between 200 hPa meridional wind responses denoted by their figure panel, computed over the latitude band 30°N-70°N. Patterns in Fig. 7.4b, Fig. 7.4d and Fig. 7.9d have been multiplied by -1 before computing the correlations.

	Fig. 7.1a	Fig. 7.1b	Fig. 7.1c	Fig. 7.4c	Fig. 7.4d	Fig. 7.4a	Fig. 7.4b	Fig. 7.9c	Fig. 7.9d	Fig. 7.10f	Fig. 7.10g
Fig. 7.1a	1	0.70	0.80	0.85	0.75	0.84	0.45	0.34	0.24	0.39	0.06
Fig. 7.1b	-	1	0.72	0.70	0.58	0.71	0.49	0.23	0.26	0.32	0.08
Fig. 7.1c	-	-	1	0.77	0.69	0.73	0.40	0.30	0.18	0.37	0.09
Fig. 7.4c	-	-	-	1	0.68	0.66	0.36	0.31	0.19	0.35	-0.05
Fig. 7.4d	-	-	-	-	1	0.77	0.67	0.18	0.14	0.21	0.07
Fig. 7.4a	-	-	-	-	-	1	0.60	0.22	0.20	0.28	0.18
Fig. 7.4b	-	-	-	-	-	-	1	0.11	0.27	0.19	0.24
Fig. 7.9c	-	-	-	-	-	-	-	1	0.67	0.93	0.33
Fig. 7.9d	-	-	-	-	-	-	-	-	1	0.75	0.54
Fig. 7.10f	-	-	-	-	-	-	-	-	-	1	0.46
Fig. 7.10g	-	-	-	-	-	-	-	-	-	-	1

Table 7.3: Anomaly correlation coefficients between precipitation responses denoted by their figure panel, computed over the latitude band 15°N-5°S and across longitudes 60°W-60°E. Patterns in Fig. 7.4b and Fig. 7.4d have been multiplied by -1 before computing the correlations.

	Fig. 7.1a	Fig. 7.1b	Fig. 7.1c	Fig. 7.4c	Fig. 7.4d	Fig. 7.4a	Fig. 7.4b
Fig. 7.1a	1	0.45	0.51	0.88	0.57	0.74	0.67
Fig. 7.1b	-	1	0.47	0.52	0.37	0.53	0.37
Fig. 7.1c	-	-	1	0.54	0.42	0.54	0.51
Fig. 7.4c	-	-	-	1	0.62	0.72	0.66
Fig. 7.4d	-	-	-	-	1	0.49	0.57
Fig. 7.4a	-	-	-	-	-	1	0.83
Fig. 7.4b	-	-	-	-	-	-	1

NAT ensembles (GHG response), and SST and NAT ensembles (SST response) (Figs. 7.4c,d). The GHG response shows increased precipitation over the tropical region of Africa, and decreased precipitation over the tropical Atlantic. The stationary wave is present, with the same wave centres as the CO₂ and AMIP4xCO₂ responses. In agreement with the AMIP4K response, the SST response is almost diametrically opposed to the CO₂ and AMIP4xCO₂ responses: drying over the tropical regions of Africa and wetting over the tropical Atlantic. The stationary wave is also apparent in the SST response with the same wave centre locations as the AMIP4K response.

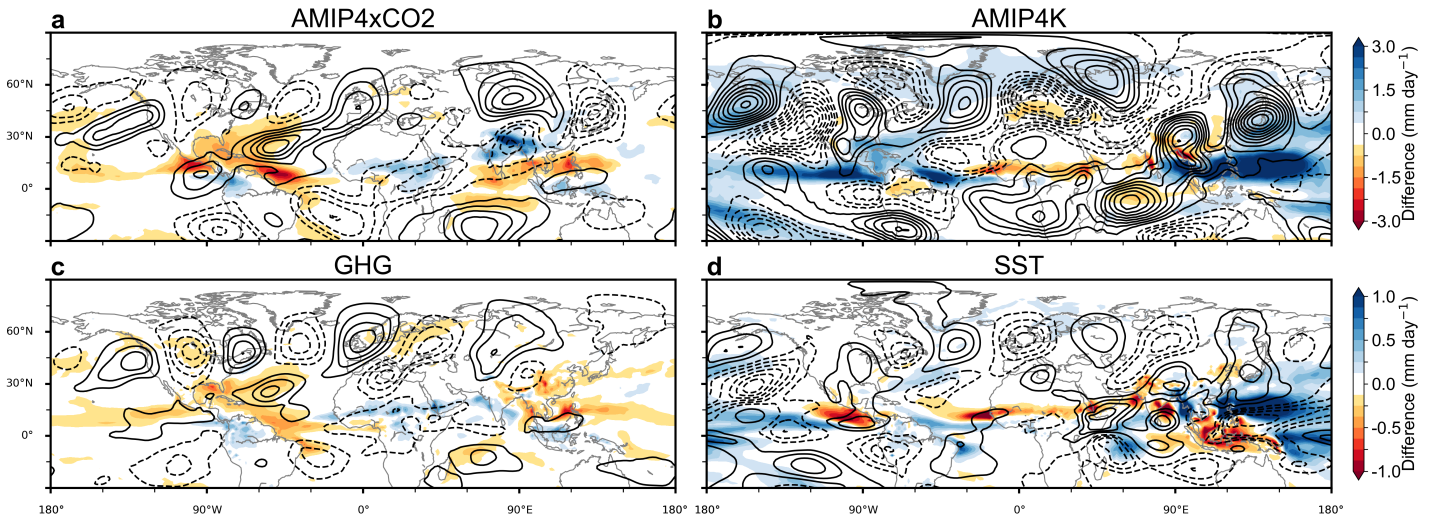


Figure 7.4: Responses to GHG and SST forcing. (a) AMIP4xCO₂, (b) AMIP4K, (c) HadAM3P GHG and (d) HadAM3P SST. Contours show the JJA 200 hPa meridional wind response (interval: 0.4 ms⁻¹ (a,b) and 0.2 ms⁻¹ (c,d) with the zero contour suppressed), colours show the JJA precipitation response.

To summarise the precipitation and meridional wind changes, we plot the differences between the AMIP4xCO₂ and AMIP4K responses and the differences between the GHG and SST responses in Fig. 7.7. This emphasises the important features consistent between the responses to direct CO₂ forcing and SST warming, and removes elements of noise. It is clear that the consistent precipitation changes between the CMIP5 experiments and Hadley experiments are located in the tropics over the Atlantic, Africa and Central-West Pacific. The wave 5 pattern and wave centres in the two experiments are also co-located, with slight differences over Europe. The similarity of the wave and precipitation responses across the range of models (quantified in Table 7.2 and Table 7.3) illustrates their robustness. Further to this, given the wave and precipitation responses to SST warming arise not only when applying the warming pattern (in the Hadley experiments), but also in the case of uniform warming (AMIP4K response), this suggests that the response is not just due to the spatial location of the SST warming pattern, but is a robust response to broad SST warming. The comparisons also show a similar response arising from both strong forcing (in the case of the CMIP5 experiments) and

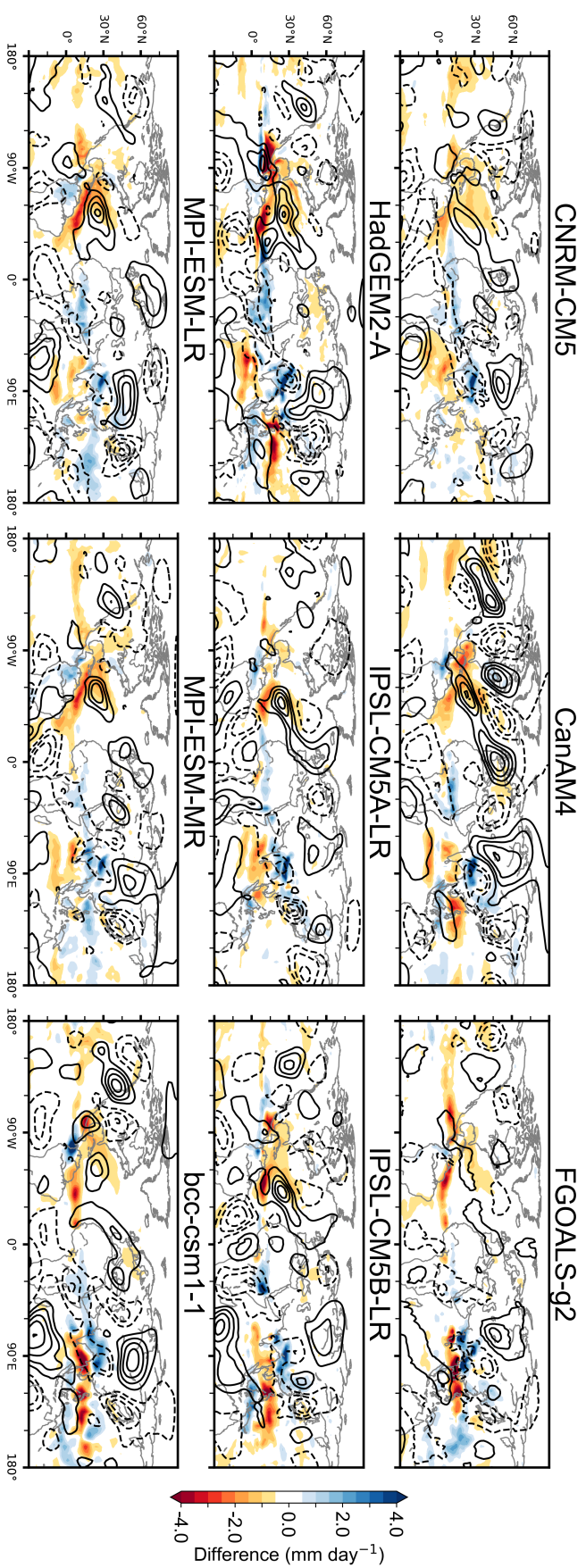


Figure 7.5: Individual AMIP responses to a 4xCO₂ increase. Contours show the JJA 200 hPa meridional wind response (interval: 0.6 ms⁻¹ with the zero contour suppressed), colors show the JJA precipitation response.

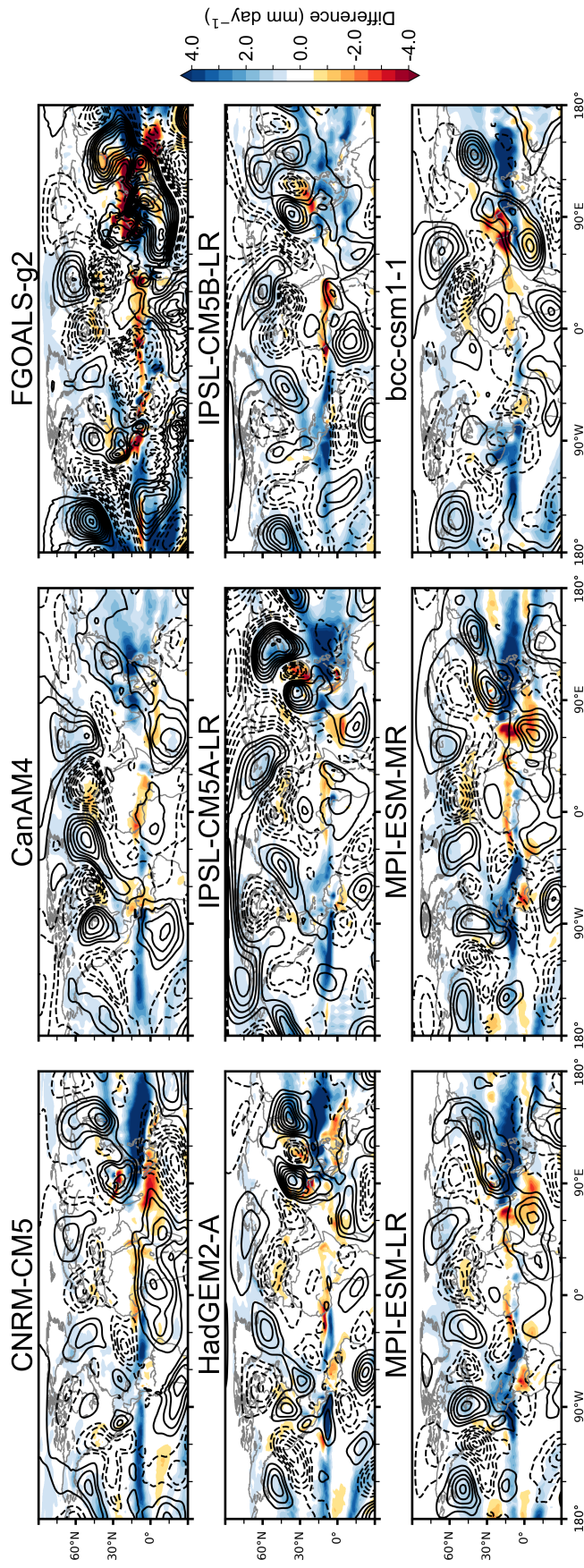


Figure 7.6: Individual AMIP responses to a uniform 4 K SST increase. Contours show the JJA 200 hPa meridional wind response (interval: 0.8 ms^{-1} with the zero contour suppressed), colors show the JJA precipitation response.

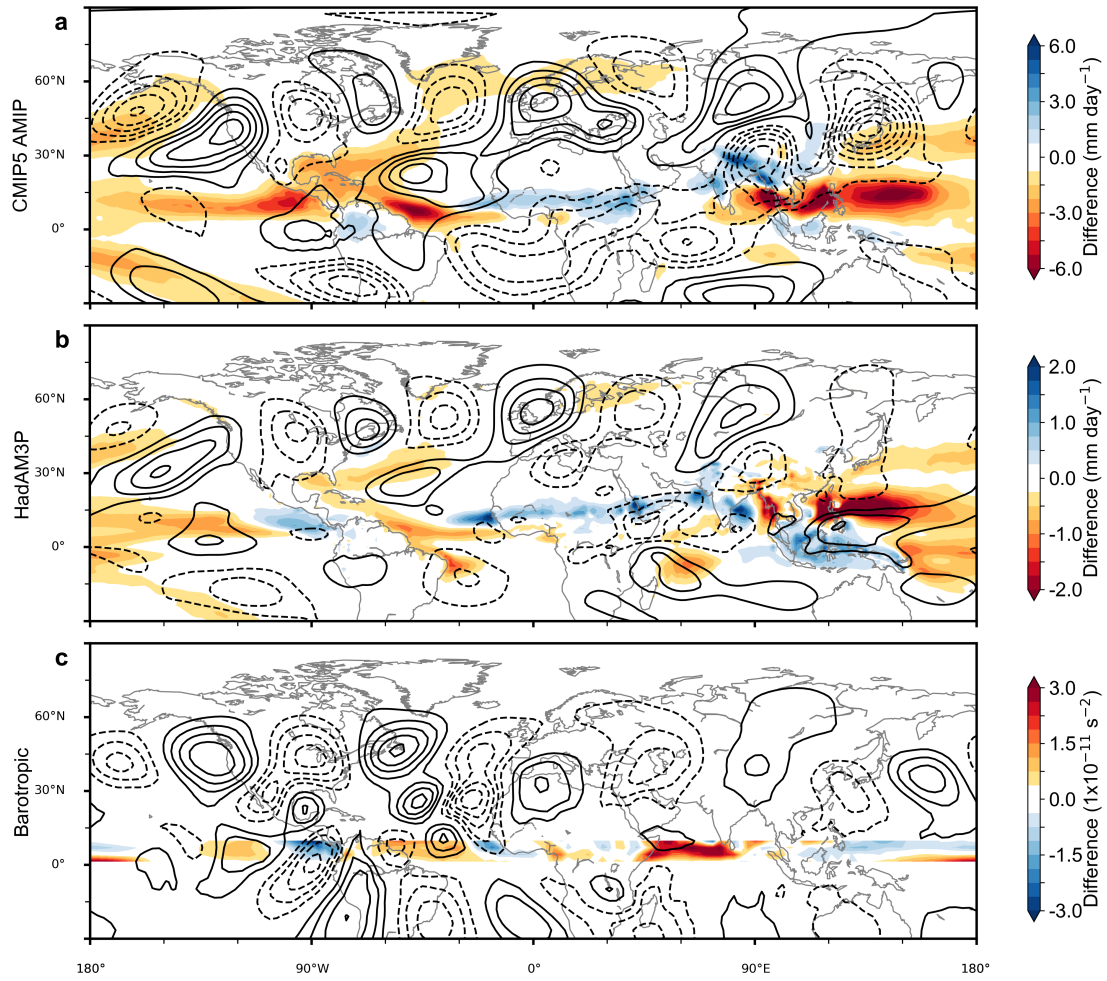


Figure 7.7: Summary of the responses to forcing in the different experiments. (a) Summary of the CMIP5 AMIP experiments showing difference the between $4\times\text{CO}_2$ and AMIP4K ensemble mean responses, (b) summary of the Hadley experiments showing the difference of GHG and SST ensemble mean responses. Contours show the JJA 200 hPa meridional wind response (interval: 0.8 ms^{-1} (a) and 0.4 ms^{-1} (b) with the zero contour suppressed), colours show the JJA precipitation response. (c) Summary of the barotropic modelling experiments. Colours show the difference between the anomalous May RWS forcings in the GHG and SST ensembles. Contours show the difference between the model response when forced with the anomalous GHG RWS and when forced with the anomalous SST RWS (interval: 0.3 ms^{-1} with the zero contour suppressed).

weak forcing (in the case of the Hadley experiments).

Before examining the origin of the stationary wave, we investigate what impact such a wave has on NH temperature. The ensemble mean JJA temperature response in the CO_2 forced ensembles is a wavelike positive-negative temperature anomaly pattern across the NH in JJA, which may be associated with increases in extreme temperature events

(explored further in Chapter 8). We focus on the 1.5°C and Hadley experiment results as we have large ensembles for these models, as opposed to the one realization from each model in the CMIP5 AMIP ensemble. We use a simple linear regression to diagnose the relationship between the temperature anomalies and the stationary wave.

We first define the response of a particular member as the difference between the HCO₂ ensemble member 10-year mean and the LCO₂ ensemble mean across the ensemble members. For each member we create a circulation index by first identifying the wave centres of the wave as the locations of the 200 hPa meridional wind maxima and minima in the wavetrain from the JJA ensemble mean 200 hPa meridional wind response fields (see dots in Fig. 7.8 for wave centre locations). We then sum the positive wave centres (pink dots) and subtract the negative wave centres (green dots) to give a circulation index for each ensemble member response. We then regress the standardized circulation indices for each ensemble member against the ensemble member mean temperature response at each gridpoint.

The regression coefficients are plotted in Fig. 7.8. They show that members with a strong circulation index response have strong temperature anomalies compared to members with weaker circulation index responses. Regions where the stationary wave is anticyclonic are associated with regions of increased temperature, and cyclonic regions are associated with regions of decreased temperature. The wave centres are in quadrature with the temperature anomalies as they are meridional wind wave centres and hence positive temperature anomalies have a positive meridional wind anomaly on the west and negative meridional wind anomaly on the east, indicating the presence of an anticyclone (and vice versa for negative temperature anomalies). Anticyclonic regions are typically characterised by clear skies, which leads to greater incident solar radiation on the Earth's surface, hence an increase in surface temperature (Bieli et al., 2015). The temperature anomalies are completely opposite for the SST case, which supports the link between the wave and the temperature anomalies further. In the CO₂ cases, the regions of positive temperature anomaly may be co-located with the NH regions that see increases in

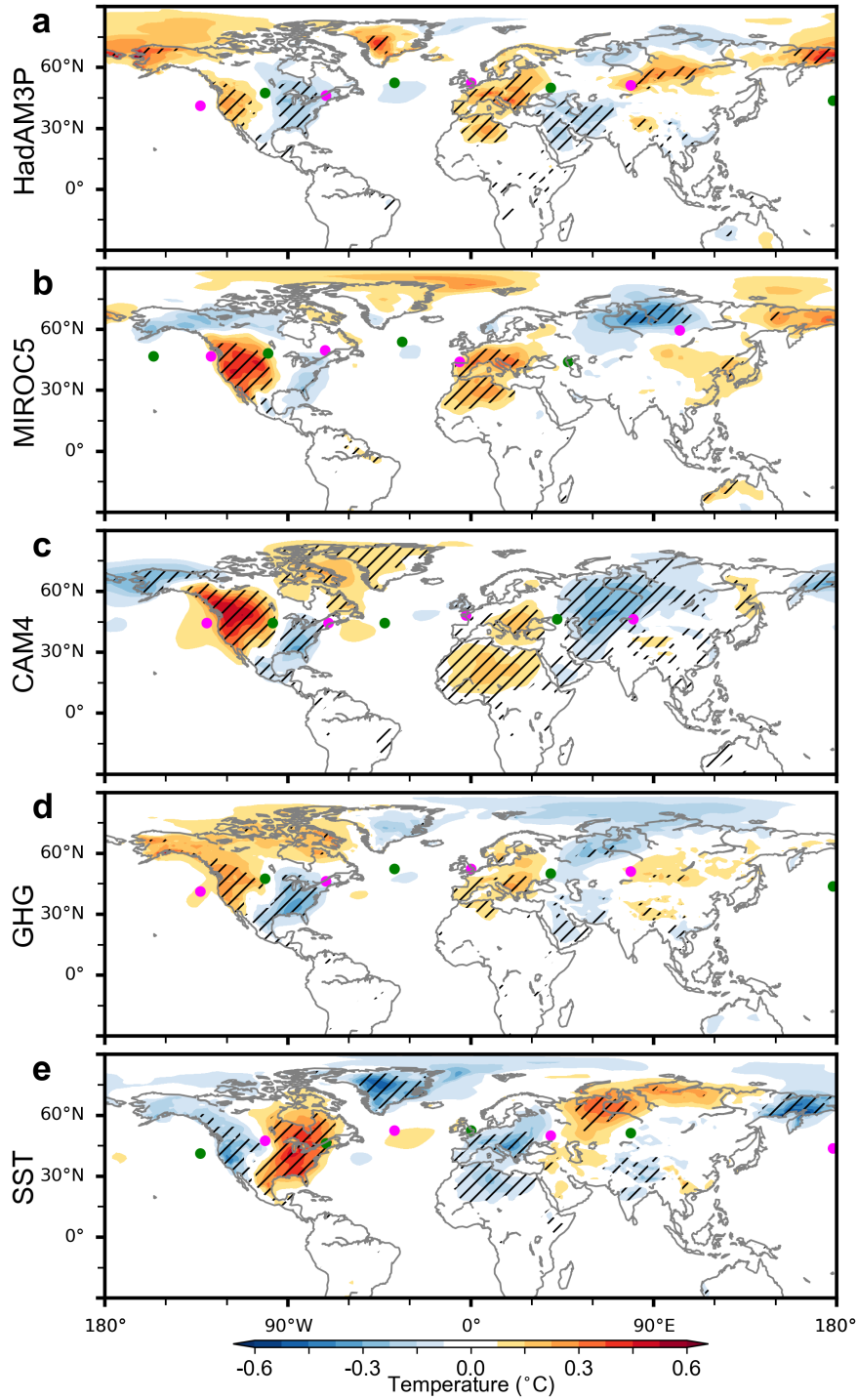


Figure 7.8: Effect of the wave on summer temperatures. JJA daily mean temperatures regressed onto the standardized circulation index for (a) HadAM3P CO₂, (b) MIROC5 CO₂, (c) CAM4 CO₂, (d) HadAM3P GHG and (e) HadAM3P SST. The hatching shows where the regression is significant at the 5% level using a two-sided Wald Test. Pink and green dots mark the positive and negative meridional wind wave centres respectively used to create the circulation index for each ensemble (see text for details).

heat extremes under higher CO₂ concentrations (investigated further in Chapter 8). This suggests that the direct circulation response to higher CO₂ concentrations may govern the regional differences in increases in extremes. It is worth noting that this regression approach likely underestimates the effect of the circulation on temperature as the circulation index is comprised of wave centres across the NH. As we will show in Section 7.4, the wave is actually forced from several different locations which contribute to parts of the wavetrain meaning the full wave is not necessarily present at all longitudes in each ensemble member. This suggests that if a single wave centre were used as the circulation index, there would be a larger local regression coefficient and therefore a larger effect on the local temperature.

A potential mechanism for the emergence of a stationary wave response to forcing is through changes in convection in the tropics forcing a Rossby wave that then propagates into the midlatitudes (Haarsma and Selten, 2012). These changes in convection can cause changes in upper-level divergence (with associated precipitation changes) which can be captured by changes in the RWS. One robust precipitation change is the decrease in precipitation in the tropical Western Atlantic and Caribbean Sea for the CO₂ forced cases (Fig. 7.1 and Figs. 7.4a,c), and an increase in the SST forced cases (Figs. 7.4b,d). A second robust change is the increase in precipitation in the tropical regions of Africa for the CO₂ forced cases (Fig. 7.1 and Figs. 7.4a,c), and a decrease in the SST forced cases (Figs. 7.4b,d). The reasons behind these observed changes will be explored in Section 7.6, for now we just seek to highlight these regions as potentially important regions where changes in the divergence lead to changes in the RWS. Another potential mechanism that could explain the stationary wave is by changes in the zonal mean westerlies in the sub-tropical upper troposphere (Simpson et al., 2016). The zonal mean westerly changes can cause changes in the midlatitude waveguides, affecting the propagation of Rossby waves. To investigate both these effects, we choose to focus on the wave response further in HadAM3P given the data available from the Hadley experiments for both GHG and SST forcing cases plus the fact that HadAM3P exhibits the clearest wave

response (see Fig. 7.1).

7.4 Barotropic model results

7.4.1 Full response

To investigate the wave pattern, we perform a series of experiments with the barotropic model. We choose a barotropic model as it is the simplest model where we can investigate both wave forcing from the tropics by applying an anomalous RWS, and also the effect of the background state on wave propagation by applying either the control background state or SST/GHG forced background state. The impact of changing the background state on wave propagation can be explained via the linear theory of barotropic Rossby waves in a zonal flow (Simpson et al., 2016), and so this effect is captured in the barotropic model. The RWS from the full model which acts as wave forcing can be backed out simply from the full model wind response field. The main disadvantages of the barotropic model are that it only has one level and so no vertical profile. This requires a choice of level as background state, and this choice is not always obvious (O'Reilly et al., 2018). It also requires strong meridional vorticity gradients to propagate waves across, meaning that sometimes wave responses will not appear fully in response to forcing where they perhaps might in complex GCMs or in the real world. Despite this, we employ the barotropic model as a tool to provide potential insight into the stationary wave pattern.

As outlined in Subsection 7.2.2, we compute the RWS (neglecting anomalous vorticity terms). We do this using the NAT ensemble mean as the 200 hPa climatological fields (overbar variables in equation 7.3) and define the anomalous fields (primed variables) as the GHG (or SST) ensemble minus the NAT ensemble 200 hPa fields. The reason for using the fields at 200 hPa is explained in Subsection 7.2.2. To test whether the barotropic model responses are linear, we multiply the input forcings by 10^{-6} (and then scale the

output fields back up by 10^6). Sensitivity experiments with and without the linearisation applied display no differences, and so the remainder of the barotropic experiments use the unscaled forcings. To first test that the barotropic model can produce the stationary wavelike response in the full GCMs, we force the barotropic model using the anomalous May RWS forcings from the GHG and SST ensembles, with the June NAT vorticity background state. The May forcing and June background state are chosen as sensitivity tests using different background state months and anomalous forcing months (not shown) show this combination gives the largest amplitude and clearest Rossby wave response, although forcing with the June RWS gives a similar, albeit slightly less well defined, wave. The June background state has the strongest vorticity gradients which are necessary for the barotropic model to propagate a wave response across. The July, August and JJA mean vorticity gradients are weaker, which suggests why the barotropic model with the June background state reproduces the wave response best. Also, we use the ensemble mean background state where vorticity gradients will have been smoothed from the individual members from the averaging, which potentially reduces the ability of the barotropic model to propagate a wave response.

The results of forcing the barotropic model with the GHG and SST RWS are shown in Fig. 7.9. Anomalous forcing is only applied between 0° - 10° N (pink box in Figs. 7.9a,b). The decision to use forcing between 0° - 10° N in the barotropic model clearly has an impact on the wave response, but without conducting an in-depth sensitivity study to determine the best forcing latitude and longitudes in different regions of the globe (which is beyond the scope of this chapter), we feel our choice is valid given we can reproduce the wave 5 pattern. The forcings in the GHG and SST cases are opposite in several regions including the Western Atlantic, Eastern Atlantic and in the Indian Ocean, just off the east coast of Africa. Anomalous RWS forcings arise from changes in tropical convective heating (Sardeshmukh and Hoskins, 1988), caused by the changes in precipitation in the tropics. The co-location of precipitation responses and anomalous RWS forcings suggest that the vertical motion associated with precipitation changes are

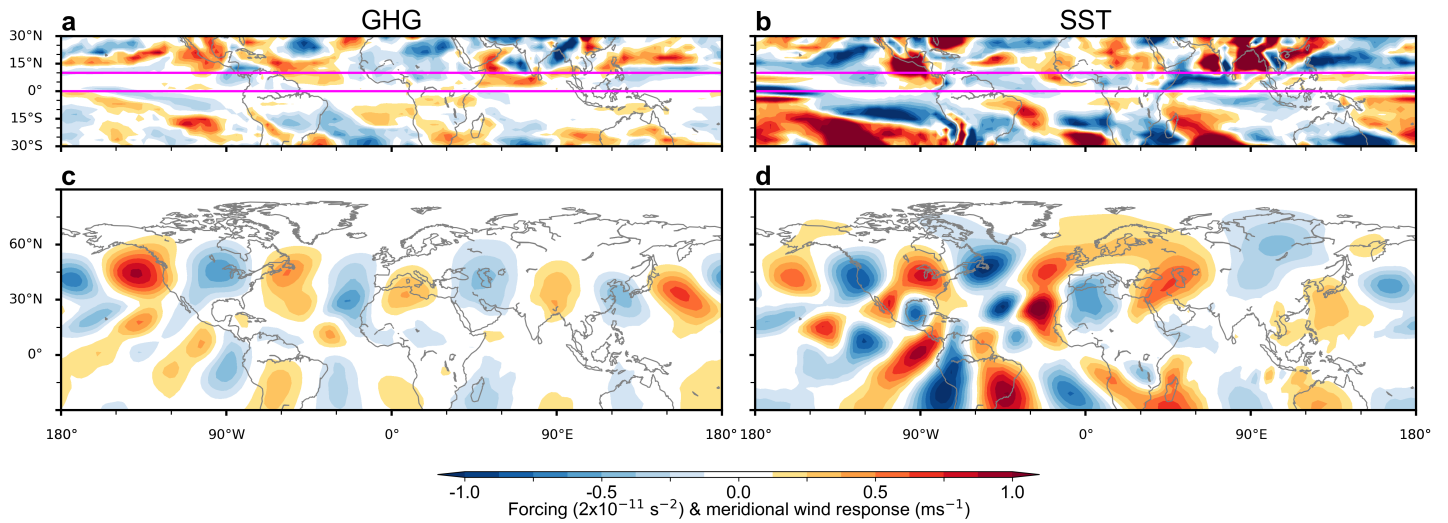


Figure 7.9: Barotropic model forcing and response. May Rossby wave source forcing from HadAM3P (a) GHG and (b) SST runs. The pink box demarcates the region used to force the barotropic model (0° - 10° N). Barotropic model meridional wind response averaged from day 16-20 for (c) the GHG forcing and (d) the SST forcing. Both cases use the NAT ensemble mean June 200 hPa vorticity as the background state.

responsible for the anomalous RWS forcings.

A comparison between the barotropic meridional wind responses (Figs. 7.9c,d) and the full model responses (Figs. 7.4c,d) reveals that the barotropic model can capture the main features of the full model. The wave pattern is clearly visible in both GHG and SST barotropic model experiments, with the same wavenumber as in the full model responses. Wave centres in the barotropic cases are shifted slightly eastward compared to the wave in the JJA mean in the Hadley experiments. However, the June mean wave centres in the Hadley experiments are also shifted by a similar amount eastward from the JJA mean (not shown), indicating that the shift between the wave in the barotropic model and the Hadley experiments could be due to the background winds differing between June and the JJA mean. Some wave centres are not as far poleward as in the full model response. This is evident in the jet exit regions, where climatological vorticity gradients are weak (e.g. over Northern Europe) and so a simple barotropic model will struggle to propagate the wave. Other reasons for the wave differences include the barotropic

model not capturing important feedbacks that occur in the full model, such as zonal wind changes, eddy-mean flow interactions, or potentially important non-linear orographic interactions (Ting, 1994). Nevertheless, the fact that the barotropic model results bear strong similarity to the full model is remarkable given the model’s simplicity.

Fig. 7.7c summarises the forcing used and the wave response in the barotropic model, taking the difference between GHG and SST RWS forcings and barotropic model meridional wind responses to highlight the areas of consistent change between the two experiments. The strongest RWS signal is in the Indian Ocean; there is a dipole in the tropical Atlantic and various consistent regions across the Pacific too. The stationary wave responses in the barotropic model are very similar to the GCM responses (quantified in Table 7.2), suggesting the GCM responses are indeed forced from the tropics. Correlations are lower than between the full model responses, this is partly because, as mentioned before, the wave centres are shifted slightly eastward in the barotropic model, and also because the wave centres are slightly further equatorward in the barotropic model.

To test whether the observed wave is a new forced wave, or an excitation of a wave in the climatological background state, we perform the same experiments after removing wavenumbers 4-6 from the background state. The barotropic model responses in this case still display the same wave pattern with wavenumber five (not shown), implying that the wave is a new forced wave due to the anomalous RWS. We also test whether the change of the background state has a substantial impact on the summer stationary waves (cf. to the effect of the background state on winter stationary waves in Simpson et al., 2016) by using the barotropic model with the GHG ensemble background state and comparing it to results using the NAT ensemble background state. If the background state were to be having a substantial effect, one would expect the waves to differ between the two experiments noticeably, however we observe no large difference in the wave responses (not shown). These results suggest that the stationary wave arises from anomalous wave forcing, not from changes in the zonal winds.

7.4.2 Sensitivity tests

To determine the forcing locations responsible for producing the wave response in the barotropic model, we perform a set of sensitivity experiments using the barotropic model. Following Barsugli and Sardeshmukh (2002), we apply a series of 64 cosine-square patch forcings of the form:

$$F'_i(\lambda, \phi) = A \cos^2 \left(\frac{\pi}{2} \frac{\phi - \phi_0}{\phi_w} \right) \cos^2 \left(\frac{\pi}{2} \frac{\lambda - \lambda_i}{\lambda_w} \right) \quad (7.7)$$

on the rectangle $\phi_0 \pm \phi_w, \lambda_i \pm \lambda_w$ (where ϕ is the latitude and λ is the longitude), and zero elsewhere. The half-widths of the patches are fixed at two gridpoints in both the latitudinal and longitudinal directions, yielding $\phi_w \approx 5.58^\circ$ and $\lambda_w = 5.625^\circ$. We apply the patch centres at a single fixed latitude $\phi_0 = 7.0^\circ$, whilst increasing λ_i in 5.625° increments starting at 0° , giving 64 patches in total. The amplitude of the patches, $A = 2.5 \times 10^{-11} \text{ s}^{-2}$, is chosen to be on the order of magnitude of, but slightly larger than, the forcing observed in the GHG forced ensemble so as to produce a strong signal in the barotropic model.

To assess the importance of a particular patch location in forcing the HadAM3P meridional wind response, we assign the following weight to each patch experiment response:

$$w_i = \frac{v_i^b \cdot \bar{v}}{\bar{v} \cdot \bar{v}}, \quad (7.8)$$

where v_i^b is the meridional wind response averaged between days 16-20 in the barotropic model for forcing F'_i , and $\bar{v}$ is the June ensemble mean meridional wind response from the HadAM3P model. The dot products are computed using cosine latitude weighting, and over the region 30°N - 70°N and all longitudes.

We perform two sets of sensitivity experiments, one using the HadAM3P NAT ensemble mean June background state, and the other using the NCEP Reanalysis 2 1979-2015 June mean background state. We then calculate a set of weights (comprising of 64

weights, one for each patch position) for each of the following four cases: the meridional wind responses of the barotropic model using the NAT ensemble mean background state projected onto the GHG ensemble mean meridional wind response, NAT background onto SST ensemble wind response, NCEP2 background onto GHG ensemble wind response and NCEP2 background onto SST ensemble wind response.

We calculate the weights for the sensitivity experiment responses with the full model GHG (Fig. 7.10a) and SST (Fig. 7.10b) responses in the 200 hPa meridional wind. Weights are shown for NAT background (blue). The RWS forcings from the full model (used to force the barotropic model in Fig. 7.9) averaged over 0° - 10° N are shown in red. Red lines are normalised so the maximum in each figure is the same as the respective blue line showing the NAT weights. Lines in Fig. 7.10b have been multiplied by -1 to allow direct comparison with Fig. 7.10a. Regions where the amplitude of the full model forcing and weights agree and are large indicate regions where forcing is likely to be important in producing the full model wave responses. From the weights, we identify three key regions: the Western Atlantic, Eastern Atlantic and in the Indian Ocean, just off the east coast of Africa (marked by the crosses in Figs. 7.10c,d,e). For the SST case, there also seems to be a region of importance in the Pacific. Here we focus on the other three regions as they appear important in both GHG and SST cases. Figs. 7.10c,d,e show the wave responses for the HadAM3P background case at each of the identified forcing locations. The forcing in the Eastern Atlantic has a negative weight, and so we multiply the response of this patch experiment by -1 to give the correct meridional wind response. When the three forcings are combined (Fig. 7.10f), they reproduce the wave response that the full model forcing produces in the barotropic model (compare to Fig. 7.9c). We combine the responses by summing them linearly to produce the response in Fig. 7.10f, but we verify that applying the three forcings in the same run produces the same wave response (not shown) indicating that the forcings combine linearly.

To test the robustness of the response to the identified forcing locations and jet conditions, we analyse the sensitivity results from the barotropic experiments run with the

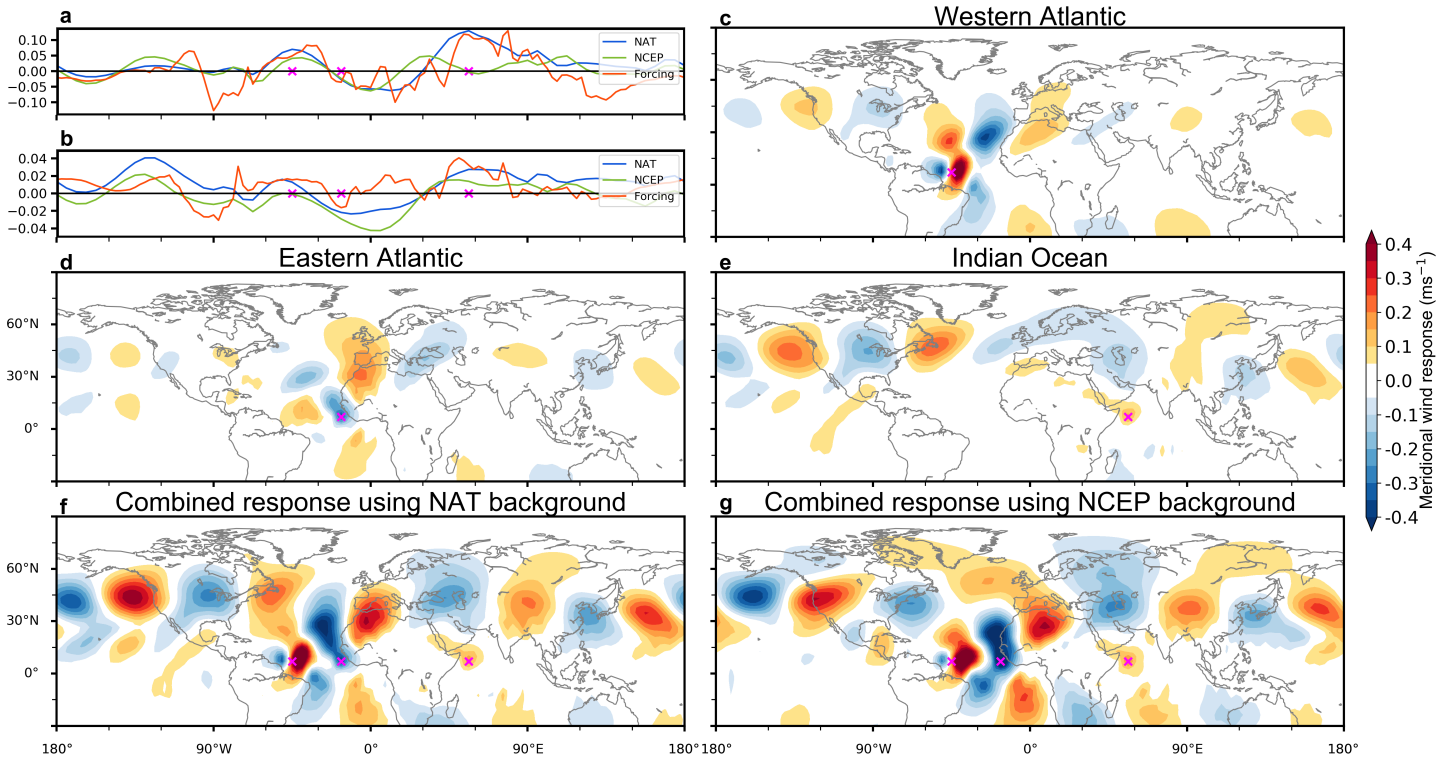


Figure 7.10: Idealised barotropic model results. Weights from idealised patch experiments for NAT and NCEP2 backgrounds and HadAM3P Rossby wave source for (a) GHG ensemble response and (b) SST ensemble response. Rossby wave source is averaged over 0° - 10° N and then normalised so its maximum corresponds to the June weight maximum. The weights in the SST case have been multiplied by -1 . Idealised meridional wind responses are shown for (c) West Atlantic, (d) East Atlantic and (e) Indian Ocean forcing patches (location marked by pink cross). Combined response of Western and Eastern Atlantic and Indian Ocean forcings with (f) NAT background state and (g) NCEP2 background state.

NCEP2 background state. The green lines in Figs. 7.10a,b show the weights when using the NCEP2 background state. The NCEP2 weights are broadly consistent with the NAT weights, albeit with a weaker sensitivity to the Indian Ocean region. When forced at the same three identified locations (marked by the crosses) the combined wave response is broadly the same as when using the NAT background state (Fig. 7.10g), with differences mainly arising over jet exit regions for the reasons previously discussed regarding weaker climatological vorticity gradients. From these sensitivity experiments we can conclude that there exist three important forcing locations to produce the full model wave response, and that the wave response also occurs on an observed background state

(the NCEP2 case), which implies that the wave responses are not just possible in the three GCMs, but also in reality.

7.4.3 Discussion

The replication of the wave 5 in the barotropic model is only possible using the June background state from HadAM3P (in both the Hadley experiments and 1.5°C experiments) and from reanalysis. We tried different months and also using the CAM4 and MIROC5 200 hPa vorticity fields as the background state. In all other cases besides the June reanalysis and HadAM3P cases, a forced wave is present, but as a wave 4 pattern and not as a circumglobal wave. We investigate the three model background states and their biases to suggest possible reasons for this.

The first thing to note is that in MIROC5 and CAM4 the wave pattern in the ensemble mean is weaker, suggesting that the background states of these two models might be less conducive to propagating the wave response to forcing. Fig. 7.11 shows the 200 hPa June ensemble mean zonal wind biases of the three models compared to the 1979-2015 200 hPa June mean zonal wind from the NCEP2 Reanalysis. The Pacific jet in both CAM4 and MIROC5 exhibits poleward biases, whilst in HadAM3P, the jet has a weak bias. All three model jets exhibit a poleward biased jet over land. Over the Atlantic, the biases in HadAM3P are again the smallest, with only a slightly weak biased jet compared to strong biased jets in the other two models. The HadAM3P jet is narrower in the zonal mean than the MIROC5 and CAM4 jets, and it has been suggested that this increases the ability of a jet to act as a waveguide (Manola et al., 2013). The biases in the zonal wind speeds could also be affecting the waveguiding properties of the jet due to the dependence of Rossby wave phase speed on jet speed (Ambrizzi et al., 1995). In the mean states, the HadAM3P jet exhibits a smaller poleward bias, meaning the waveguide is closer to the potential wave forcings in the tropics. This could lead to a stronger wavetrain, as seen in the HadAM3P full model response, as anomalous diver-

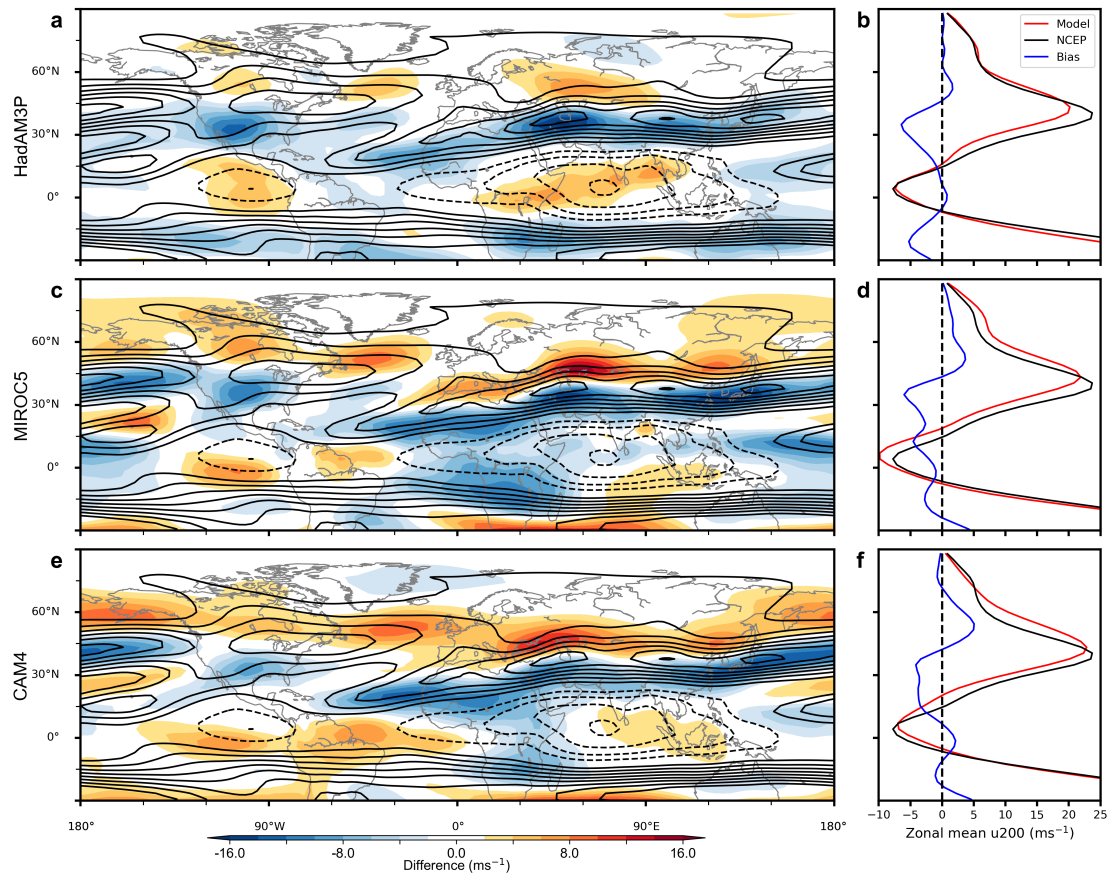


Figure 7.11: Model zonal wind biases for June in (a,b) HadAM3P, (c,d) MIROC5 and (e,f) CAM4. Contours in (a,c,e) show the June 1979-2015 200 hPa zonal wind from the NCEP2 reanalysis (contours at 5 ms^{-1} intervals with the zero contour suppressed). Colours in (a,c,e) show the difference between the June 200 hPa zonal wind in the model (2006-2015) and the NCEP2 reanalysis (1979-2015). (b,d,f) show the zonal mean zonal winds and the bias, defined as the difference between the model and NCEP2 reanalysis means, shown in blue.

gent winds in the tropics are closer to the vorticity gradients in the midlatitudes required for the wave to propagate. The narrowness and lack of a poleward bias in HadAM3P may suggest why using it as the background state for the barotropic model can give the wave 5 pattern whereas the other models cannot. The larger biases in the ensemble mean states of MIROC5 and CAM4 may explain why using a mean background state means the barotropic model is not able to reproduce the wave.

The ensemble mean background vorticity states are also smoothed in comparison to the individual model background states which allow the wave to occur in the full model

responses. This smoothing leads to weaker vorticity gradients, which are required to be strong for the barotropic model to propagate the wave response. Given that the response in the ensemble mean is already weaker in CAM4 and MIROC5, the background states in these models may be less conducive for the wave to occur. It is possible that the necessary vorticity gradients are not present in the CAM4 and MIROC5 smoothed background states, thus preventing the barotropic model from propagating the wave.

Differences in the three-dimensional structure could also be important. We use the 200 hPa background vorticity state as this level has the largest anomalous tropical forcing and because the vorticity gradients, upon which the signal propagates out of the tropics, are stronger in the upper-troposphere. We find it necessary to have basic states with strong waveguides to achieve the response pattern. In reality, the equivalent barotropic level for external Rossby waves is lower in the troposphere (Held et al., 1985), and so lower model levels could be important in allowing full model to produce the wave response.

With the large number of individual ensemble members, and the fact that the wave response is largest and biases smallest in HadAM3P, we have focused on analyzing the mean state forcing and background state, and just in the HadAM3P model. The logical next step to investigate and confirm the forcing mechanism from the tropics, and to reproduce the wave using the MIROC5 and CAM4 background states, is to perform experiments using a three-dimensional baroclinic model where we could cleanly test the tropical forcing hypothesis and the hypothesis that the changing background state is not important. However, in this thesis we do not carry out this more involved modelling approach.

7.5 Forcing mechanisms

We identify two classes of mechanism which cause the changes in tropical precipitation in response to the CO₂ and SST forcings. The first of these is due to thermodynamic

changes. Increases in atmospheric CO₂ concentration lead to increases in the static stability over the tropical oceans. This is because the lapse rate in the tropics is close to the moist adiabatic lapse rate which decreases as the atmosphere warms; a warmer air parcel can contain more water vapour when it becomes saturated, thus releasing more latent heat as it rises and the water vapour condenses out (Held and Soden, 2006). This means that the upper troposphere will warm more than the lower troposphere, thus increasing the static stability and suppressing convection (Sugi and Yoshimura, 2004; Andrews et al., 2009; Dong et al., 2009; Cao et al., 2012; Bony et al., 2013). On the other hand, an increase in SSTs results in increased atmospheric moisture, leading to an increase in precipitation across tropical wet regions (Held and Soden, 2006; Seager et al., 2010). These thermodynamic effects are responsible for the model changes in precipitation over the tropical Western Atlantic. In the CO₂ forcing cases, the suppressed convection leads to a decrease in precipitation in the tropical Western Atlantic (Fig. 7.1 and Figs. 7.4a,c). In the SST forcing cases, the enhanced moisture leads to an increase in the precipitation in the tropical Western Atlantic (Figs. 7.4b,d).

The second mechanism arises from the land-sea θ_e contrast. Increases in CO₂ cause the land to warm in response to increased downwelling longwave radiation, enhancing the land-sea θ_e contrast (Shaw and Voigt, 2015) and leading to increases in precipitation over land (Chadwick et al., 2014; Richardson et al., 2016). However, when the SSTs are increased, the land-sea θ_e contrast changes in the opposite direction (Shaw and Voigt, 2015), causing an increase in the vertical velocity over the oceans (likely due to a reduction in static stability) thus enhancing convective precipitation over the oceans (Chadwick, 2016). The changes in the land-sea θ_e contrast in the GHG and SST responses are shown in Fig. 7.12. In the GHG case (Fig. 7.12a) on the east coast of Africa, there is a positive land-sea θ_e gradient (along the blue arrow in Fig. 7.12a) and associated increase in precipitation over the same region in the CO₂ forcing cases (Fig. 7.1 and Figs. 7.4a,c). On the west coast Sahel region of Africa however, there is a negative land-sea θ_e gradient (black arrow) which is associated with a decrease in precipitation. The changes in θ_e

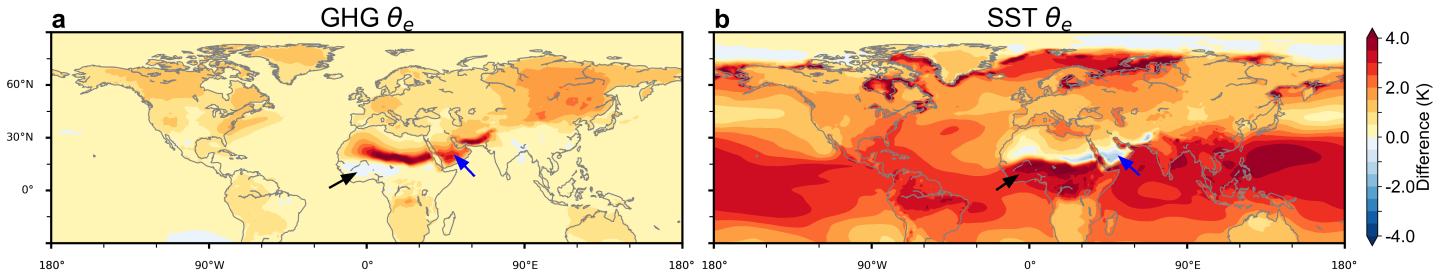


Figure 7.12: Responses to GHG and SST forcing. Surface θ_e anomaly for (a) HadAM3P GHG response and (b) HadAM3P SST response. See Subsection 7.2.3 for details of calculation.

correspond to the RWS forcing in the two regions: positive RWS for a positive land-sea θ_e contrast and negative RWS for a negative contrast (see Fig. 7.9a). These regions of positive and negative RWS forcing are consistent with the Eastern Atlantic weight being opposite in sign to the Western Atlantic and Indian Ocean weights in Fig. 7.11a. In the SST case, the land-sea θ_e gradient is negative on the east coast of Africa (along the blue arrow in Fig. 7.12b), and positive on the west coast Sahel region of Africa (black arrow). These θ_e gradients again have corresponding changes in precipitation and RWS forcings: a decrease in precipitation (Figs. 7.4b,d) and negative RWS forcing on the east coast of Africa and an increase in precipitation and positive RWS forcing in the west coast Sahel region of Africa. Again, this is consistent with the opposite sign of the Eastern Atlantic weight in Fig. 7.11b. This forcing mechanism is the same mechanism responsible for summer stationary wave changes in response to direct radiative and SST forcings found by Shaw and Voigt (2015).

There is also a large change in RWS just off the west coast of South America (around 90°W; Fig. 7.10a), which presumably arises due to the same mechanisms causing the other changes in RWS at land-sea interfaces. However, forcing the barotropic model here does not produce a response which projects onto the full model wave response. This may be due to the meridional distance of this RWS from large enough vorticity gradients needed to act as a waveguide.

The response of the land-sea θ_e contrast on the west coast of Africa is tied to the

precipitation response in the region, which arises due to a shift in the ITCZ (Fig. 7.4). The ITCZ shifts toward the anomalously warm hemisphere (e.g. Bischoff and Schneider, 2014, and Chapter 4 of this thesis): in the GHG case the ITCZ shifts northwards as the NH warms more due to having more land; in the SST case, the ITCZ shifts southwards as the Southern Hemisphere warms more due to having more ocean. Previous work has found the same sign ITCZ shifts in this region as we find in this chapter: northwards due to greenhouse gas forcing and southwards due to global SST warming (e.g. Hoerling et al., 2006; Skinner et al., 2012). In the GHG case, on the west coast where the precipitation anomaly is negative, θ_e decreases going from ocean to land, consistent with a decrease in convection. In the SST case, the precipitation anomaly is positive, with an increase in θ_e from ocean to land, consistent with an increase in convection. The convection responses which arise from the ITCZ shift lead to the RWS responses over the Eastern Atlantic, which are opposite in sign compared to the Western Atlantic and Indian Ocean regions in both the GHG and SST responses.

We identify two key coastal regions just off the east coast of continents, the east coast of South America and just off the east coast of Africa (co-located with the western part of the Indian Ocean basin), and a further region just off the west coast of Africa. Positive RWS forcing in the east coast regions associated with decreased upper-level divergence, and negative RWS forcing in the west coast region associated with increased upper-level divergence, acts to force the NH stationary wave in summer.

The regions we identify for forcing this particular wave have been shown to be important in forcing extratropical Rossby waves in other studies. Changes in outgoing long-wave radiation in observations in summer over the tropical Atlantic, coinciding with the regions where we see precipitation changes in this chapter, have been associated with summer circulation patterns over Europe (Cassou et al., 2005). The Summer East Atlantic pattern, as observed in reanalysis by Wulff et al. (2017) has an associated circumglobal wavenumber 5 wave, which is linked to an anomalous tropical precipitation pattern with similarities to the precipitation patterns we find in this chapter: drying in

the Western Atlantic and wetting over the Sahel region in Africa, producing a similar convergent forcing as we find in this chapter. Yasui and Watanabe (2010) also find that precipitation changes over equatorial Africa are important in forcing the observed circumglobal teleconnection (CGT, see Ding and Wang, 2005) in boreal summer. The wave we observe is not an excitation of the CGT: the wave centres we observe do not coincide with the CGT, and as demonstrated in Section 7.4, the wave we force is not an excitation of a wave present in the background state. However, the fact that studies on the CGT find similar key forcing regions as this chapter suggests that these regions are important generally in forcing boreal summer Rossby waves. As well as equatorial Africa, the Indian summer monsoon region (which includes the area of ocean just off the east coast of Africa) has also been found important in forcing aspects of the CGT (Ding and Wang, 2007), and other NH teleconnection patterns (Lin, 2009).

7.6 Discussion

The opposing effects of radiative forcing versus sea surface warming have important implications on regional climates and extremes. Framed in terms of a global mean temperature target, the changes in regional climates and extremes depend significantly on the climate sensitivity (defined as the global mean temperature warming due to a doubling of atmospheric CO₂ concentrations), which is subject to large uncertainties (IPCC, 2013). If the Earth has a low climate sensitivity, the temperature target may be met, but with corresponding dangerous changes in certain regional extremes, due to increases in tropical precipitation over land and the wavelike temperature increase pattern over the NH. For higher climate sensitivities, less CO₂ would be emitted to reach the same temperature target, and so the opposing effect of the sea surface warming would balance out the direct CO₂ effect to a greater extent. If the climate sensitivity is very high, the direct CO₂ effect will be relatively unimportant compared to the sea surface warming effect, leading to drying over the land in the tropics, and different regions (those under the

anticyclonic regions of the SST stationary wave pattern) experiencing increases in heat extremes. Geoengineering techniques aiming to tackle climate change impacts without removing CO₂ would exacerbate the direct CO₂ effect when stabilising at a temperature target as the techniques would reduce global mean temperature without removing CO₂. Ultimately, the interplay of the two opposing effects will be dictated by the Earth's climate sensitivity, and further work is needed to quantify the exact magnitudes of the direct CO₂ and sea surface warming effects on changes in precipitation and on the forced stationary wave.

On the question of the interaction between the direct CO₂ effect and sea surface warming effect, Lambert et al. (2011) found no substantial nonlinearities in indices of atmospheric circulation when the doubling of CO₂ and spatially uniform SST warming were added to an atmosphere-only GCM. Deser and Phillips (2009) also observe a high degree of linearity in the response to direct CO₂ effects and sea surface warming in atmospheric circulation trends during 1950-2000 simulated in an atmosphere-only GCM. This suggests the opposing effects on circulation may combine linearly as in the barotropic modelling experiments in this chapter, although further work is needed to confirm this.

The SST forcing model results presented here rely on one global warming pattern change, and the direct CO₂ runs all have the same fixed SST pattern. However, it has been suggested that the pattern of SST warming plays an important part in certain regional circulation responses (He and Soden, 2015) and the tropical precipitation response to SST warming (Chadwick, 2016). The robustness of the wave response in the AMIP4K response suggests that the wave is forced by changes in mean SST warming, but further work investigating the uncertainty arising from the SST pattern change is needed before quantitative predictions about the opposing effects of radiative forcing and sea surface warming can be made.

There are several other possible contributing factors to the direct CO₂ effect, which require further study to quantify. Radiative changes in clouds have been linked to cir-

culation changes (Voigt and Shaw, 2015; Ceppi and Shepherd, 2017) and large scale cloud cover changes over the NH in summer are associated with the direct CO₂ effect (Richardson et al., 2016), with a decrease in cloud cover over regions that see strong increases in temperature due to CO₂. However, determining the causality between the cloud cover changes and the temperature changes is a complex issue. One possibility is that the anticyclonic regions of the forced stationary waves cause cloud to be less prevalent in these regions, adding to the increases in local temperature. Midlatitude stationary wave changes can also be affected by changes in the zonal mean westerlies in the sub-tropical upper troposphere (Simpson et al., 2016). We do not observe this effect in this chapter, and Grise and Polvani (2014) show that whilst there are differing effects on the zonal circulation under direct radiative forcing and sea surface warming, the effects are not opposite. However, with large variations among future predictions in zonal wind changes (Simpson et al., 2014), this effect could be important in dictating the characteristics of the forced stationary wave.

A modelling intercomparison project as part of the new CMIP6 ensemble, The Cloud Feedback Model Intercomparison Project Phase 3 (CFMIP3) (Webb et al., 2017), consists of AMIP style simulations designed to investigate effects on clouds and circulation of CO₂ radiative effects, physiological effects, SST uniform warming and SST pattern effects. Chadwick et al. (2017) use several models from CFMIP3 to demonstrate in winter that the physiological effect and SST pattern change play important roles in the hydrological cycle and European circulation changes. The full CFMIP3 ensemble will enable more quantitative predictions and analysis surrounding the opposing effects of radiative forcing and sea surface warming in summer to be carried out.

7.7 Conclusions

Using results from three GCMs and CMIP5, we have shown here that there are opposing effects of direct CO₂ forcing and sea surface warming on tropical precipitation and

forced summer NH stationary waves. The direct effect of CO₂ forcing leads to a drying over the land and wetting over the ocean in the tropics in boreal summer in key forcing locations. This forces a wavenumber 5 stationary wave pattern over the NH. We observe the opposite precipitation response due to sea surface warming, which forces the same wave over the NH but with opposite sign. The three key locations for forcing the summer wave are found to be the Western Atlantic, Eastern Atlantic and in the Indian Ocean off the east coast of Africa. These forcing locations are diagnosed using a barotropic model, but further work with a three-dimensional baroclinic model would help to confirm the forcing locations. Further work using coupled models would also be useful to quantify the linearity of how the sea surface warming and direct radiative forcing effects combine, whilst also allowing a quantification of which of the two effects might be expected to dominate in a fully coupled system.

This chapter has important implications for changes in regional climates and extremes which depend heavily on the Earth's climate sensitivity and land-sea temperature contrast response. The opposing effects of radiative forcing and sea surface warming and their relative importance at a specific global mean temperature are determined by how much CO₂ is emitted to reach such a temperature. A low climate sensitivity means a greater contribution from the direct CO₂ effect, i.e. enhanced precipitation and precipitation extremes over tropical land, and increases in heat extremes under the anticyclonic centres of the CO₂ forced stationary wave. On the other hand, a high climate sensitivity means a greater contribution from the sea surface warming effect, i.e. reduced precipitation over tropical land. It also means that the increases in heat extremes are shifted to the cyclonic centres of the CO₂ forced stationary wave, which are anticyclonic for the sea surface warming forced wave. The future change in land-sea temperature contrast will also play an important role in determining the balance between the opposing effects. A larger land-sea temperature contrast response could lead to a larger forcing of the direct CO₂ wave. The relative magnitude of the combination of the opposing effects of direct radiative forcing and sea surface warming is still uncertain due to uncertainty

in the Earth's climate sensitivity and land-sea temperature contrast response (Byrne and O'Gorman, 2013). This chapter points to the need to reduce the uncertainties in the climate sensitivity and land-sea contrast response to better constrain how regional precipitation and temperature patterns will change under global warming.

Chapter 8

Impacts of direct CO₂ radiative forcing

In the previous chapter, we investigated the opposing effects of radiative forcing and sea surface warming on the atmospheric circulation. We found that direct CO₂ forcing produces a wavenumber 5 stationary wave over the Northern Hemisphere in summer. We showed sea surface warming produces the same wave, but in antiphase with the CO₂ forced wave. By analysing tropical precipitation changes, equivalent potential temperature changes, and results from a simple barotropic model, we showed that the wave is forced from the tropics, particular the West Atlantic, west coast of Africa and off the Horn of Africa. The stationary wave has a significant impact on regional temperature anomalies in the Northern Hemisphere summer, with a potential direct effect of CO₂ on temperature extremes. In this chapter, we will look more closely at the changes in extremes caused by the direct effect of CO₂, and explore the implications of these changes in extremes on current climate policy.

8.1 Introduction

The long-term goal of the Paris Agreement (United Nations/Framework Convention on Climate Change, 2015) aims to strengthen the global response to the threat of climate change by: *‘Holding the increase in the global average temperature to well below 2°C*

above pre-industrial levels and to pursue efforts to limit the temperature increase to 1.5°C above pre-industrial levels, recognizing that this would significantly reduce the risks and impacts of climate change.' To achieve the warming goal, the agreement implements an explicit temperature-based goal-driven regime, with which mitigation efforts are intended to be revised to over time. Therefore, no explicit long-term emissions targets are directly associated with the agreement aims, aside from the need to ultimately achieve a balance between net anthropogenic greenhouse gas sources and sinks into the atmosphere (Fuglestedt et al., 2018). Since the Paris Agreement, there has been a call for research into impacts associated with 1.5°C and 2°C globally-average surface temperature anomalies (Mitchell et al., 2016; Schleussner et al., 2016). However, previous work (Steinacher et al., 2013; Seneviratne et al., 2016) suggests global mean temperature targets alone are unable to comprehensively limit the risks from anthropogenic emissions, and that global mean temperature is not the sole driver of changes in temperature and precipitation patterns (Folland et al., 1998; Sexton et al., 2003; Bony et al., 2013; Dong and Sutton, 2015; Richardson et al., 2016) and extremes (Kamae et al., 2014a; Seneviratne et al., 2014; Dong et al., 2016).

Here we isolate the direct effect of CO₂ concentration, investigating the difference between estimated likely (>66% probability that the value will lie within this range) upper and lower bounds on CO₂ concentrations leading to 1.5°C of global warming in 2100 for adaptive pathways to meet the Paris Agreement goals under climate response uncertainty (Millar et al., 2017). We define the direct effect of CO₂ concentration as all the effects of CO₂ on climate beside those occurring through ocean warming but including feedbacks over land (e.g. from soil moisture), therefore excluding most of the global temperature change since the latter is to a large extent determined by ocean temperatures. We run five AMIP style ensembles (Taylor et al., 2012) with HadAM3P and MIROC5: a present day ensemble (2006-2015); three ensembles with sea surface temperature (SST) levels equivalent to 1.5°C global mean warming relative to 1850-1900 with low, best-estimate and high CO₂ concentrations; and a 2.0°C ensemble with best-estimate CO₂

concentrations and SSTs for a 2.0°C world. By considering the difference between high and low CO₂ concentration ensembles, we find substantial differences in the patterns of change in temperature and precipitation, pointing to an increase in extreme event likelihood. These differences could have important consequences if international climate policy does not seek to limit CO₂ concentrations and instead concentrates exclusively on the global mean temperature target. This idealised experimental design helps focus on the risks associated with different atmospheric CO₂ concentrations, consistent with different climate responses, after global temperatures have stabilised at approximately the same level. We also compare the results from HadAM3P and MIROC5 to results from CAM4, drawing similar conclusions.

8.2 Methods

8.2.1 Experimental design

As stated above, in this chapter we make use of the three climate models used in Chapter 7: HadAM3P, MIROC5 (Watanabe et al., 2010) and CAM4 (Gent et al., 2011), at the same resolutions as before. The experimental setup follows the Half a degree Additional warming, Prognosis and Projected Impacts (HAPPI; www.happimip.org) design (Mitchell et al., 2017a). Forcing conditions are as in the AMIP design, including SSTs and sea ice (Taylor et al., 2012). The HAPPI experiments are designed to simulate conditions in the present decade (2006-2015), and 1.5°C and 2.0°C warmer than pre-industrial (1861-1880) conditions. SSTs for the 1.5°C case are calculated by adding to the observed 2006-2015 SSTs a change in SST between the decadal average of the modelled 2006-2015 period and the decadal average of the modelled 1.5°C world over 2091-2100 (Mitchell et al., 2017a). Hence the SST patterns are still time varying because they are based on the 2006-2015 observations, but they have an additional warming added to them. As CMIP5 historical simulations stopped in 2005, the decadal average of the

2006-2015 SSTs is estimated from RCP8.5 simulations, as this is the scenario that is closest to observations over this period. The decadal average of the 2091-2100 SSTs is estimated from CMIP5 RCP2.6 simulations. The process to calculate the 2.0°C SST pattern is similar and outlined in detail in Mitchell et al. (2017a). Here we use the HAPPI Tier 1 experimental design (Mitchell et al., 2017a), which uses the multi-model mean patterns, thus across all three models used in this chapter, all the present day ensembles are run with an identical present day SST pattern, all the 1.5°C ensembles are run with an identical 1.5°C SST pattern, and all the 2.0°C ensembles are run with an identical 2.0°C SST pattern. Full details are discussed in Mitchell et al. (2017a).

For HadAM3P, we run five ensemble experiments, each over a 10-year period. A present day ensemble is run using HAPPI present day setup (Mitchell et al., 2017a) over the period 2006-2015 with an average CO₂ of 390.4 ppm (90 ensemble members). A 1.5°C ensemble is run using HAPPI 1.5°C setup (Mitchell et al., 2017a), which uses the RCP2.6 forcing scenario boundary conditions from the last decade of the 21st century, with CO₂ fixed at 423.1 ppm (71 members). Two further ensembles are run using the 1.5°C setup (Mitchell et al., 2017a), but with CO₂ fixed at 395.8 ppm and 550.0 ppm (76 and 88 members) to represent the lower and higher likely CO₂ concentrations averaged over 2091-2100 in adaptive pathways that succeed in achieving warming below 1.5°C in 2100 for the assessed ranges of climate response uncertainty (Millar et al., 2017).

The range of CO₂ concentration is intended to illustrate the scale of the difference in concentrations that may be consistent with a 1.5°C world. Thus the range of concentrations used does not affect the qualitative results of changes in extremes. As the concentrations are only used to fit the regression model and not calculate changes in the carbon budget directly, the reduction in the upper bound of the carbon budget is not influenced by the exact choice of CO₂ concentrations. It is however influenced by the selection of models used in the chapter, due to their different climate sensitivities. A similar range in concentrations can be obtained by converting the spread of individual model temperatures in CMIP5 under RCP2.6 at the end of the century (2081-2100) to a

concentration range. To see this, we start by noting that the change in temperature at the end of the century from the pre-industrial baseline stated in the IPCC Fifth Assessment Report (IPCC, 2013) is 1.6°C with a standard deviation of 0.4°C and 5-95% uncertainty range of $0.9\text{-}2.3^{\circ}\text{C}$. The radiative forcing averaged over 2081-2100 is 2.60 Wm^{-2} , from which we can compute the ratio of radiative forcing to temperature. When multiplied by 1.5°C and converted to a concentration range, this gives a range of 400.2-510.2 ppm using the range of temperatures associated with one standard deviation from the mean, or 381.6-624.7 ppm when using the 5-95% range.

A 2.0°C ensemble is run with the HAPPI 2.0°C SST pattern (Mitchell et al., 2017a) and CO_2 fixed at 486.6 ppm (96 members). All CO_2 concentrations are prescribed as a global mean atmospheric concentration. Following previous work (Massey et al., 2015), initial condition perturbations are applied between ensemble members via perturbations to the potential temperature. For MIROC5, we perform exactly the same ensemble runs, with 50 members per experiment using the same CO_2 concentrations as HadAM3P. For CAM4, we use the data from three experiments, a present day, 1.5°C , and 1.5°C with a CO_2 concentration of 379.0 ppm (thus for CAM4, the 1.5°C ensemble with CO_2 at 423.1 ppm doubles as the ‘best-estimate’ and ‘high’ ensembles and the 1.5°C with CO_2 at 379.0 ppm becomes the ‘low’ ensemble), all using the HAPPI experimental design with 501 ensemble members.

Following standard AMIP design protocol (Taylor et al., 2012), soil moisture is allowed to vary freely in our simulations. The primary aim of this chapter is to disentangle the direct CO_2 effect on regional climate from the ocean warming, since the latter strongly affects global mean warming and, additionally, may cause some large scale circulation changes. SST patterns can contribute to regional climate variability far further afield than their perturbation, whereas soil moisture effects are more local and do not affect global mean temperature significantly (Vogel et al., 2017). Some of the regional effect, and indeed perhaps a large part of this effect over land areas (Vogel et al., 2017), is due to changes in soil moisture. Holding soil moisture constant would mask some of

the changes due to the direct effect of CO₂, preventing us from disentangling this effect from that of the global mean warming of the ocean.

We select the HAPPI experimental design as it allows us to isolate the direct effect of changing CO₂ on climate extremes. However, other possible experimental designs do exist where it would be possible to investigate this effect too (Palter et al., 2018; Sanderson et al., 2017). In the setup described in Palter et al. (2018), 1.5°C warming is arrived at by 2100 under two different emissions scenarios, one keeping concentrations below 440 ppm, and the other overshooting and then ramping down CO₂ concentrations to arrive at 1.5°C. In this setup however, there are many other differences between the simulations at 1.5°C, e.g. sea ice-cover, and so the setup makes it hard to disentangle the direct response on extremes due to differing atmospheric CO₂ concentrations. In the setup described in Sanderson et al. (2017), first the CO₂ emissions are determined that drive a particular coupled model to arrive at 1.5°C and 2.0°C global warming, and then the coupled model is run with this CO₂ concentration. Unlike the HAPPI setup, this guarantees the physical consistency of the setup and does not result in an SST response and CO₂ concentration that are potentially inconsistent. However, to then investigate the impact of changed atmospheric CO₂ concentration on extremes, one would then have to compare the impact across different models, making it hard to disentangle the direct effect of CO₂ from all the other differences that may give rise to that difference in sensitivity (e.g. model physics and SST anomaly pattern).

8.2.2 Regression model

To compute the contribution of changes in radiative forcing and changes in global mean temperature to a change in a particular variable, we fit the changes in the 10-year ensemble mean variables between present and 1.5°C runs to the model: $\Delta X = \alpha \Delta F + \beta \Delta \bar{T}$. ΔX is the change in variable. ΔF is the change in the radiative forcing due to CO₂ from the present decade, $\Delta F = \frac{3.71}{\ln 2} \cdot \ln \left(\frac{C}{C_p} \right)$ (IPCC, 2013) where C and C_p are the CO₂

concentration in the forced run and the present decade respectively in ppm. $\Delta\bar{T}$ is the change in global mean temperature between the 1.5°C ensembles and the present day ensemble (not the change in mean SST, thus $\Delta\bar{T}$ is different for the three 1.5°C ensembles). The regression framework allows us to account for the change in variables from the increase in global mean temperature arising from increased CO₂ concentrations, and so separate the effect of direct radiative CO₂ from the effect of the global mean temperature increase. The fit parameters α and β indicate the dependence of the change in variable to changes in CO₂ radiative forcing and global mean temperature respectively. Uncertainty estimates in warming for a given radiative forcing are calculated from the covariance matrix of α and β which accounts for the spread in 10-year ensemble member mean quantities.

To apply the global mean temperature change correction, for each variable we subtract $\beta (\bar{T}_H - \bar{T}_{BE})$ from the high CO₂ ensemble variable, and add $\beta (\bar{T}_{BE} - \bar{T}_L)$ to the low CO₂ ensemble variable (subscripts: H - high ensemble, BE - best-estimate ensemble, L - low ensemble). For maps, β is calculated for each individual grid-point, for area mean indices, β s are calculated from the area mean of the variable in question.

8.2.3 Extreme indices

The extreme measures TX90p and R95p are taken from the dictionary of the European Climate Assessment and Dataset project (ECA&D), which has been commonly used in previous studies about climate extremes (Frich et al., 2002; Meehl and Tebaldi, 2004; Alexander et al., 2006). TX90p, a heat extreme measure, is defined as the number of days per season exceeding the 90th percentile of daily maximum temperature in the present day ensemble. R95p, an extreme precipitation measure, is defined as the number of days per season where precipitation exceeds the 95th percentile of daily precipitation on wet days ($>1\text{mm/day}$) in the present day ensemble. Precise definitions of the full list of indices are available at the ECA&D website (<http://eca.knmi.nl/indicesextremes/indicesdictionary.php>).

The percentile thresholds for computing TX90p and R95p are calculated from the present day ensemble. Due to the quantity of daily data, we have not applied the five day filtering window when calculating the percentile threshold for TX90p.

The simplified wet-bulb globe temperature (WBGT) (Fischer and Knutti, 2012) is given by:

$$\text{WBGT} = 0.567T + 0.393e + 3.94, \quad (8.1)$$

where T is the air temperature in degrees Celsius and e denotes the water vapour pressure in hPa. Water vapour pressure is calculated from relative humidity by $rH = \frac{e}{E} \times 100\%$. The saturation water pressure, E (in hPa), is approximated using the Magnus formula (Sonntag, 1990):

$$E(T) = 6.112 \times \exp\left(\frac{17.62T}{243.12 + T}\right). \quad (8.2)$$

We define the index WBGT95p for a particular month as the number of degree months per year at each grid point above the 95th percentile of WBGT in the same month of the present day ensemble.

8.2.4 Significance testing

We test for a discernible signal against the decadal variability in the model by assessing the difference between the means of the two sets of 10 year extreme index means for each ensemble member. To test for discernible changes between the high and low CO₂ runs, the t -statistic does not depend on ensemble size, and is given by $t = \frac{\mu_h - \mu_l}{\sqrt{\sigma_h^2 + \sigma_l^2}}$, where the μ_i are the ensemble means in the high and low CO₂ cases, and the σ_i the standard deviations of the 10-year ensemble member means. This tests for a decadal signal against the internal variability inherent in the climate system. We assess discernibility at the 10% significance level. We also test for significant changes in the extreme index distributions using a two-sample Kolmogorov-Smirnov test at the 10% significance level.

Table 8.1: Ensemble global mean temperature ($\bar{T}$) and precipitation ($\bar{R}$) change relative to the present day ensemble with 5-95% sampling uncertainty of 10-year ensemble means.

Ensemble	CO ₂ (ppm)	$\Delta\bar{T}$ (°C)	$\Delta\bar{T}_{land}$ (°C)	$\Delta\bar{R}$ (%)	$\Delta\bar{R}_{land}$ (%)
Present day	390.4	-	-	-	-
HadAM3P					
1.5° C low CO ₂	395.8	0.69 ^{+0.01} _{-0.02}	0.85 ^{+0.04} _{-0.05}	3.09 ^{+0.11} _{-0.10}	2.97 ^{+0.51} _{-0.75}
1.5° C best-estimate CO ₂	423.4	0.71 ^{+0.02} _{-0.02}	0.91 ^{+0.06} _{-0.05}	2.84 ^{+0.10} _{-0.10}	3.12 ^{+0.53} _{-0.69}
1.5° C high CO ₂	550.0	0.81 ^{+0.01} _{-0.01}	1.20 ^{+0.04} _{-0.05}	1.91 ^{+0.09} _{-0.08}	4.21 ^{+0.59} _{-0.57}
2.0° C best-estimate CO ₂	486.6	1.18 ^{+0.02} _{-0.02}	1.59 ^{+0.06} _{-0.06}	3.72 ^{+0.08} _{-0.10}	4.68 ^{+0.70} _{-0.66}
MIROC5					
1.5° C low CO ₂	395.8	0.66 ^{+0.02} _{-0.02}	0.85 ^{+0.06} _{-0.05}	1.82 ^{+0.10} _{-0.09}	1.09 ^{+0.55} _{-0.81}
1.5° C best-estimate CO ₂	423.4	0.68 ^{+0.02} _{-0.02}	0.90 ^{+0.06} _{-0.05}	1.59 ^{+0.16} _{-0.12}	1.29 ^{+0.49} _{-0.53}
1.5° C high CO ₂	550.0	0.77 ^{+0.02} _{-0.02}	1.11 ^{+0.05} _{-0.05}	0.75 ^{+0.13} _{-0.12}	1.77 ^{+0.70} _{-0.70}
2.0° C best-estimate CO ₂	486.6	1.15 ^{+0.03} _{-0.02}	1.52 ^{+0.07} _{-0.06}	2.25 ^{+0.11} _{-0.16}	2.05 ^{+0.68} _{-0.63}
CAM4					
1.5° C low CO ₂	379.0	0.62 ^{+0.02} _{-0.02}	0.68 ^{+0.07} _{-0.07}	2.42 ^{+0.11} _{-0.11}	0.65 ^{+0.65} _{-0.62}
1.5° C best-estimate CO ₂	423.4	0.66 ^{+0.02} _{-0.02}	0.78 ^{+0.06} _{-0.07}	2.02 ^{+0.11} _{-0.12}	1.00 ^{+0.59} _{-0.61}

8.3 Patterns of extremes

The differences in global mean temperature and precipitation between the five ensembles are presented in Table 8.1. Global mean temperature differences between the 1.5°C ensembles are small, as expected due to the prescribed SSTs, with only a 0.12°C difference between low and high CO₂ ensembles in HadAM3P and 0.11°C in MIROC5. Over land, these differences are slightly more pronounced. To correct for the differences in global mean temperature between low, best-estimate and high ensembles, we use a simple linear regression model (see Section 8.2 for details), fitting changes in variables to changes in global mean temperature and radiative forcing. Using the regression coefficients and global mean temperature difference between 1.5°C ensembles, we then adjust all variables to have a value associated with the global mean temperature in the best-estimate ensemble. This means all differences between the 1.5°C ensembles are

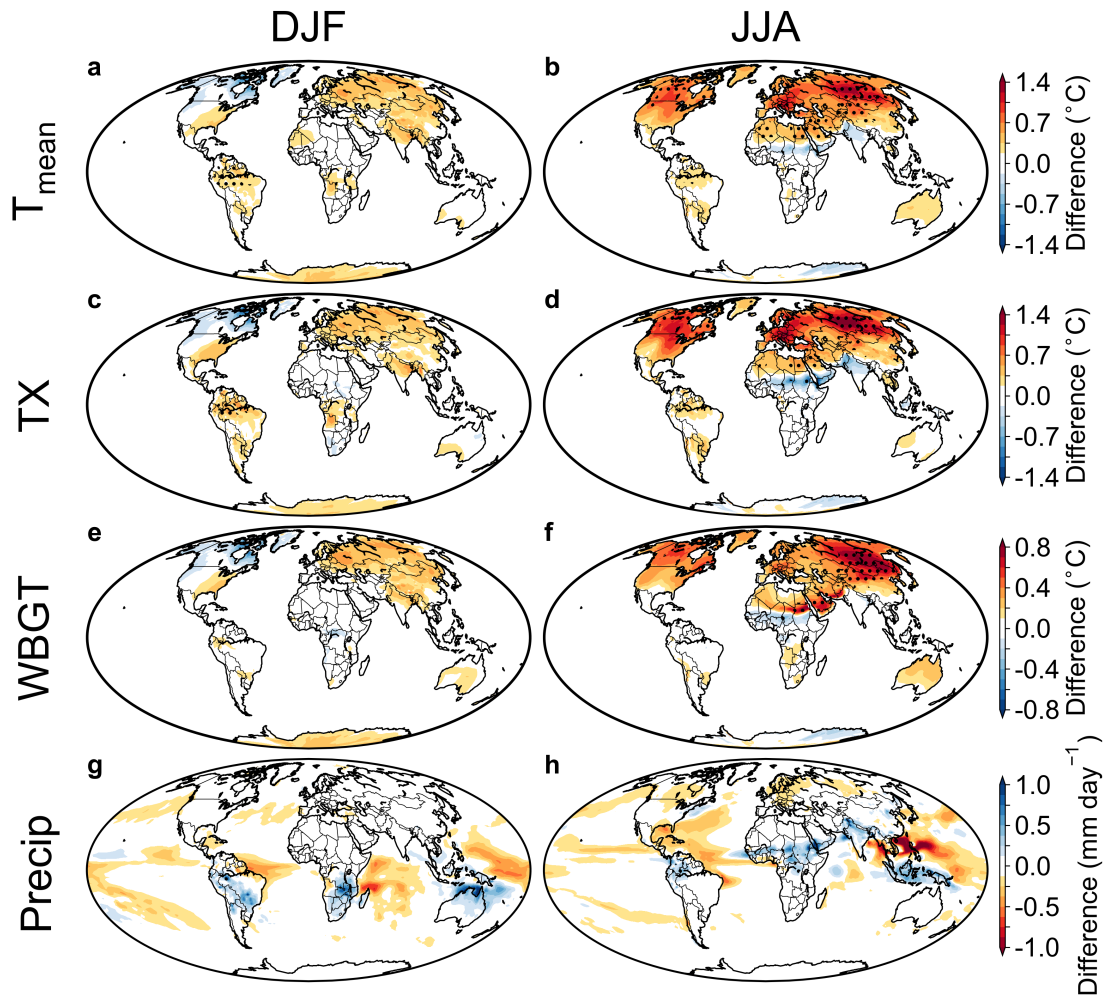


Figure 8.1: Changes associated with CO_2 concentration at 1.5°C warming in HadAM3P with temperature correction applied. Differences in T_{mean} (a, b), TX (mean daily maximum temperature) (c, d), WBGT (see text for details) (e, f) and precipitation (g, h) between high and low CO_2 forced ensembles. Stippling indicates statistical significance at the 10% level when testing for discernible impact (i.e. with an effective sample size of one, see Section 8.2 for details).

due to the direct radiative forcing effect from differing CO_2 concentrations. All variables and figures (unless specified) for the rest of this chapter use the corrected ensembles (i.e. with the effect of the global mean temperature differences between the 1.5°C ensembles removed). Uncorrected figures are shown in Appendix A. The spatial patterns of change in HadAM3P due to the direct radiative forcing effect are shown for several variables in Fig. 8.1 (and without temperature correction in Appendix A, Fig. A1). It may be argued that the SST pattern and CO_2 concentrations are physically inconsistent with each

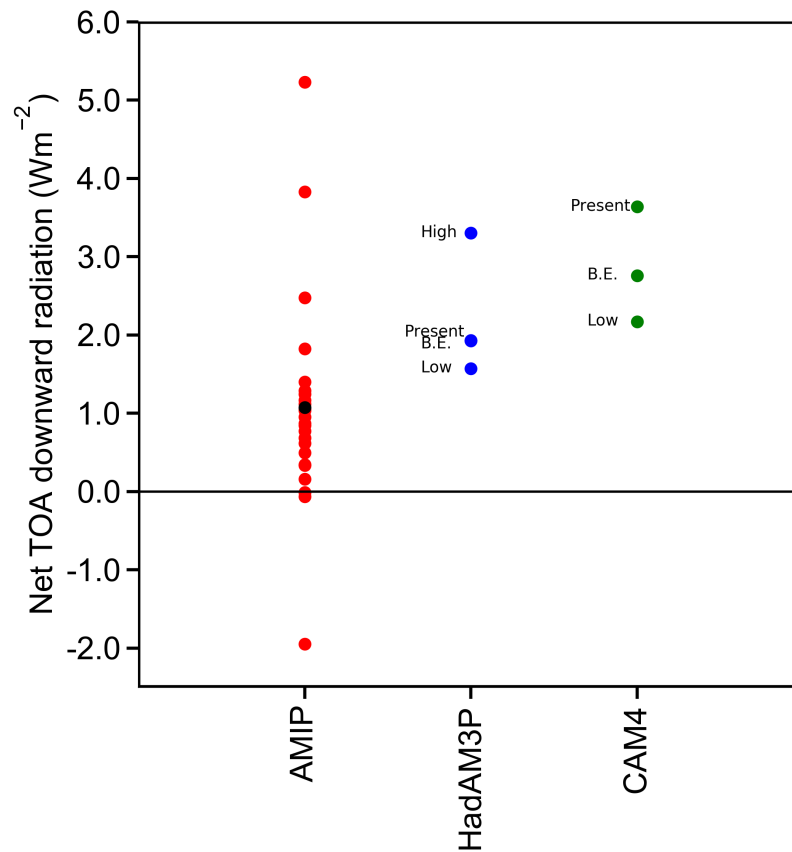


Figure 8.2: Net downward radiation at the top of the atmosphere. The red dots mark the individual historical AMIP simulations (1979-2008), with the multi-model mean shown in black. Coloured dots mark the HadAM3P (blue) and CAM4 (green) present, low, best-estimate and high ensembles.

other, but the simulated atmosphere is still physically consistent within the model with its SSTs and CO₂ boundary conditions. The top of atmosphere radiative imbalances in our ensembles lie within the range of imbalances in the historical AMIP simulations (see Fig. 8.2), and so any conclusions we draw possess the same caveats as conclusions drawn from using the AMIP simulations.

We now consider the differences in the extreme indices between the ensembles. Differences between the high and low CO₂ concentration ensembles are shown for HadAM3P, MIROC5 and CAM4 in Figs. 8.3, 8.4 and 8.5 respectively (and without temperature corrections in Figs. A2, A3, A4). The stippling shows differences discernible against the decadal variability in the model (see Section 8.2 for details). We test for statistical sig-

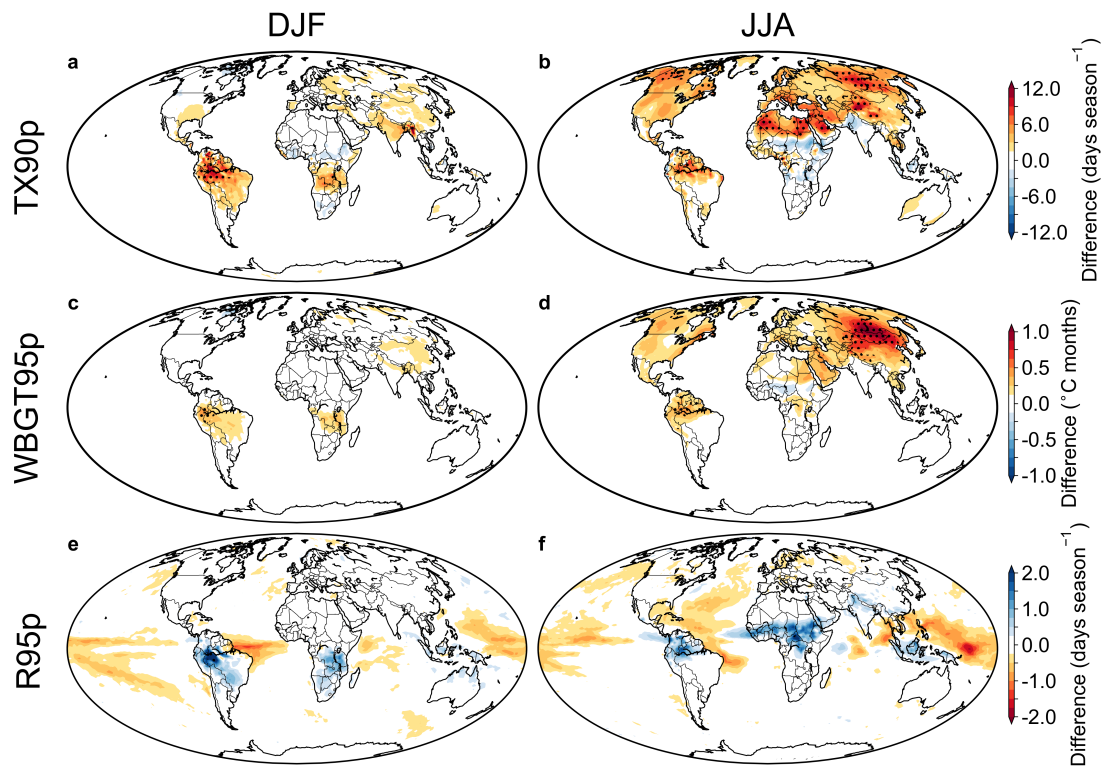


Figure 8.3: Changes associated with CO₂ concentration at 1.5°C warming in HadAM3P. Differences in TX90p (a, b), WBGT95p (c, d) and R95p (e, f) between high and low CO₂ forced ensembles. Stippling indicates statistical significance at the 10% level when testing for discernible impact (i.e. with an effective sample size of one, see Section 8.2 for details). Note the reversal of the colourbar for e,f.

nificance at the 10% level using a two-sample Kolmogorov-Smirnov test; the hatching in Fig. 8.6 indicates significant changes in extreme indices in HadAM3P (maps for the other ensembles are not shown, but display similar hatching regions). Due to large ensemble sizes, we find significant changes across large regions of the globe. There is considerable spread in extreme indices in the CMIP5 models at the same temperatures (Seneviratne et al., 2016). We sub-sample the CMIP5 ensembles, computing the JJA TX90p when each model reaches 1.5°C under the RCP8.5 scenario. The differences in the spatial patterns of TX90p between models with low and high global mean temperature sensitivity to carbon emissions (see Fig. 8.7), support our findings that increases in CO₂ concentrations lead to increases in extreme temperature.

The WBGT, a heat stress metric and proxy for human discomfort (Sherwood and Hu-

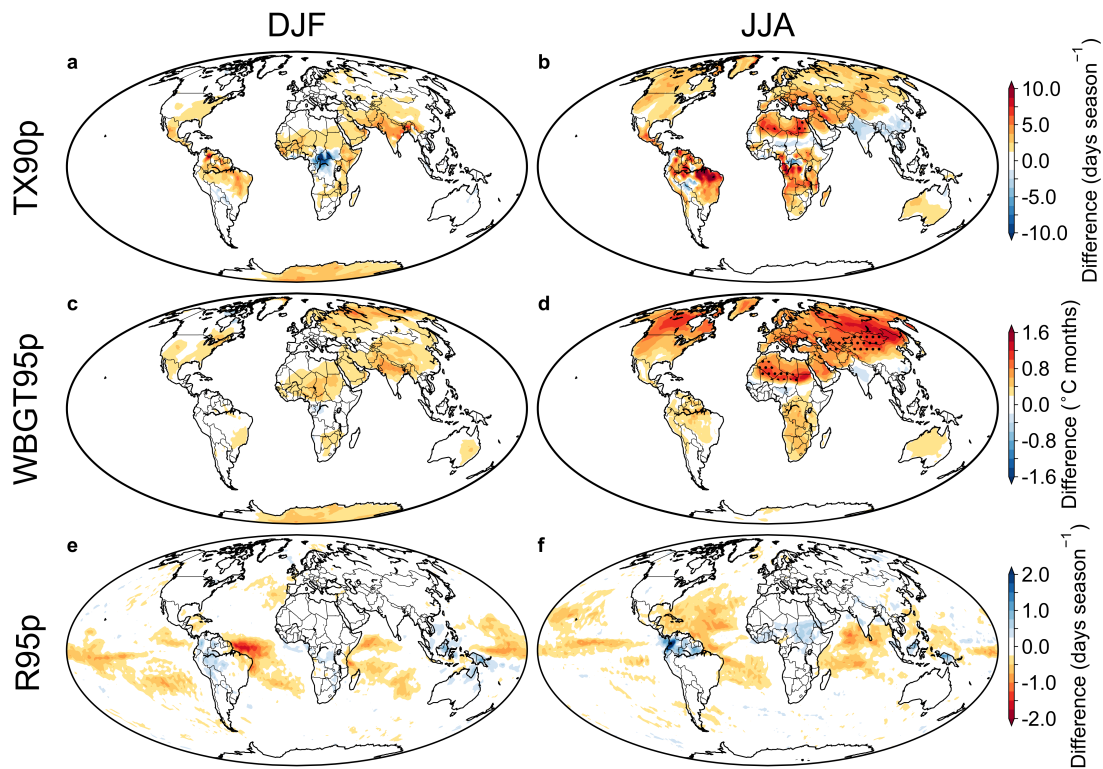


Figure 8.4: Changes associated with CO₂ concentration at 1.5°C warming in MIROC5. Differences in TX90p (a, b), WBGT95p (c, d) and R95p (e, f) between high and low CO₂ forced runs. Stippling indicates statistical significance at the 10% level when testing for discernible impact (i.e. with an effective sample size of one, see Section 8.2 for details).

ber, 2010; Willett and Sherwood, 2012), allows us to better quantify the impact of temperature extremes on humans. We consider changes in the extreme metric WBGT95p due to increases in CO₂. In JJA, increases are co-located with high population density areas in the Eastern US seaboard, central Europe, the Arabian Peninsula and North-East China and Korea. These regions are likely to experience increases in likelihood and severity of humid heatwaves under climate change (Russo et al., 2017). Our results suggest these increases may be partly due to differences in atmospheric composition. Increases in population over the coming decades, combined with rising CO₂ concentrations, mean more people will be exposed to the extreme changes in WBGT (Im et al., 2017) with potential economic impacts (Takakura et al., 2017) and potentially drastic societal implications (Murari et al., 2015).

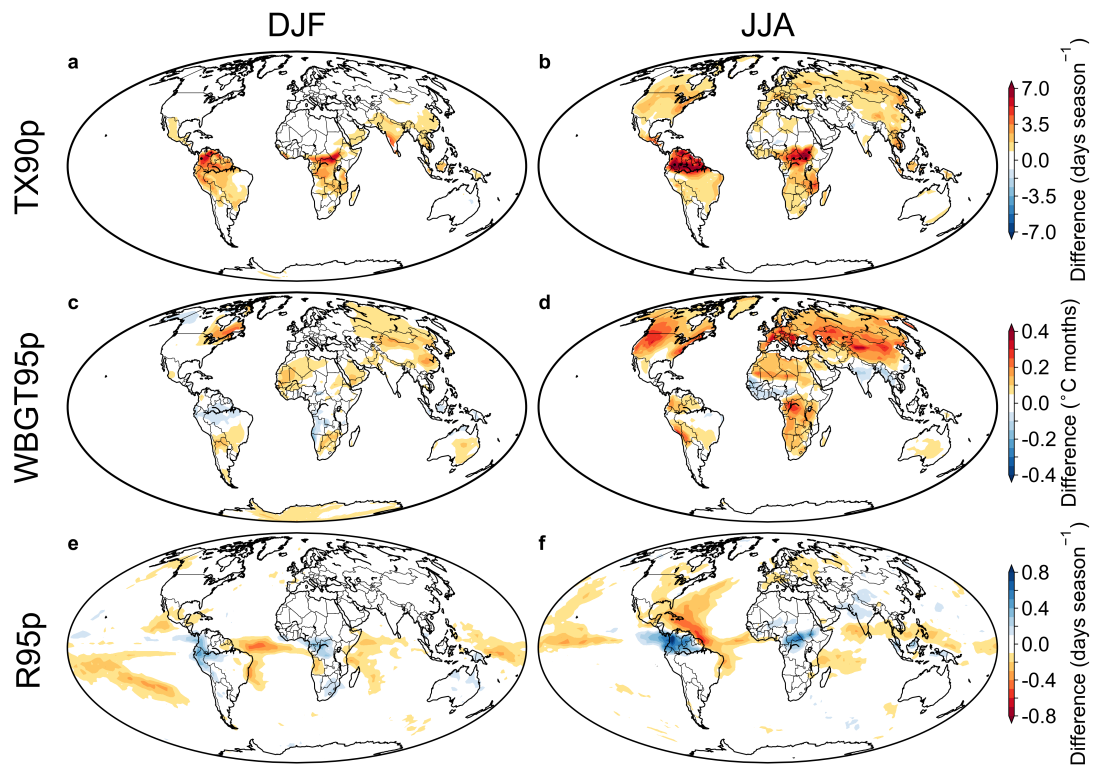


Figure 8.5: Changes associated with CO₂ concentration at 1.5°C warming in CAM4. Differences in TX90p (a, b), WBGT95p (c, d) and R95p (e, f) between best-estimate and low CO₂ forced runs. Stippling indicates statistical significance at the 10% level when testing for discernible impact (i.e. with an effective sample size of one, see Section 8.2 for details).

Area mean changes from present day are summarized in Fig. 8.8 (and without temperature correction in Fig. A5). Global mean changes in temperature and precipitation are shown, as well as land averaged Northern Hemisphere extratropical (NH_ET), 30°N-90°N, and tropical, 30°S-30°N, JJA changes for the mean and extreme indices. It is clear from the differences between high and low CO₂ ensembles that the area averaged changes in JJA means and extremes are highly significant in both the NH extratropics and tropics for WBGT95p and TX90p. R95p changes are significant in HadAM3P (excluding the extreme NH extratropical precipitation), but not in MIROC5.

We compare the difference in extreme indices due to the range in CO₂ concentrations consistent with 1.5°C warming scenarios with the difference between the best-estimate CO₂ concentration cases at 2.0°C and 1.5°C global mean warming. The CO₂

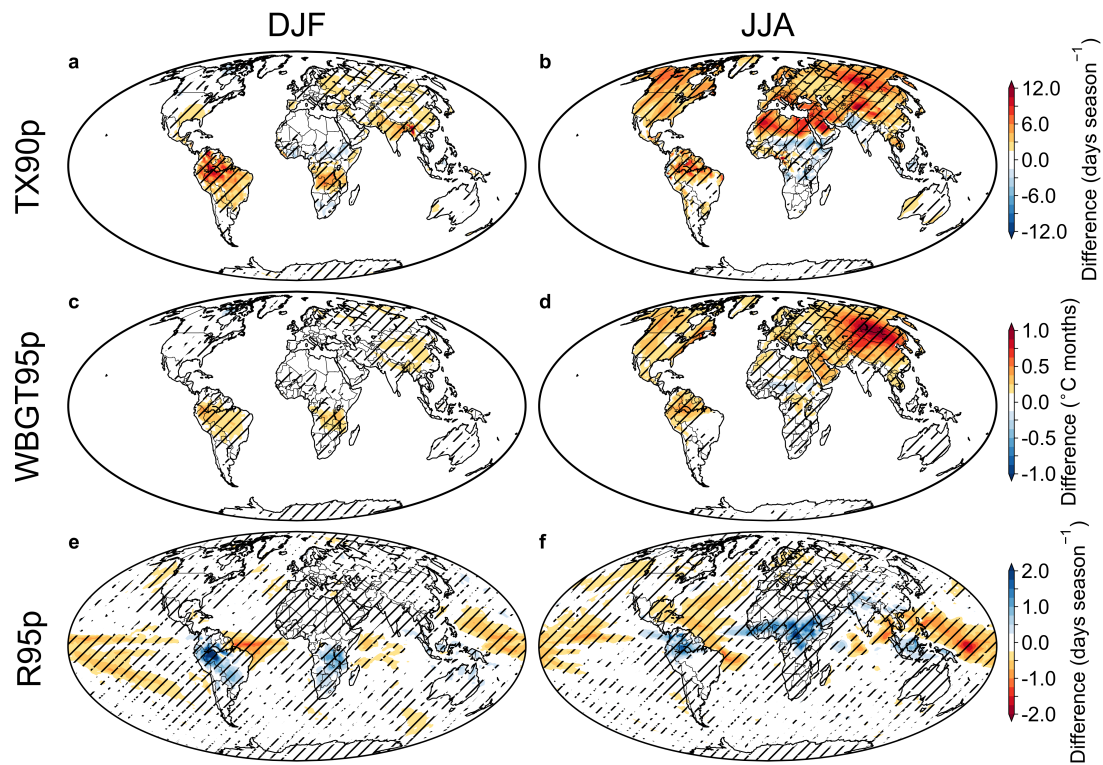


Figure 8.6: Changes associated with CO₂ concentration at 1.5°C warming in HadAM3P with temperature correction applied. Differences in TX90p (a, b), WBG95p (c, d) and R95p (e, f) between best-estimate and low CO₂ forced ensembles. Hatching indicates significance at the 10% level, computed using a two-sample Kolmogorov-Smirnov test.

induced differences as a percentage of the global mean warming induced differences for HadAM3P are shown for TX90p (Fig. 8.9a,b) and WBG95p (Fig. 8.9c,d) (and without temperature correction in Fig. A6). In JJA, over the NH midlatitudes, there are regions where the differences due to CO₂ are greater than those due to the extra 0.5°C of global mean warming. For R95p, we show the zonally averaged differences for ensemble means in the solid colours, and individual ensembles member differences in the light colours (Fig. 8.9e,f) (and without temperature correction in Fig. A6). In both DJF and JJA, the differences due to CO₂ in the tropics are on the same order of magnitude as the differences due to the extra 0.5°C of global mean warming. The importance of CO₂ on driving tropical extreme precipitation is contrary to the extratropics where global mean warming is the main driver (Pendergrass et al., 2015). The comparisons for MIROC5 (Fig. 8.10, and without temperature correction in Fig. A7) vary from HadAM3P, with

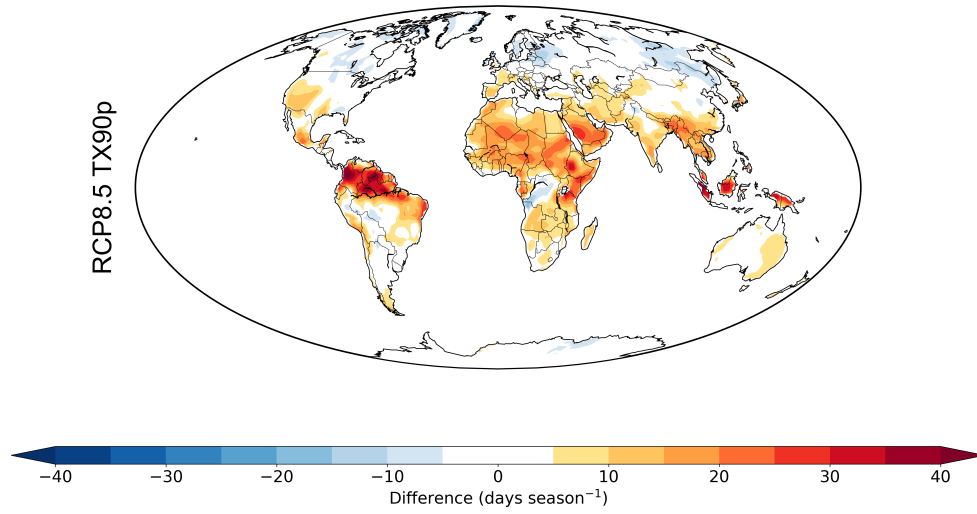


Figure 8.7: The difference in JJA TX90p at 1.5°C under RCP8.5 (averaged over 5 years each side of when the model reaches 1.5°C) from the mean of 3 low sensitivity CMIP5 models (Inmcm4, GFDL-ESM2M, HadGEM2-CC) to 3 high sensitivity CMIP5 models (CanESM2, MIROC-ESM, MPI-ESM-LR). Data from <http://climate-modelling.canada.ca/climatemodeldata/climindex/> (Sillmann et al., 2013a,b).

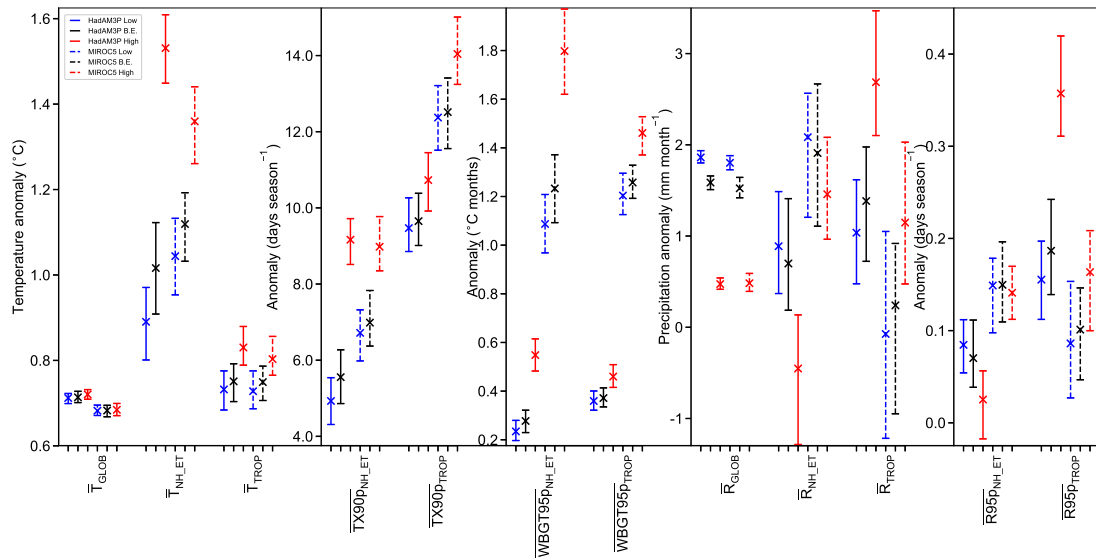


Figure 8.8: Area averaged differences associated with CO₂ concentration at 1.5°C warming with temperature correction applied. Whisker plots show the difference in indices between the low, best estimate and high CO₂ ensembles relative to the present day ensemble. Crosses mark the 50th percentile and the caps mark the 10th and 90th percentiles. Global mean temperature ($\bar{T}$) and precipitation ($\bar{R}$) differences (GLOB indices) are calculated as averages over all seasons and the whole globe. Precipitation anomalies are expressed as the mean change over the gridpoints.

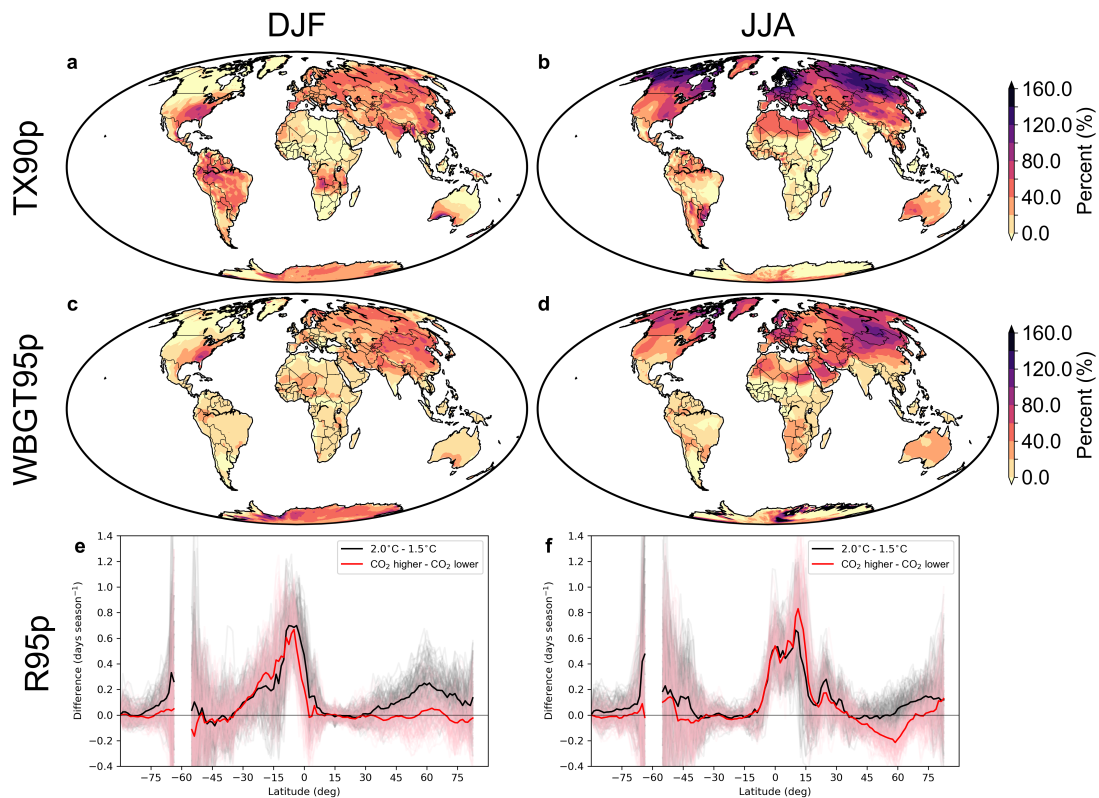


Figure 8.9: Comparison of changes due to CO₂ concentration with changes due to 2.0°C to 1.5°C warming difference in HadAM3P. Differences between high and low CO₂, with temperature correction applied, as a percentage of the difference between 2.0°C and 1.5°C ensembles for TX90p (a, b) and WBG95p (c, d). Differences between high and low CO₂ (red), and the differences between 2.0°C and 1.5°C (black) for R95p (e, f). Plots show the zonally averaged ensemble mean R95p changes in solid colours, and individual ensemble members in the light colours.

slightly lower magnitude, but nevertheless show substantial changes due to CO₂ concentration differences compared to differences due to the extra 0.5°C of global mean warming.

8.4 The carbon budget

These results highlight that increases in extreme indices due to the direct effect of CO₂ have important implications on the upper limit of CO₂ that can be emitted in order to limit changes in specific costly climate extremes. If we accept a level of extreme index increase consistent with 1.5°C of global mean warming associated with the increased

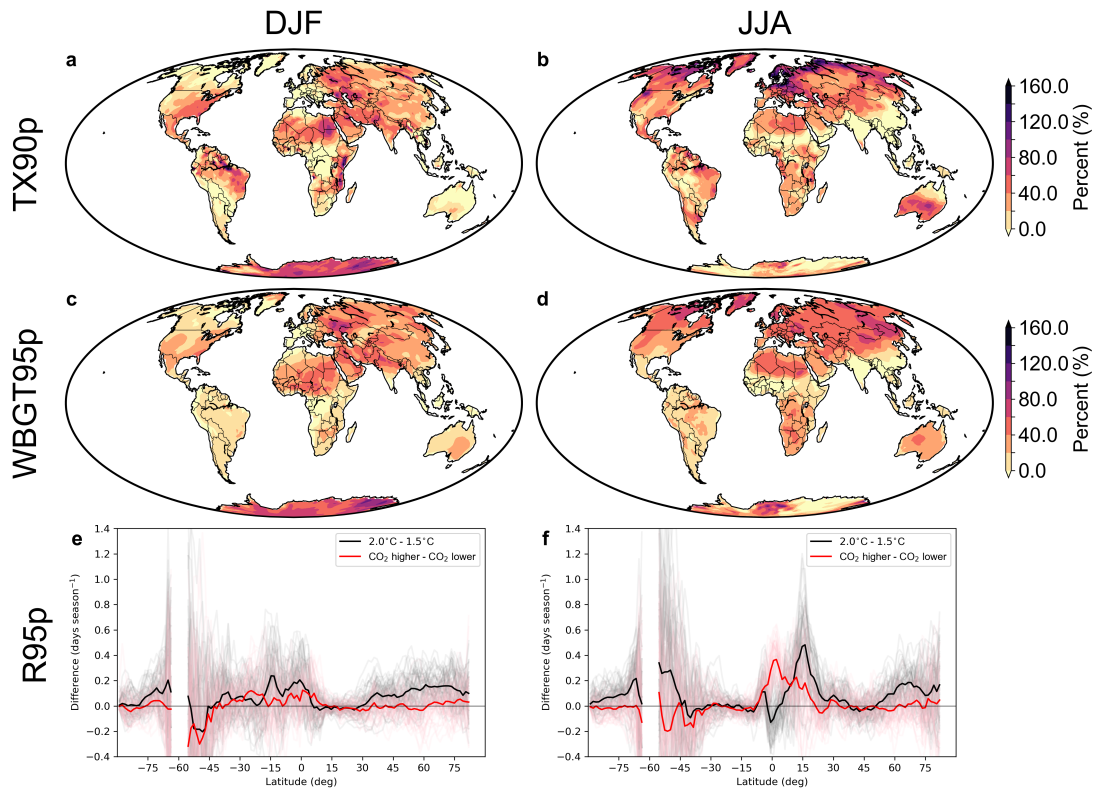


Figure 8.10: Comparison of changes due to CO₂ concentration with changes due to 2.0°C to 1.5°C warming difference in MIROC5. Differences between high and low CO₂ as a percentage of the difference between 2.0°C and 1.5°C ensembles for TX90p (a, b) and WBG95p (c, d). The differences between high and low CO₂ (red), and the differences between 2.0°C and 1.5°C (black) for R95p (e, f). Plots show the zonally averaged ensemble mean R95p changes in solid colours, and individual ensemble members in the light colours.

CO₂ in the best-estimate climate sensitivity case, we can ask what the upper bound of the carbon budget is that will keep extreme event likelihoods at this level. This will reduce the upper bound of the carbon budget, which is currently set by the uncertainty in the CO₂ emission levels which lead to 1.5°C of global mean warming. This allowable emissions uncertainty arises due to the global mean temperature response uncertainty (IPCC, 2013).

Using the simple regression model (see Section 8.2 for details), we calculate how each extreme index varies with global mean temperature and CO₂ concentration. Taking the extreme index value at 1.5°C of global mean warming in the best-estimate climate sensitivity case, we can then compute pairs of values of global mean temperature and

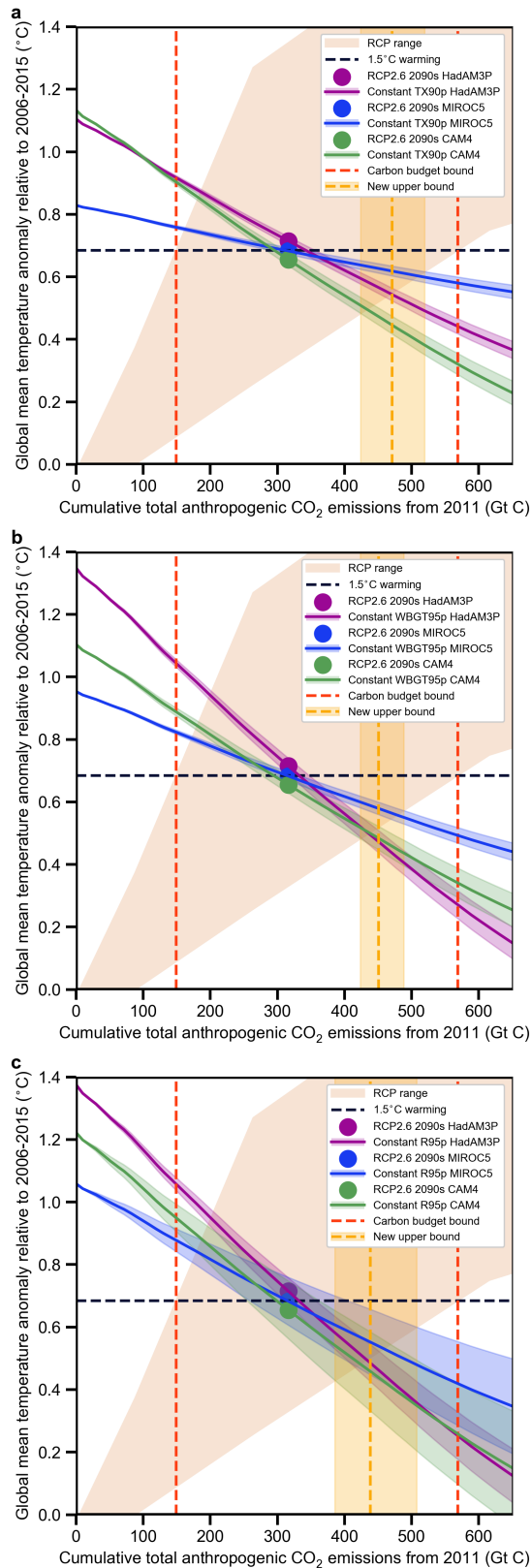


Figure 8.11: The reduction in the upper bound of the carbon budget. TX90p (JJA 30°N-90°N land) (a), WBGT95p (JJA 30°N-90°N land) (b) and R95p (JJA 30°S-30°N land) (c). The pink plume shows the 5-95 percentile of the CMIP5 response under all four RCPs relative to the 2006-2015 mean. The black dashed line marks the warming from the present decade equivalent to 1.5°C of mean global warming from the pre-industrial period (1850-1900). Dots indicate the warming observed in the best-estimate CO₂ concentration ensemble for each model (corresponding to the RCP2.6 2090-2099 mean). Curves show lines of constant extreme index, calculated from the regression model, with 5-95% uncertainty plumes. Red dashed lines mark the upper and lower uncertainty bounds of the carbon budget. The yellow dashed line is the new multi-model mean upper bound in the carbon budget, with uncertainty plume given by the pink plume's intersection with the model 5th percentile that gives the largest reduction in the carbon budget, to the intersection with the model 95th percentile that gives the smallest reduction in the carbon budget.

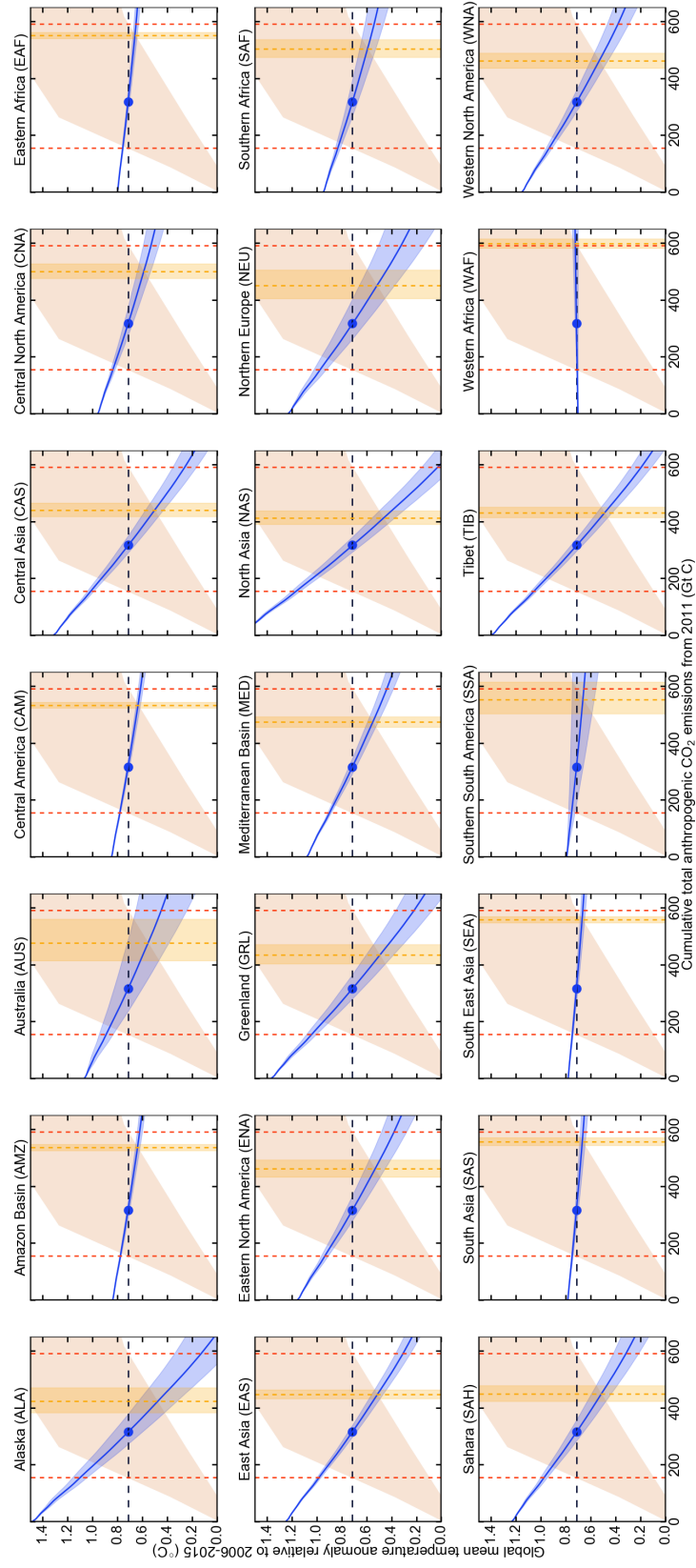


Figure 8.12: The reduction in the upper bound of the carbon budget for WBG95p in HadAM3P, shown for the Giorgi regions (Giorgi and Francisco, 2000).

CO₂ concentration resulting in that same extreme index value. This allows us to determine an upper bound for the carbon budget consistent with this extreme index change, as illustrated by Fig. 8.11 (see caption for details). For TX90p (Fig. 8.11a), the decrease in the upper bound of the carbon budget is from 569 to 471 GtC, giving a $23^{+11}_{-12}\%$ decrease in the uncertainty range of the carbon budget. For WBGT95p and R95p the new upper bounds are 450 GtC and 438 GtC, and reductions in uncertainty are $28^{+6}_{-9}\%$ and $31^{+13}_{-17}\%$ respectively (Figs. 8.11b,c). Reductions due to regional WBGT95p changes are shown in Fig. 8.12. Besides Western Africa, all regions display a decreased upper bound. We stress that these reductions in the carbon budget are only estimates based on the results from the three models. The use of a larger number of climate models run under a similar experiment, if they were available, would allow us to quantify the reduction more accurately. With the full range of CMIP5 models, there would be a greater spread in the uncertainty of the new upper bound of the carbon budget. However, based on the magnitude of the reduction in the upper bound of the carbon budget from these three models, it seems likely that the direct CO₂ effect is important, and should be taken into consideration when formulating carbon budgets to avoid any given level of climate impacts on extremes.

8.5 Policy implications

We demonstrate significant differences in temperature and precipitation extremes between the higher and lower likely CO₂ concentrations in a 1.5°C world. This highlights the importance of direct, local, CO₂ forcing effects on regional climates and extremes. We must acknowledge these effects so that we can avoid ‘dangerous’ changes in extremes which, in the context of the Paris Agreement, are defined as impacts beyond those expected at 1.5°C. The results also make clear that impacts on extremes in a world warming past 1.5°C would be different from those after stabilising at 1.5°C.

Differences in extremes at the same global mean temperature, due purely to differing

CO₂ concentrations, directly impact the use of the pattern scaling technique (Tebaldi and Arblaster, 2014). These differences provide compelling evidence that when using pattern scaling, we must account for CO₂ concentration and not just the changes in global mean temperature (which is the dominant method used in academic and policy work).

This chapter supports findings (Curry et al., 2014) that geoengineering schemes aimed at reducing the global warming impacts without reducing CO₂ concentration would not fully mitigate changes in extremes whose likelihoods have increased by the direct effect of increasing CO₂ concentrations.

Whilst it is important to note these findings are from only three models, the use of large ensembles and the statistical significance of the differences give weight to the reliability of our findings. Previous work (Richardson et al., 2016) shows the rapid regional precipitation response to increases in CO₂ is robust among CMIP5 models. Further work must be carried out to explore the spatial patterns and magnitudes of temperature and precipitation differences in other models to better quantify how extremes change with different atmospheric CO₂ concentrations. Despite this, it is clear that complementing global mean temperature goals with explicit limits on atmospheric CO₂ concentrations would reduce the risk of unexpectedly high changes in high-impact weather extremes.

Chapter 9

Conclusions

Uncertainty surrounding the circulation response to forcing is a key challenge in understanding how weather and climate will be impacted under climate change. In particular, the uncertainty in the exact nature of the forcing and the size and mechanisms of the response to different forcings provide a strong motivation for the in-depth study into the range of possible circulation responses due to the range of possible forcing natures. Therefore, as described in Chapter 1, this thesis aimed to use a systematic, sensitivity analysis based approach to understand how the circulation responds to localized forcings. The rationale for using this approach was to identify key forcing locations and to be able to investigate the mechanisms by which the circulation responds to localized forcing so that we could build upon current understanding of how the circulation might respond to climate change.

9.1 Findings

To help formalise our conclusions, we revisit our original main aims:

1. To quantify the sensitivity of the atmospheric circulation to forcing.
2. To investigate the seasonal dependence of the sensitivities.

3. To determine the dynamical mechanisms by which the atmospheric circulation responds to forcing.
4. To explore potential impacts of the range of circulation changes on temperature and precipitation extremes.

The first part of this thesis pursued these aims in a dry idealized GCM. We applied a series of localized thermal forcings systematically throughout the atmosphere and quantified the eddy-driven jet and Hadley cell sensitivities, as well studying the response mechanisms in selected case studies. In Chapter 3, we found that the speed and latitude of the jet display different sensitivities to each other in response to forcing, and these sensitivities also depend on the season. Latitude shift sensitivities are highest in the midlatitudes and polar regions near to the jet core, and these sensitivities move poleward in summer as the climatological jet in summer is further poleward. Jet speed sensitivities extend more prominently into the subtropics and tropics. The winter jet latitude is, however, affected by heating in the summer hemisphere. Consistent across both seasons and indices is the in-tandem response of the storm track with the jet, suggesting that the jet sensitivities are vital for our understanding of how midlatitude weather will be affected.

Seeking to explore the cross-hemispheric jet responses to forcing, in Chapter 4 we first quantified the sensitivities of the ITCZ and Hadley circulation. As with the jet, we found seasonally varying sensitivities which are largely barotropic in nature. Heating in one hemisphere affects the jet and Hadley circulation in that same hemisphere in a similar way, suggesting that the midlatitude eddies are feeding back onto the Hadley circulation. However, when we forced in the opposite hemisphere to the jet, we found that the Hadley circulation changes drive the jet changes. Winter hemisphere heating forces the summer Hadley cell extent, which is highly correlated with summer jet speed changes; summer hemisphere heating forces the winter Hadley cell extent which is highly correlated with winter jet shifts. We proposed that the jets respond to the Hadley circulation

through modifications of the overturning circulation projecting onto the heat fluxes and subtropical jets.

We then proceeded, in Chapter 5, to run a series of spin-up experiments to fully investigate the mechanisms responsible for the jet responses. We found that the main mechanism by which the jet responds to heating applied in the same hemisphere is: changes in heat fluxes and low level baroclinic stirring modify the eddy momentum fluxes which in turn cause the jet to respond as regions of momentum convergence and divergence shift and change in intensity. In practice, this means that shifts of the jet are driven by changes in the position of the maximum low level meridional temperature gradient which determines where the maximum meridional heat flux occurs. Changes in jet strength arise from changes in the low level meridional temperature gradient across the storm track, which dictates the amount of meridional heat flux.

The second part of this thesis went on to explore the circulation responses in more complex GCMs. The primary model used was HadAM3P, but we also compared results to MIROC5 and CAM4. The first experiments conducted with HadAM3P were sensitivity experiments to explore the North Atlantic circulation sensitivity to SST anomalies. Presented in Chapter 6, these experiments used a Green's function type approach to identify key regions that force the winter and summer NAO. We found that the Atlantic tripole, Indian Ocean and Pacific Ocean all contribute to the NAO state. We demonstrated skill in reconstructing both the winter and summer NAO from observed SSTs, and also some significant forecast skill using the SSTs from the month preceding the seasonal NAO we wanted to forecast. The skill arises from both the tropical and extratropical SSTs. The North Atlantic jet speed and latitude sensitivities are broadly similar to the NAO, albeit with some differences between all three sensitivity maps. It appears that fast timescale variations in the jet speed and latitude are important for reconstructing the NAO.

Comparing the idealized model and the HadAM3P results, it is clear they agree to a large extent, and this illustrates the robustness of the responses. Jet latitude sensitivities

in the Atlantic and Pacific agree with the idealized results: heating on the equator leads to an equatorward jet shift, whilst heating in the subtropics, up to the southern flank of the jet, shifts the jet poleward. Poleward of the jet, heating acts to shift the jet equatorward. In agreement with the idealized results, there is also a cross-hemispheric sensitivity observed in the Pacific and Atlantic. The results differ in the response of the jet to the Indian Ocean, and Western Pacific in summer, as these are longitudinally remote forcings from the jet, and the mechanism by which the jet responds is via Rossby waves, which are not captured in the idealized model.

Having investigated the circulation sensitivity to SSTs in the more complex GCM, we then proceeded, in Chapter 7, to investigate the circulation sensitivity to different types of forcing. We showed that there are directly opposing effects from SST and direct CO₂ forcing on tropical precipitation and on forced summertime Northern Hemisphere stationary waves. The phase of the wave depends on whether it was forced by the SSTs or the direct CO₂ forcing; the two forced waves are in antiphase. The wave, with wavenumber 5, is forced from the tropical Western and Eastern Atlantic, and in the Indian Ocean off the east coast of Africa. The forcing mechanism comes about due to changes in atmospheric static stability and land-sea temperature contrasts. The stationary wave has an impact on the regional surface temperature: where the anticyclonic wave centres are, the temperature has a positive anomaly, and where the negative centres are, there is a negative temperature anomaly.

A key difference across the hierarchy of models is the seasonality of the circulation response. In Chapter 3, we showed that the summer eddy-driven jet was more sensitive to forcing localized near the jet core, whilst the winter jet was sensitive to more broad ranging forcing. In Chapter 6, we showed that in the HadAM3P GCM, the winter NAO also displayed more sensitivity to remote forcings than the summer NAO. However, in Chapter 7, we showed that in summer, broad SST and GHG forcings led to a tropical forcing which produced a stationary wave circulation response in the Northern Hemisphere. These results suggest that the circulation response in winter is more governed

by a change in the zonal mean circulation, especially in response to remote forcings, whereas in summer, unless forcing is localized near the jet, the response of the circulation is dominated by wavelike changes. This wavelike response appears to arise due to the monsoon type effect in summer, where the land-sea contrast plays an important role in forcing the midlatitude circulation.

Focusing in on this summer stationary wave and associated temperature anomalies, in Chapter 8, we investigated the sensitivity of extreme heat and heat stress likelihood to CO₂ forcing. We also considered how the sensitivity of the tropical precipitation to CO₂ forcing affected precipitation extremes. We found significant differences in extreme event likelihoods between the higher and lower bounds of the carbon budget for 1.5°C for both summer Northern Hemisphere extreme temperatures and tropical extreme precipitation. The difference in extremes between the higher and lower carbon budget bounds was comparable to the rise in extremes between 1.5°C and 2.0°C of global mean temperature warming.

9.2 Implications and future work

9.2.1 Eddy-driven jet and NAO sensitivities

Changes in both the eddy-driven jet and Hadley circulation can impact the weather and climate experienced by a significant fraction of the global population. Understanding their response to climate change forcings, and other atmospheric forcings, is therefore vital to understand how the weather and climate may change. Our sensitivity studies highlight key regions where the atmospheric circulation is sensitive to forcing. In these key regions, it is important that we redouble efforts to reduce the uncertainty in the forcing amplitude and spatial distribution of the forcings. The idealized studies point to the need to reduce uncertainty in the spatial distribution of heating in the tropical upper troposphere, as the latitudinal extent of the heating is key to determine the sign and extent

of circulation changes, and the regions near the jet cores in the midlatitudes. Reducing the forcing uncertainty is not trivial; cloud feedbacks can act to alter diabatic heating in the atmosphere with an impact on the jets (e.g. Voigt and Shaw, 2015; Ceppi and Hartmann, 2016; Tandon and Cane, 2017), and on top of this, the extent of the heating and the circulation changes are inherently coupled. Although the model is a dry GCM, diabatic heating forcings could conceivably arise from changes in cloud cover. This highlights the importance of the sensitivity results presented in Chapter 3 and Chapter 4, as these cloud feedbacks may not necessarily manifest in the tropical upper troposphere and Arctic where the radiative and Arctic amplification warmings occur.

Our investigation into the mechanisms by which the eddy-driven jet is forced is conducted in the simple GCM with zonal mean forcing profiles. In reality, many atmospheric forcings, e.g. the aforementioned cloud changes or the Indian monsoon, have longitudinal structure and so to have confidence that the mechanisms we propose hold in a more realistic setting, a first step would be to perform spin-up studies similar to those presented in Chapter 5, but with longitudinally confined forcings. Whilst 80% of the storm track shifts seen in CMIP5 can be reproduced by dry dynamics (Mbengue and Schneider, 2013), further work using more complex GCMs with moist processes could verify the mechanisms we propose. Ultimately, an analysis of jet changes in an ensemble of coupled GCMs with explicitly designed experiments to test the forcing mechanisms (e.g. applying localized diabatic thermal forcings) would provide huge value. However, the cost and specific nature of this type of experiment render them untenable with current computing limitations.

In the more complex HadAM3P model, the key sensitivity regions were shown to be the tropical Indian and Pacific Oceans and the Atlantic tripole region. It is clear that in the model, the Indian Ocean SSTs force the NAO via a Rossby wave. However, when using the Indian Ocean SSTs to reconstruct the observed NAO, we observe no skill. Clearly, further work building on our work, and the Fletcher and Kushner (2011) and Fletcher and Cassou (2015) studies, is necessary to confirm the links between the Indian

Ocean and NAO. Because of the shortness of observational datasets, we suggest that a combined approach where coupled modelling studies and observations are utilised together would be a good next step to explore the linkage. The SST sensitivity study is only carried out with HadAM3P, and although Li and Forest (2014) also investigate the winter NAO sensitivity to SSTs in several other models, extending our study and investigating the differences between different models would allow the robustness of the sensitivities we observe to be explored.

Our sensitivity studies have been performed with diabatic heating applied to the atmosphere in the simple GCM and SST changes in the atmosphere-only GCM. Numerous other potentially important drivers of the jet and NAO exist. Some examples which have been shown to influence the jet include stratospheric dynamics and coupling to the troposphere (Polvani and Kushner, 2002; Gerber and Vallis, 2009; Son et al., 2009; Manzini et al., 2014) and sea ice processes (e.g. Pedersen et al., 2016). It is important to understand the sensitivities of the jet and NAO to these drivers, but also to include these processes in sensitivity studies. For example, SST perturbations at high latitudes will impact sea ice distributions; some mechanisms may also propagate via the stratosphere (e.g. Wang et al., 2012; Zhang et al., 2017), or result from heatings in the stratosphere (e.g. Chan and Plumb, 2009; Simpson et al., 2009).

Using the SST sensitivities, we demonstrated statistically significant skill in forecasting the NAO in certain time periods. In winter, there is some skill in recent decades using November SSTs to forecast the NAO, whilst in summer we found there was skill in the 1925-1994 period, using the May SSTs to forecast the summer NAO, which to this date has not been reported elsewhere (Dunstone et al., 2018). Improvements to forecasting seasonal weather patterns in advance generates great benefit to society as it allows prior planning for the risk of severe weather (Scaife et al., 2014). Clearly, our results are therefore important, but further work understanding the non-stationarity in the forecast skill for both seasons is vitally important so that we can use these types of forecast reliably. One avenue of investigation is to use long GCM control runs and isolate the

periods where there is skill in reconstructing the NAO from the GTO. This would allow us to search for common atmospheric states in the periods with skill versus without skill.

9.2.2 Sensitivity to forcing type

We now focus on the sensitivity of the summer circulation to direct radiative and SST forcings. We proposed a tropical forcing mechanism for the stationary wave observed in response to direct CO₂ and SST forcing, identifying the land-sea contrast and static stability in the tropics as potential mechanisms explaining why the key forcing locations appear to be located in coastal tropical regions. We identified these regions with the help of a set of barotropic sensitivity simulations. An alternative mechanism for causing stationary wave changes, in winter at least, is via changes in the zonal wind basic state (Simpson et al., 2016). We tested this mechanism in our barotropic experiments which suggested that it was not important. An obvious extension to our work with the barotropic model is to use a three-dimensional baroclinic model. This would allow a set of clean experiments to be carried out, where we could test the tropical forcing hypothesis and also test the importance of the background state.

The stationary wave, and the dependence of its phase due to whether it is forced by direct CO₂ forcing or SST forcing, have important implications for future changes in regional climate extremes. The wave is associated with anomalous surface warming under anticyclonic regions and anomalous cooling under cyclonic regions. These anomalously warm regions also exhibit increases in extreme temperature likelihood. However, we show which sign of the wave dominates the climate change response, and hence where the regions of anomalous extremes occur, depends on the Earth's climate sensitivity and land-sea temperature contrast response. Therefore, it is important that further work continues to reduce the uncertainty in the Earth's climate sensitivity and land-sea temperature contrast response to better constrain our understanding of how patterns of extremes will change in a changing climate.

In reality, the opposing effects of direct CO₂ and SST forcing are connected. There are two important aspects to understand in relation to this. The first is how the two effects would combine in a coupled ocean-atmosphere system. Previous work (Deser and Phillips, 2009; Lambert et al., 2011) suggests that the two effects may combine linearly. The other aspect is quantifying the dependence of the stationary wave response on the climate sensitivity and land-sea temperature contrast. An ideal set of experiments to test both of these aspects is the third phase of the Cloud Feedback Model Intercomparison Project (Webb et al., 2017). As well as the experiments consisting of atmosphere only runs with increased CO₂ and runs with uniform SST increases and SST pattern changes (allowing the importance of SST pattern changes in forcing the wave to be investigated), there are also coupled runs. Therefore the experiments will allow a more quantitative analysis of the opposing effects of radiative forcing and sea surface warming in summer.

In Chapter 8, we investigated the implications of the changes in extremes due to the stationary wave on climate policy and the range of the carbon budget for 1.5°C. The direct effect of CO₂ on extremes has several implications. Firstly, the upper bound of the carbon budget is reduced to avoid ‘dangerous’ changes in extremes. Secondly, impacts on extremes in a world warming past 1.5°C would be different from those after stabilising at 1.5°C. Thirdly, when using pattern scaling, one must account for CO₂ concentration and not just the changes in global mean temperature (which is the dominant method used in academic and policy work). Finally, geoengineering schemes aimed at reducing the global warming impacts without reducing CO₂ concentration would not fully mitigate changes in extremes whose likelihoods have increased by the direct effect of increasing CO₂ concentrations.

Due to the use of a limited sample of models compared to those used to compute the range of the carbon budget, it is necessary to carry out further, similar, modelling studies with a greater range of models to quantify accurately the reduction in the upper bound of the carbon budget, and more broadly to quantify the exact size of the direct effect of CO₂ on extremes. Another approach to quantifying the effect is to run perturbed physics

ensembles for a particular model, choosing parameters to give a range of model climate sensitivities. It would then be possible to quantify the effect of climate sensitivity and direct CO₂ effect on extremes at a specific temperature target. Once the effect has been accurately quantified, it would also then be possible to account for it when pattern scaling and calculating the range of the carbon budget.

9.3 Final remarks

This thesis has, through the use of a hierarchy of models, explored and quantified the sensitivity of the atmospheric circulation to different types of forcing. We investigated the seasonal dependencies of the circulation sensitivity, and showed how these seasonal circulation sensitivities could lead to seasonally varying climate change impacts, which could be pertinent to society. By conducting spin-up experiments, we formulated a mechanistic understanding of how the jet and Hadley circulation respond to forcing. We also put forward a mechanism by which SST and direct CO₂ forcing can force Northern Hemisphere stationary waves, which could have important societal impacts due to increased temperature extreme risk. Whilst further work would confirm the findings here, this thesis has mapped out a broad range of atmospheric circulation sensitivities and response mechanisms which are of tangible benefit to understanding the circulation response to climate change and other types of atmospheric forcing.

Appendix A

Supplemental figures for Chapter 8

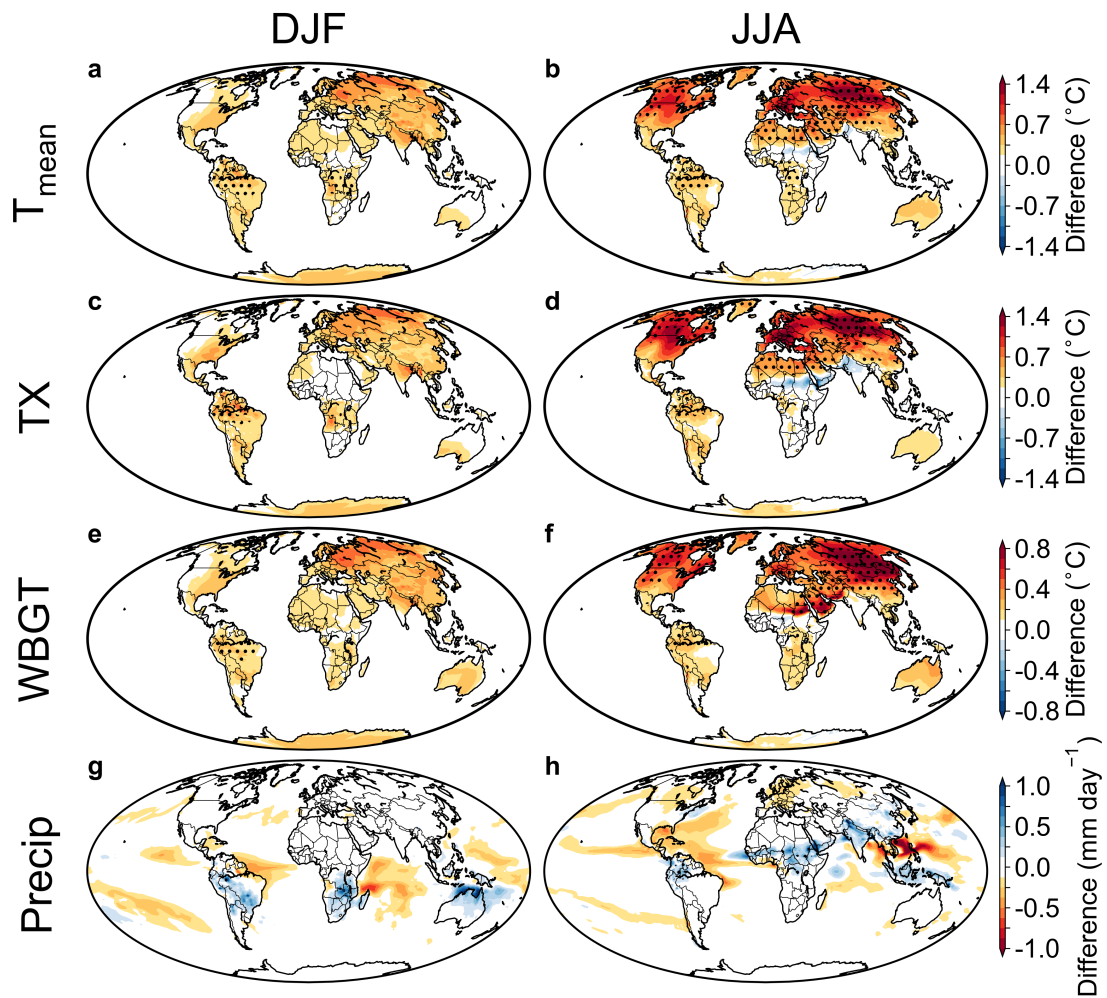


Figure A1: Changes associated with CO₂ concentration at 1.5°C warming in HadAM3P without temperature correction applied. Differences in T_{mean} (a, b), TX (mean daily maximum temperature) (c, d), WBGT (e, f) and precipitation (g, h) between high and low CO₂ forced ensembles. Stippling indicates statistical significance at the 10% level when testing for discernible impact (i.e. with an effective sample size of one, see Section 8.2 for details).

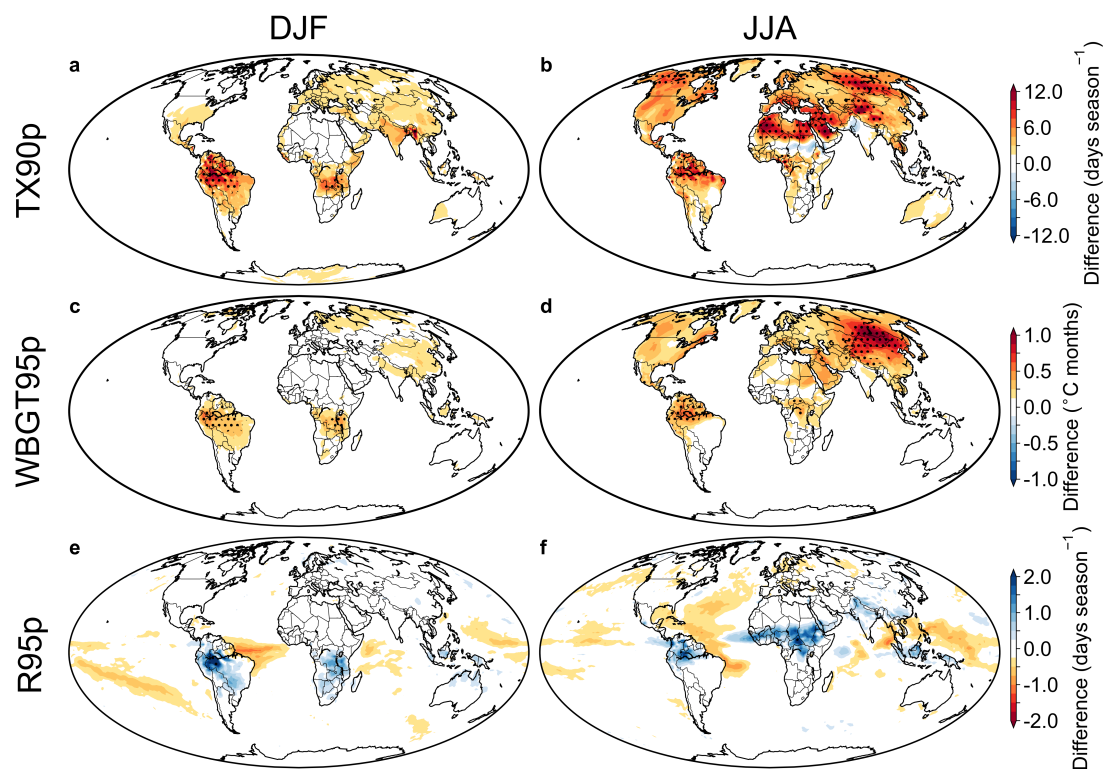


Figure A2: Changes associated with CO₂ concentration at 1.5°C warming in HadAM3P without temperature correction applied. Differences in TX90p (a, b), WBGT95p (c, d) and R95p (e, f) between high and low CO₂ forced ensembles. Stippling indicates statistical significance at the 10% level when testing for discernible impact (i.e. with an effective sample size of one, see Section 8.2 for details).

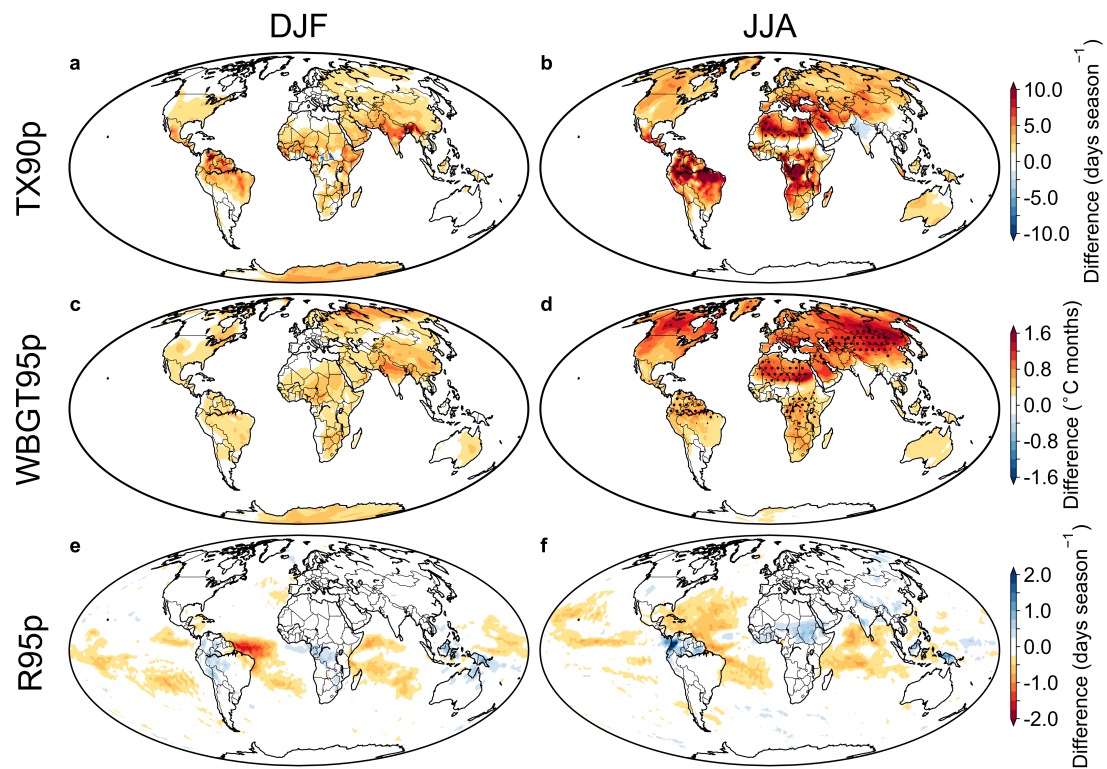


Figure A3: Changes associated with CO₂ concentration at 1.5°C warming in MIROC5 without temperature correction applied. Differences in TX90p (a, b), WBGT95p (c, d) and R95p (e, f) between high and low CO₂ forced ensembles. Stippling indicates statistical significance at the 10% level when testing for discernible impact (i.e. with an effective sample size of one, see Section 8.2 for details).

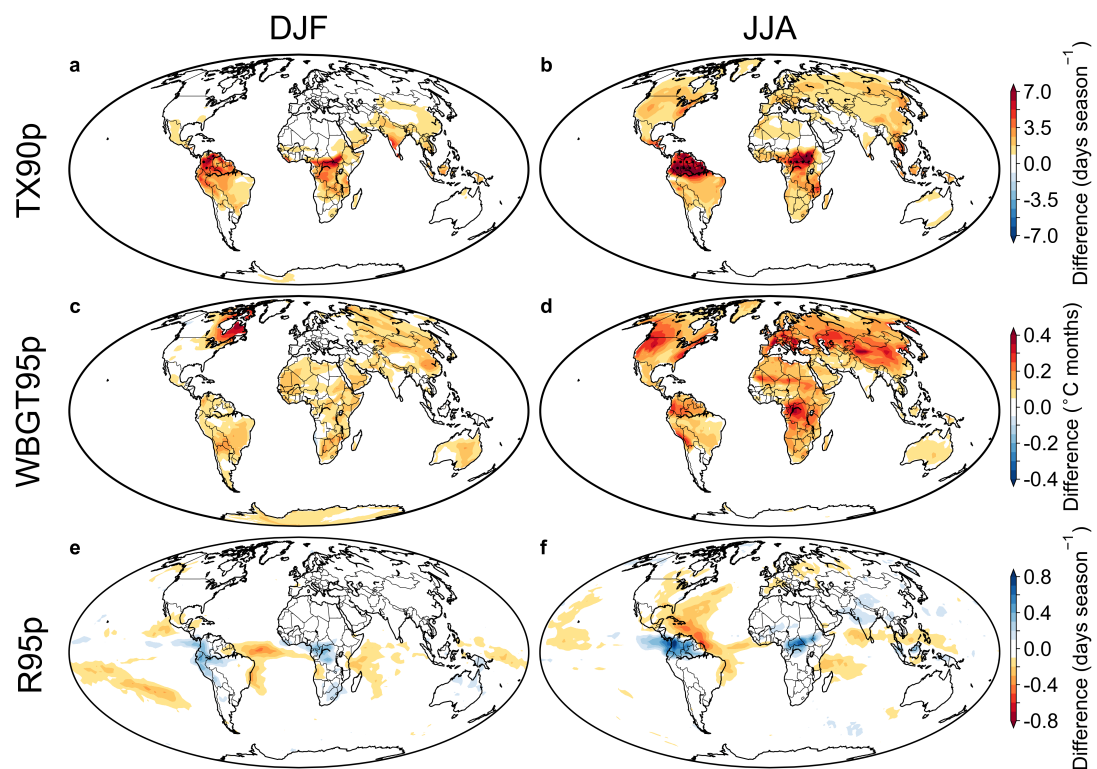


Figure A4: Changes associated with CO₂ concentration at 1.5°C warming in CAM4 without temperature correction applied. Differences in TX90p (a, b), WBGT95p (c, d) and R95p (e, f) between best-estimate and low CO₂ forced ensembles. Stippling indicates statistical significance at the 10% level when testing for discernible impact (i.e. with an effective sample size of one, see Section 8.2 for details).

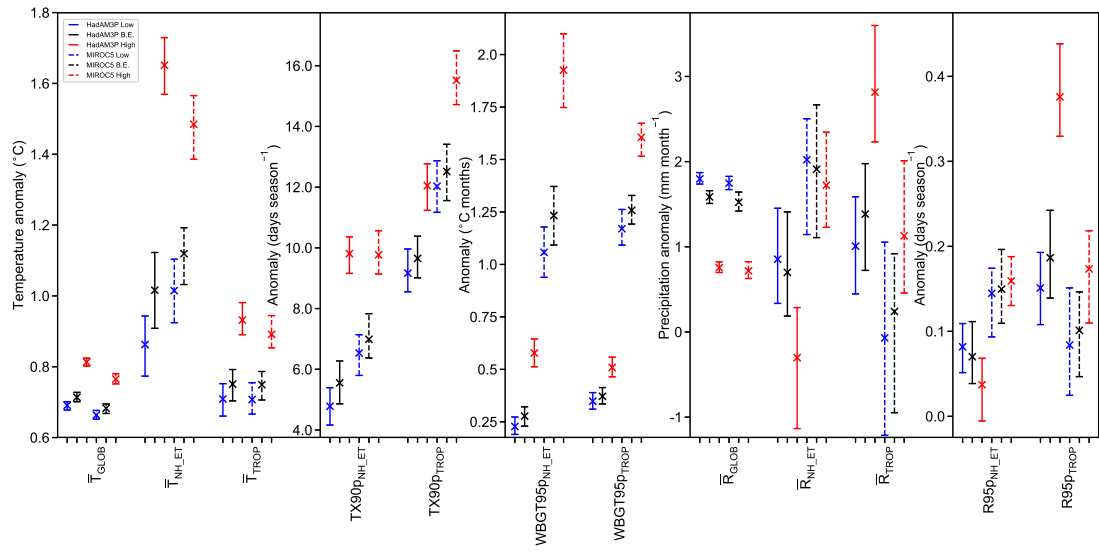


Figure A5: Area averaged differences associated with CO_2 concentration at 1.5°C warming without temperature correction applied. Whisker plots show the difference in indices between the low, best estimate and high CO_2 ensembles relative to the present day ensemble. Crosses mark the 50th percentile and the caps mark the 10th and 90th percentiles. Global mean temperature ($\bar{T}$) and precipitation ($\bar{R}$) differences (GLOB indices) are calculated as averages over all seasons and the whole globe. Precipitation anomalies are expressed as the mean change over the gridpoints.

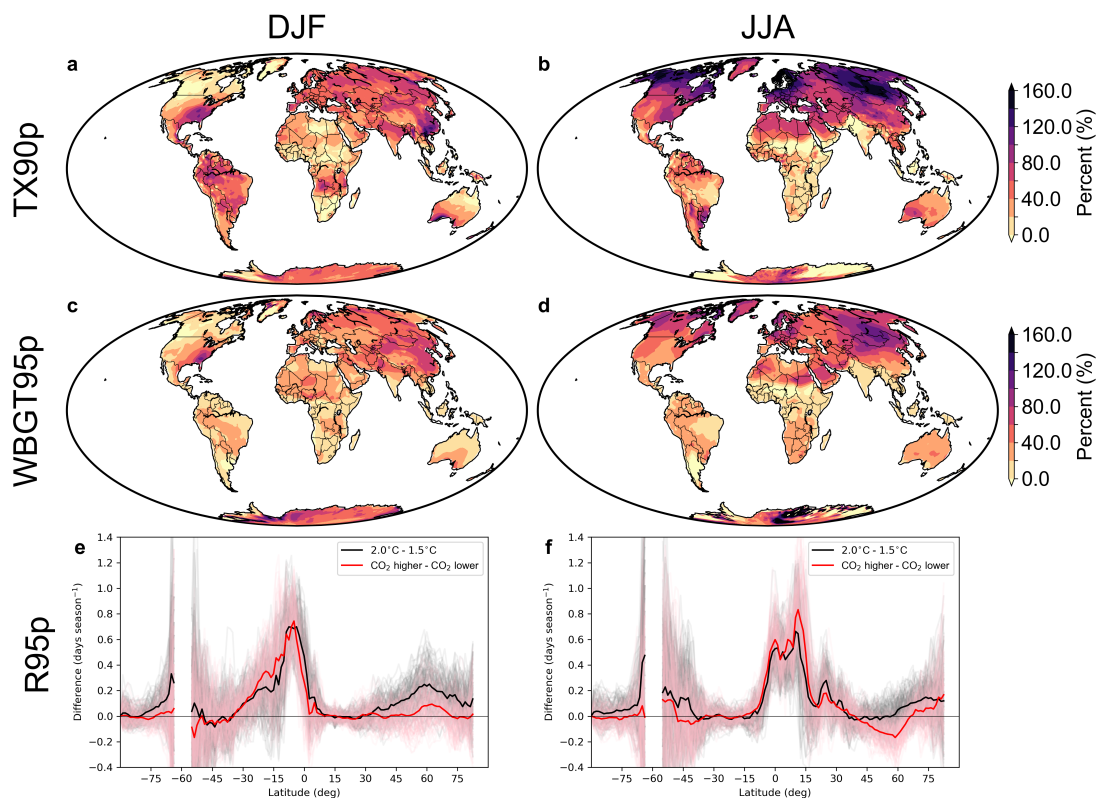


Figure A6: Comparison of changes due to CO₂ concentration with changes due to 2.0°C to 1.5°C warming difference in HadAM3P. Differences between high and low CO₂, without temperature correction applied, as a percentage of the difference between 2.0°C and 1.5°C ensembles for TX90p (a, b) and WBG95p (c, d). The differences between high and low CO₂ (red), and the differences between 2.0°C and 1.5°C (black) for R95p (e, f). Plots show the zonally averaged ensemble mean R95p changes in solid colours, and individual ensemble members in the light colours.

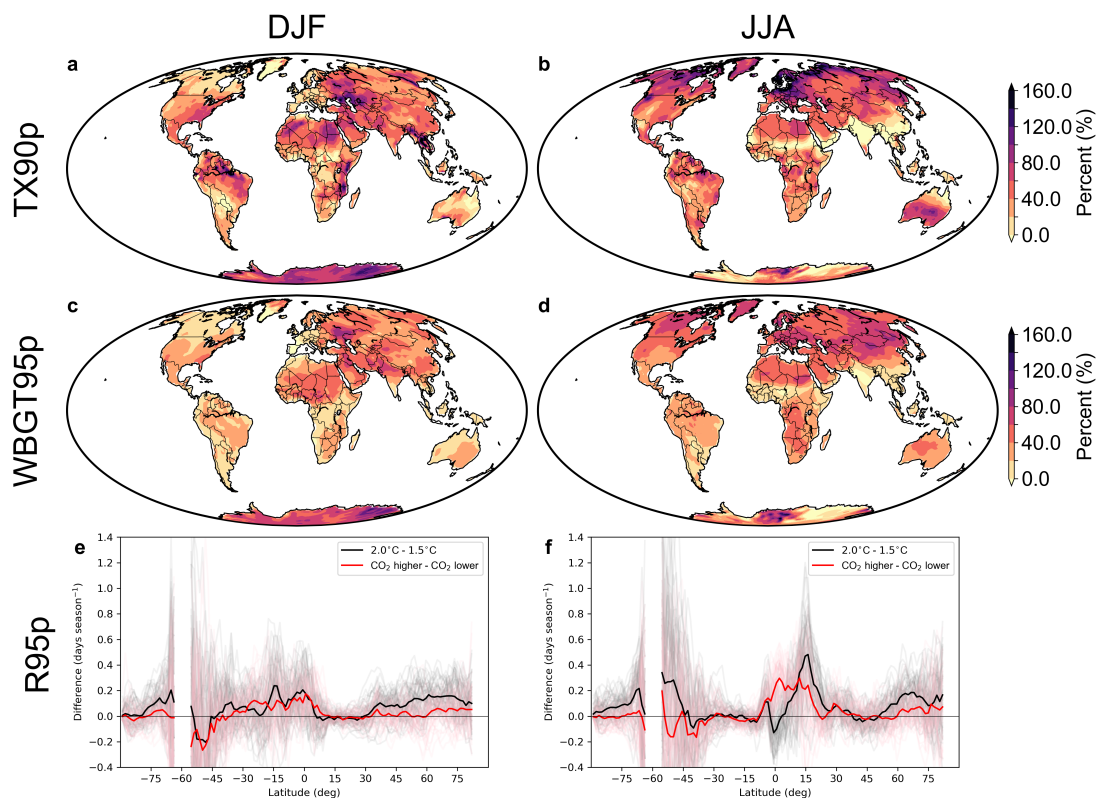


Figure A7: Comparison of changes due to CO₂ concentration with changes due to 2.0°C to 1.5°C warming difference in MIROC5. Differences between high and low CO₂, without temperature correction, applied as a percentage of the difference between 2.0°C and 1.5°C ensembles for TX90p (a, b) and WBG95p (c, d). The differences between high and low CO₂ (red), and the differences between 2.0°C and 1.5°C (black) for R95p (e, f). Plots show the zonally averaged ensemble mean R95p changes in solid colours, and individual ensemble members in the light colours.

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