The Large-Scale Atmospheric Circulation Response to Climate Change Drivers: A Multi-Model Comparison Study

by

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Submitted in accordance with the requirements for the degree of

Doctor of Philosophy

The University of Leeds

School of Earth and Environment

January 2022

UNIVERSITY OF LEEDS

Declaration of Authorship

The candidate confirms that the work submitted is their own, except where work which has formed part of jointly-authored publications has been included. The contribution of the candidate and the other authors to this work is explicitly indicated below. The candidate confirms that appropriate credit has been given within the thesis where reference has been made to the work of others.

Two-thirds of the research included in this thesis has been peer reviewed and published. The remaining third has been prepared for submission to a suitable journal. This thesis is published using the University of Leeds alternative thesis format. The research within can be readily identified and accessed. The thesis consists of an introductory chapter, three first-author articles, a conclusions and recommendations chapter and appendices containing supplementary information for the results chapters.

The publication **Wood et al. (2020a) The Southern Hemisphere Midlatitude Circulation Response to Rapid Adjustments and Sea Surface Temperature Driven Feedbacks**, Journal of Climate, 33(22), doi: 10.1175/JCLI-D-19-1015.1, jointly authored with A. C. Maycock, P. M. Forster, T. B. Richardson, T. Andrews, O. Boucher, G. Myhre, B. H. Samset, A. Kirkevåg, J.-F. Lamarque, J. Mülmenstädt, D. Olivié, T. Takemura and D. Watson-Parris, is included as Chapter 2 of this thesis. The text was solely written by the candidate, with comments from co-authors. The candidate performed data analysis and produced all figures. A. C. Maycock contributed to interpretation of results. Model simulations were performed by various members of the Precipitation Driver and Response Intercomparison Project (PDRMIP) group.

The publication **Wood et al. (2020b) Role of sea surface temperature patterns for the Southern Hemisphere jet stream response to CO_2 forcing**, Environmental Research Letters, 16(1), doi: 10.1088/1748-9326/abce27, jointly authored with C. M. McKenna, A. Chrysanthou and A. C. Maycock, is included as Chapter 3 of this thesis. The text was solely written by the candidate with advice from co-authors except for the 5th paragraph of the methods which was jointly written with C. M. McKenna. The candidate performed data analysis and produced all figures. IGCM4 experiments were conceived and designed by the candidate and A. C. Maycock. The IGCM4 model runs were performed by C. M. McKenna under direction of the candidate. All co-authors contributed to interpretation of results.

The publication **Wood and Maycock (2022) Aerosol-forced tropical expansion in the late 20th century: robust anthropogenic signal or internal variability?**, prepared for submission to Environmental Research Letters, jointly authored with A. C. Maycock, is included as Chapter 4 of this thesis. The text was solely written by the candidate with comments from A. C. Maycock. The candidate performed data analysis and produced all figures. A. C. Maycock contributed to the interpretation of results.

For Marley and Rudy

Acknowledgements

When I started this project in January 2017, global atmospheric CO_2 concentrations (measured at Mauna Loa, Hawaii) were approximately 405 ppm. Much has changed since then. I have become a father to two children, a pandemic has swept the globe, creating a step change in the way that society is organised and how we live our lives, and at the time of writing, almost exactly 5 years later, CO_2 concentrations have risen to approximately 417 ppm - an increase of about 3%. Climate change is the most pressing issue facing Earth, affecting everything from politics, economics and culture, to biodiversity, weather systems and sea levels. Its impact on the world over the next several decades has the potential to be catastrophic. It is arguably the greatest existential threat that has ever faced civilisation. I embarked on this project because I wanted to be part of the global effort to do something about it. With this contribution to scientific knowledge, I hope that I have taken a step toward achieved that aim and that when my children read this in the years to come, they can say that Dad did his best.

I would firstly like to thank my primary supervisor, Amanda Maycock. Her support throughout this project has been incredible. Her kind and understanding nature has helped immensely when I have faced issues both with the project itself and in my personal life. Having children and being simultaneously hit by a pandemic (and the restrictions to life that came with it), all while undertaking PhD research, has meant this has almost certainly been the most challenging period of my life. Amanda has been thoroughly understanding of this and has always found time from her extremely busy schedule to meet with me (with our meetings invariably lasting longer than planned!) and support my research, and has enabled me to work in the way that suited me best.

I also thank Piers Forster for his support and encouragement when I applied for this project and for his input in the early stages. I approached Piers following a public outreach meeting held at the University of Leeds discussing the 2015 IPCC Paris climate accord, when I was first considering applying for this project. His encouragement and enthusiasm was a major factor in my decision to leave my previous career in energy and emissions consultancy and return to academia.

I also thank my office colleagues who have helped with this project. In particular, Richard Rigby for his technical support, Chris Smith for sharing his code in the early stages, and Christine

McKenna and Andreas Chrysanthou for interesting and helpful scientific discussions and feedback.

Without the support of my family, none of this would be possible. My parents, Tina and Harry, have supported me financially, without which the unfunded paternity breaks I have taken would not have been possible. My parents-in-law, Jen and Rob, have also been incredibly supportive, helping with childcare on numerous occasions when I have been snowed under with work. My siblings, Tim, Kate, Russ and George have all helped me along the way, either directly with the project, or indirectly by being generally supportive and providing distractions when I needed them (and sometimes when I didn't!). Tim, a computer scientist, answered coding questions which helped me get to grips with a discipline that, up to the point of starting the project, I hadn't had much contact with. George, a sociologist, has helped enormously with the statistical side of this work, discussing various methodologies and answering questions at all times of the day.

A huge thank you is owed to my best friend, Ad. His moral support has been nothing short of heroic at times. At my lowest points, he always finds a way to dig me out. Thanks, mate.

I would also like to thank my closest friends (you know who you are) whose good humour and support always helps take my mind away from work pressures. A special thanks to Josh with whom I had interesting coding and statistics discussions. Another special shout to the Listening Sessions community. Music is a fundamental part of my life and it truly helps to keep me grounded. Special thanks to Mark (Yilan) with whom I have had many collaborative music production sessions over these years, providing the very best form of escapism.

Endless gratitude goes to my two children, Marley (4) and Rudy (1). No matter what sort of day I've had, these two rays of sunshine will always make me smile. It's an honour to be Dad to these two cheeky monkeys.

Finally, the biggest and most heartfelt thanks goes to my amazing wife and soul mate, Row. Without her unwavering love and emotional support, I would not have got to this point. This year, we celebrate 20 years together, and throughout those two decades she has always been my biggest supporter and cheerleader, helping me in ways that would be impossible for me to state. Thank you, Bot.

Abstract

In mid-latitudes, changes in large-scale atmospheric circulation can substantially alter regional climate including the likelihood of extreme events. Over recent decades, robust changes in circulation have been identified in observations, including a poleward shift of the summertime Southern Hemisphere (SH) jet and widening of the tropical belt. The midlatitude circulation is sensitive to external forcing and climate models project significant future trends under high radiative forcing scenarios. However, there remain significant uncertainties in these responses, including the relative importance of different external drivers, the physical mechanisms that underpin responses and the robustness of responses in ensembles of climate models.

This thesis aims to advance understanding through multi-model assessments of large-scale tropospheric circulation responses to CO_2 and non-CO_2 drivers by investigating dynamical sensitivity, rapid adjustments and sea-surface temperature (SST) mediated responses, and sources of uncertainty arising from structural model differences and internal variability, focusing on the SH.

SST-mediated feedbacks are shown to be the most important component of the large-scale response to greenhouse gases, scattering aerosols and solar forcing, with the evolution of SST warming patterns in the initial decade after abrupt forcing perturbations of particular importance. Rapid adjustments also contribute a substantial fraction of the midlatitude circulation response, especially to black carbon (BC) aerosols, with 75% of the SH jet shift to a 10xBC perturbation mediated by direct tropospheric heating anomalies. 80-90% of the intermodel difference in summertime midlatitude low-level zonal wind responses can be explained by differences in tropics-to-pole temperature gradient responses.

A role for anthropogenic aerosols in recent tropical width trends is suggested. However, ensemble sizes in current generations of Coupled Model Intercomparison Project (CMIP) models are too small to robustly detect and quantify this response. The results motivate development of large ensemble single forcing simulations to improve modelled representation of historical and projected future circulation shifts.

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Abbreviations

AA	Anthropogenic aerosols	**ITCZ**	Intertropical convergence zone
BC	Black carbon	**IPCC**	Intergovernmental Panel on Climate Change
BMB	Biomass burning aerosols	**IV**	Internal variability
CH_4	Methane	**JJA**	June - July - August
CMIP	Coupled Model Intercomparison Project	**MJO**	Madden-Jullian oscillation
CMIP5	Phase 5 of CMIP	**MAM**	March - April - May
CMIP6	Phase 6 of CMIP	**MMC**	Mean meridional circulation
CO_2	Carbon dioxide	**MMM**	Multi-model mean
D&A	Detection and attribution	**NH**	Northern Hemisphere
DJF	December - January - February	**PDRMIP**	Precipitation Driver and Response Model Intercomparison Project
EC	Emergent Constraints	**Psi500 (Ψ_{500})**	Meridional mass streamfunction at 500 hPa
ECS	Effective climate sensitivity	**PV**	Potential vorticity
EDJ	Eddy Driven Jet	**RCP**	Representative concentration pathway
EKE	Eddy kinetic energy	**SAM**	Southern annular mode
ENSO	El Niño Southern Oscillation	**SH**	Southern Hemisphere
ERF	Effective radiative forcing	**SO_4**	Sulphate aerosols
fSST	fixed-SST experiments	**Sol**	Solar radiative forcing
GCM	General circulation model	**SON**	September - October - November
GHG	Greenhouse gas	**SSP**	Shared socioeconomic pathway
GSAT	Global mean near-surface air temperature	**SST**	Sea surface temperatures
GMST	Global mean surface air temperature	**STJ**	Subtropical jet
HC	Hadley Cell	**Sul**	Sulphate aerosols
hPa	Hectopascals (1 hPa = 100 Pa)	**TTL**	Tropical tropopause layer
IGCM4	Intermediate Global Circulation Model version 4	**ZM**	Zonal mean

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Chapter 1

Introduction

1.1 Motivation

The large-scale atmospheric circulation is expected to alter under climate change (Lee et al., 2021). Observations, reanalysis datasets and climate models have shown changes in several measures of the large-scale general circulation over recent decades during which atmospheric CO_2 concentrations have monotonically increased and significant global warming has occurred (e.g. Johanson and Fu, 2009; Bender et al., 2012; Harvey et al., 2012; Grise and Medeiros, 2016; Gulev et al., 2021). Importantly, increases in global mean surface air temperature (GSAT) are not distributed evenly across the globe. Instead, the distribution of radiative heating and cooling is changing, resulting in adjustments to the thermal structure of the atmosphere accompanied by shifts of atmospheric circulation features (Lorenz and DeWeaver, 2007; Staten et al., 2012).

Changes to the large-scale circulation could have significant implications for regional climates and the hydrological cycle (e.g., Held and Soden, 2006; Kang et al., 2011, Harnik et al., 2016), both of which are largely controlled by atmospheric circulation (Vallis, 2006). Such regional changes include a narrowing of the inter-tropical convergence zone (ITCZ) (e.g., Byrne and Schneider, 2016), and a slowing (e.g., Vecchi and Soden, 2007; Kang and Lu, 2012) and poleward shift in the edge of the Hadley cell (e.g., Frierson et al., 2007; Lu et al., 2007; Davis and Rosenlof, 2012; Ceppi and Hartmann, 2013, Grise and Davis, 2020; Staten et al., 2020a), which is a measure of the width of the tropical belt. Changes of this nature could give rise to important societal impacts. For example, regions currently on the poleward side of the subtropical zones, such as the southwestern United States, and the southern Mediterranean (Seager et al., 2014), may have experienced drying and droughts associated with a poleward shift of the Hadley cell edge (Seidel et al., 2008; Seager et al., 2007; Hoerling et al., 2012; Dai, 2013; Seager et al., 2014). Subtropical desert regions could expand poleward and the incidence of wildfire could potentially increase (Scheff and Frierson, 2012; Feng et al., 2013).

In the extra-tropics, a poleward shift of storm tracks (e.g., Yin, 2005; Harvey et al., 2014) and the midlatitude jets, especially in the Southern Hemisphere (SH) (Barnes and Polvani, 2013; Simpson et al., 2014; Simpson et al., 2021), have been projected by climate models. Such changes could have potentially significant implications for regional extremes including heat waves and storms, with changes to extratropical cyclone characteristics (e.g., Tamarin-Brodsky and Kaspi, 2017; Harnik et al., 2016; Eichler, 2020). Furthermore, the prevalence and severity of wildfires in southeastern Australia in Austral winter and spring has been associated with a deficit in precipitation associated with the negative phase of the Southern Annular Mode (SAM), the leading mode of atmospheric variability in the SH (Harris and Lucas, 2019). Severe wildfires in the 2019/20 summer in southeastern Australia have been partially attributed to drought induced by an extreme negative phase of the SAM throughout 2019 (van Oldenborgh et al., 2021). Whilst that particular event is likely associated with a rare Antarctic Sudden Stratospheric Warming (SSW) event (Lim et al., 2020) and strong spring ozone anomaly with the smallest ozone hole since 1982, it serves to demonstrate the potential impacts of shifts in the mean SAM index and related large-scale circulation features associated with anthropogenic climate forcing on the SH regional climate. Biodiversity, tropical fishing grounds and agriculture could also be affected (Heffernan, 2016). For example, grain production is heavily influenced by climate (Ceglar et al., 2017), with reductions in rainfall associated with large-scale circulation changes having a potentially severe impact on crop yields (Heffernan, 2016), which may disproportionately affect low- and middle-income countries (Wilts et al., 2021). It is therefore important to understand anticipated large-scale circulation changes to inform governments and decision makers.

Shifts in circulation are also thought to have impacts on the global carbon cycle, for example by the influence of the SH surface westerly winds (Anderson et al., 2009; Waugh et al., 2013; 2019; Swart et al., 2014) and SAM (Lovenduski et al., 2007) on ocean meridional overturning circulations which mediates increased ventilation of the deep Southern Ocean, affecting carbon uptake. Alterations to the large-scale atmospheric circulation are therefore a potentially important source of large-scale climate feedbacks (Zickfield et al., 2007).

There remain large uncertainties in our quantitative understanding of large-scale circulation responses to changes in greenhouse gases (GHGs), aerosols and natural climate forcings (Harvey et al., 2015; Collins et al., 2014; Voigt and Shaw, 2015; Ceppi and Shepherd, 2017; Grise and Davis, 2020; Lee et al., 2021). Current understanding is hampered by a lack of knowledge of the responses to different climate forcings and by diversity in responses across different models, which is particularly evident in the variety of storm track (e.g. Laîné et al., 2009; Shaw et al., 2016), jet stream (e.g., Barnes and Polvani, 2013; Simpson et al., 2021) and tropical width (e.g., Garfinkel et al., 2015; Xian et al., 2021) responses in the literature. This limits our ability to make robust projections for future regional climate changes and to assess potential societal impacts. It is therefore critical that the representation of the large-scale circulation in climate models is accurate and that uncertainties arising from structural differences between models, as well as uncertainties due to internal climate variability, are understood and ultimately reduced (Bony et al., 2015).

1.2 The Vertical Structure of Earth's Atmosphere

The vertical profile of Earth's atmosphere consists of three layers. The lowermost layer is called the troposphere which typically extends to an altitude of ~16 km at the equator and ~8 km over the poles, though this varies by season. The troposphere is characterised by near-monotonically decreasing temperature with height, with a gradient of ~6.5 K km^{-1} referred to as the lapse rate. The lower troposphere also has a near-monotonically reducing equator-to-pole meridional temperature gradient, primarily a consequence of the zenith angle of incoming solar irradiance which increases with latitude. Above the troposphere is the stratosphere, which extends to a altitude of ~50 km and is typically characterised by increasing temperature with height due to the presence of ozone in the mid-stratosphere that absorbs solar radiation and releases heat via the Chapman oxygen-ozone cycle (Chapman, 1930). A relatively thin boundary exists between the two layers called the tropopause, which is thermally defined as the lowest height where the lapse rate falls to 2 K km^{-1} or less (Vallis, 2006). Above the stratosphere lies the upper atmosphere region, which consists of the mesosphere, thermosphere and exosphere. The troposphere, while being a relatively thin layer compared to the whole atmosphere in terms of its spatial height and volume, is much denser than the middle and upper atmosphere and contains the majority of atmospheric mass at higher pressure. It is where the large majority of weather occurs as well as the atmospheric transport of heat, with around two-thirds of poleward heat transport occurring through tropospheric processes and most of the rest through ocean circulation (Vallis, 2006), although this varies by region (Trenberth and Caron, 2001). The tropospheric circulation is the focus of this study.

1.3 The Large-Scale Circulation

The large-scale circulation is the flow of air on a synoptic to global scale, that is of the order of several hundreds to thousands of kilometres. This covers pressure and weather systems the size of extratropical cyclones up to mean air flows that span the entire planet. An elementary conceptualisation of the tropospheric circulation envisions a three large-scale cell structure of zonally-averaged meridional (i.e. north-south) overturning circulations in each of the Northern (NH) and Southern hemispheres, which can be generalised as the tropical, extratropical (or midlatitude) and polar regions.

The tropics are governed by the Hadley circulation, a thermally-driven overturning circulation which is dominated by moist convection and latent heating processes (more detail on the Hadley circulation is provided in section 1.3.1). The extratropical midlatitudes are more complex, being dominated by transient baroclinic eddies which govern meridional momentum, moisture and heat fluxes (Vallis, 2006) with contributions from planetary stationary waves (e.g. Kaspi and Schneider, 2013). As such the temperate extratropical midlatitude regions can experience more unsettled and fluctuating weather. The mean residual midlatitude meridional flow resulting from eddy and stationary wave activity resolves to a weak, thermally-indirect meridional quasi-overturning circulation, historically referred to as the Ferrel cell, with apparent ascending motion in the high latitudes near the polar region and descending motion in the subtropics, aligned with the sinking branch of the Hadley cell. However, the traditional Ferrel cell is considered an overly simplistic understanding of processes that govern the extratropical circulation (more detail on the midlatitude circulation is provided in section 1.3.2). The high latitudes comprise the Polar cell, which, like the Hadley cell, is thermally direct, but much weaker with a smaller meridional temperature gradient.

Near the surface, average zonal (i.e. east-west) winds follow in an easterly–westerly–easterly pattern from equator to pole, which must be maintained against surface friction by momentum convergence for midlatitude westerlies, which have a strong barotropic component, and by momentum divergence for tropical and polar easterlies. High-latitude surface easterlies are much weaker than those in the tropics, which are colloquially known as the trade winds. Surface winds are generally stronger in the winter, primarily due to a stronger meridional tropics-to-pole temperature gradient, and are also stronger in the SH as the smaller land mass and thus reduced orography means there is less friction and thus weaker surface drag (Vallis, 2006) (more detail on zonal winds is provided in section 1.4).

1.3.1 The Hadley Circulation

The tropical region has traditionally been defined in terms of Earth's relationship to the sun. Cartographers generally define the tropics between approximately 23.5 °N (the Tropic of Cancer) and 23.5 °S (the Tropic of Capricorn) based on the latitude where the Sun is directly overhead during the June and December solstices in the NH and SH respectively due to Earth's approximate 23.5° tilt of its rotational axis relative to its ecliptic. An oscillation in Earth's axial tilt means the location of the tropics by this definition varies between around 22.1° and 24.5° over an approximate 41,000 year cycle (Campisano, 2012). While this definition is useful for cartography and astronomy, it is not particularly useful for climatology and does not capture the behaviour of Earth's atmosphere that primarily governs differences in climate between tropical and extratropical regions. Instead, climatologists generally define the tropics based on measures of atmospheric circulation, in particular the Hadley cell circulation.

The Hadley cell is a large-scale overturning circulation named after amateur meteorologist, George Hadley, who first proposed it in 1735 (Hadley, 1735) as part of a hypothesis on the formation of the easterly trade winds that were fundamental to international trade, global exploration and thus the development of geopolitics during the 'Age of Sail' from approximately the mid-15th to the mid-19th century (Bankoff, 2017). Important and landmark early modern developments in models of the Hadley circulation were proposed by Schneider (1977) and Held and Hou (1980). The Hadley circulation is thermally direct, meaning that within the cell, warm air rises and cool air sinks. Important zonal asymmetries exist in the Hadley cell, including, for example, the Walker circulation in the Pacific region (e.g. Julian and Chervin, 1978; Schwendike et al., 2014), and Madden-Julian oscillation (MJO; Madden and Julian, 1971, 1972; Zhang, 2005).

The basic meridional components of the Hadley cell can be conceptualised as follows: An ascending branch of warm moist air rises near the equator to near the tropopause through convection (or more accurately the tropical tropopause layer (TTL; Fueglistaler et al., 2009)), where it moves polewards, carrying moisture and dissipating heat, facilitating cloud formation and precipitation and giving rise to equatorial wet zones, which present primarily as humid rainforest ecosystems. As moisture condenses out of the air, it dries, cools and descends typically at around 25 - 30° latitude (the descending branch), promoting dry zones at either poleward edge of the equatorial wet zones, which present as arid desert regions. The dry and cool air then moves back toward the equator, warming and picking up moisture as it goes, thus closing the circulation. This is often referred to as the mean meridional circulation (MMC).

The presence of Earth's anti-clockwise rotation and resulting Coriolis force causes meridionally flowing air to veer to the right in the NH and left in the SH. The ascending branch of the Hadley cell therefore veers to the east and accelerates in a zonal direction due to angular momentum conservation as the air moves poleward, closer to Earth's axis of rotation, in the upper troposphere. At a point, the winds become geostrophic, whereby the meridional pressure-gradient force is balanced by the Coriolis force, resulting in zonally flowing winds. These quasi-geostrophic winds manifest as westerly subtropical jet streams in the upper troposphere near the edge of the subtropics (see section 1.4). Near the surface, the descending branch veers to the west as it moves equatorward from the Hadley cell edge, creating the near-surface easterly trade winds that converge from the north and south near the equator in the low-pressure Intertropical Convergence Zone (ITCZ).

Due to the inherent seasonally asymmetric hemispheric heating which reaches its maximum during the two solstices, the latitude at which the Northern and Southern Hadley circulation cells meet, and thus the ITCZ, migrates with the seasons with a several week lag due to ocean thermal inertia. The two Hadley circulation cells therefore rarely meet at the equator and are rarely symmetrical in size. Because of this, the overall tropical circulation is mostly dominated by one winter Hadley cell, with its rising branch centred in the summer hemisphere and descending branch in the winter hemisphere, meaning the summer cell is entirely contained within the summer hemisphere.

1.3.2 Features of The Midlatitude Circulation

The extratropical midlatitude circulation is dominated by baroclinic instability which leads to synoptic eddy formation and turbulence and presents as the often quite drastic daily fluctuations in weather experienced in the extratropics, versus the generally more stable weather experienced in the tropics. Baroclinic instability, described by Vallis (2006) as "the form of hydrodynamic instability that most affects the human condition" due to its fundamental role in weather fluctuations, arises in stratified fluids that are both rotating and have a meridional temperature gradient. Baroclinic instability therefore appears in both the atmosphere and the oceans primarily in the midlatitudes where there is a strong meridional temperature gradient and density and pressure gradients are misaligned. Vertical shear of the geostrophic zonal wind is proportional to the meridional temperature gradient via the thermal wind relation (eqn. 1a and 1b):

$$f\frac{\partial u_g}{\partial z} = -\frac{\partial b}{\partial y}, \tag{1a}$$

$$f\frac{\partial v_g}{\partial z} = \frac{\partial b}{\partial x}, \tag{1b}$$

where f is the Coriolis parameter (= 2ΩsinΦ; where Ω = Earth's rate of rotation and Φ = latitude), u_g and v_g are the geostrophic zonal and meridional components of wind respectively, and b is the buoyancy (where b = $-g\partial p/p_0 = \partial\phi/\partial z$ in hydrostatic balance, where ϕ is a pressure deviation). We can derive the thermal wind relation from an examination of the Boussinesq horizontal and vertical momentum equations (eqns. 2 and 3):

$$\frac{Du}{Dt} + f \times u = -\nabla_z \phi\,, \tag{2}$$

$$f\frac{Dw}{Dt} = \frac{\partial \phi}{\partial z} + b\,. \tag{3}$$

Using mass conservation (eqn. 4) to scale vertical velocity, we find that where horizontal length scales are much larger than vertical height scales (as in the atmosphere), the advective term in the vertical momentum equation (3) can be neglected and therefore we can assume approximate hydrostatic balance:

$$\nabla_z \cdot u + \frac{\partial w}{\partial z} = 0\,. \tag{4}$$

In the midlatitudes, we can also assume geostrophic balance, whereby the Rossby number is small ($Ro = U/fL \ll 1$) and therefore the advection term in the horizontal momentum equation can be neglected in relation to the dominant rotation term and the pressure gradient force and Coriolis force are therefore approximately balanced. For a small Rossby number, $u \approx u_g$ and $v \approx v_g$. The combination of the geostrophic and hydrostatic balance approximations thus results in the thermal wind relation (Vallis, 2019). Importantly, when the atmosphere is assumed to be in hydrostatic balance (i.e. stably stratified) with a monotonic meridional density gradient, it is unstable to small perturbations as air parcels are able to release potential energy as kinetic energy as they move along a sloping path. An associated reversal in the potential vorticity (PV) gradient allows eddies to form and grow. The presence of baroclinic instability explains how eddies are responsible for the majority of heat transport in the midlatitudes and is key to explaining how the presence of a meridional horizontal temperature gradient is fundamental to the formation of midlatitude westerlies (Vallis, 2019).

As well as baroclinic instability, the other major source of turbulence in the midlatitudes is the existence of surface features such as orography and land-sea temperature contrasts. These asymmetries result in stationary planetary waves that contribute to momentum transport (Kaspi and Schneider, 2013). Their amplitude and structure varies seasonally, associated with zonal wind characteristics, with stronger waves in winter due to stronger land-sea temperature contrasts and zonal winds (Wang and Ting, 1999; Willis et al., 2019). Stationary waves are larger in the NH compared to the SH and contribute to regional climate features such as precipitation and temperature differences (e.g. Broccoli and Manabe, 1992; Simpson et al., 2015). Planetary waves also mediate teleconnections between phenomena in different regions, such as remote impacts arising from the El Nino southern oscillation (ENSO) (e.g. Turner and Annamalai, 2012).

Despite the inherent chaotic nature, the large-scale turbulence in the midlatitudes may be described as ‘geostrophic turbulence’ (Vallis 2006 p.539) as the mean flow has time- and zonally-averaged hydrostatically and geographically balanced features which are important to understand.

1.4 Jet Streams

There are two major westerly atmospheric jet streams that develop in Earth's troposphere, the subtropical jet (STJ), and the midlatitude (also the polar-front or extratropical) or 'eddy-driven' jet (EDJ) (Figure 1.1).

The STJ, sometimes referred to as the 'thermally-driven' jet, forms in the upper troposphere, peaking around 200 - 300 hPa (~10 km altitude) near the edge of the subtropics at a latitude of around 30°. It is primarily a consequence of angular momentum conservation through the Coriolis force acting on the poleward flow in the upper branch of the Hadley Cell, and thermal wind balance due to meridional temperature gradients (Held and Hou, 1980). It is mostly baroclinic in nature, exhibiting significant shear (Vallis, 2006).

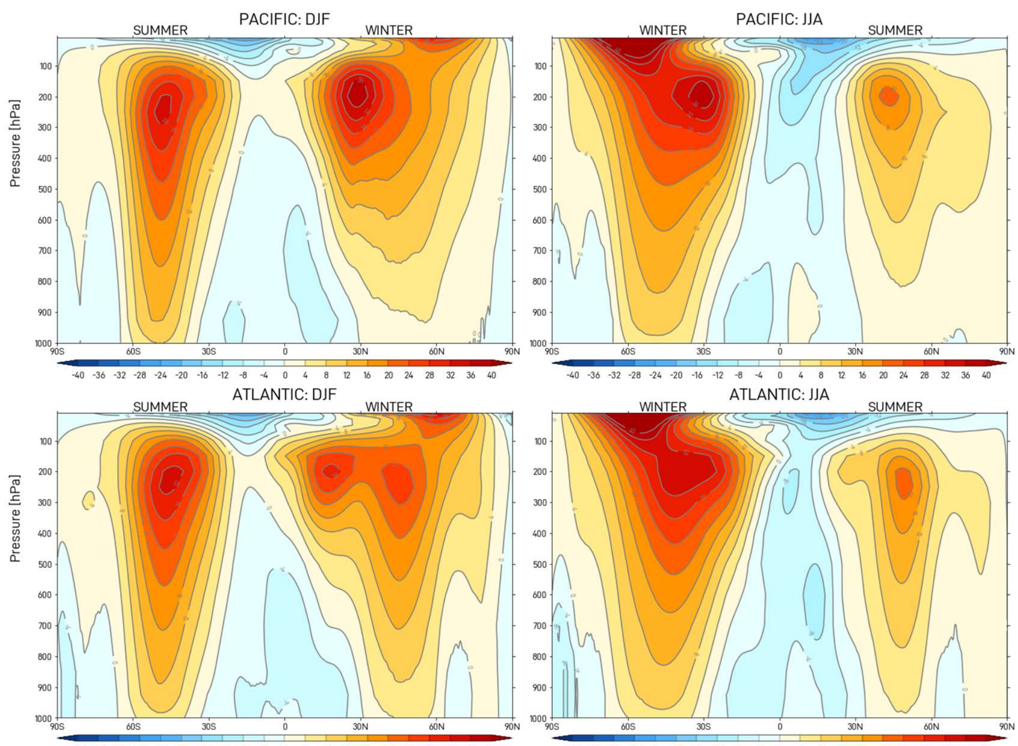

Figure 1.1: Zonal wind averaged over the (top) Pacific Ocean (150°W - 120°E) and (bottom) Atlantic Ocean (60°W - 0°) in DJF (left) and JJA (right) for the period 1980 - 2015 in the ERA5 reanalysis dataset.

The EDJ extends from near the tropopause to the surface, therefore experiencing effects from surface friction. The EDJ is formed through westerly eddy momentum flux convergence of baroclinic eddies (Held, 1975; Rhines, 1975; Panetta and Held, 1988) which is necessary to counter the effect of surface drag (Vallis, 2006). Indeed, it has been shown in two-layer quasi-geostrophic models that a westerly EDJ can form spontaneously through this mechanism in the absence of planetary angular momentum and a STJ, and that in the absence of eddies, the EDJ would not form (Panetta, 1993; Lee, 1997). In the midlatitudes on the poleward side of the Hadley Cell, there exists a region of strong baroclinicity and therefore a tendency for baroclinic eddies to form (see Section 1.3.2), creating a forcing on the mean flow with baroclinic waves that propagate meridionally from the source region. This transfers momentum and energy poleward (Trenberth and Stepaniak, 2003), the convergence of which drives the EDJ (Held, 1975, Rhines, 1975; Pfeffer, 1981; Lee and Kim, 2003). The EDJ is more barotropic than the STJ and as such exhibits less shear (Vallis, 2006). The EDJ is associated with the location of the storm track and so the location of synoptic midlatitude weather systems, including precipitation and cloud distribution.

The degree by which the STJ and EDJ can be distinguished from each other varies by location and season (Simpson et al, 2014; Manney et al, 2014). It is also important to recognise that in the real atmosphere, the two jets never operate in isolation and will always have some degree of influence on each other (Lee and Kim, 2003; Walker and Schneider, 2006; Li and Wettstein, 2012).

In the NH, typically the jets are more separated in the mid-winter (DJF) in the North Atlantic basin (Eichelberger and Hartmann, 2007; Figure 1.1.), while in the North Pacific basin, the jets are more likely to be colocated (Figure 1.1.) as the region of strongest baroclinicity, and therefore maximum eddy generation, is at approximately the same latitude as the STJ (Lee and Kim, 2003; also see Figure 1.1). When this collocation occurs, it is usually referred to as a 'merged' or 'combined' jet with each influenced by both mechanisms (Li and Wettstein, 2012). A merged jet is more likely where there is a strong STJ, while a weak STJ tends to allow for a well separated EDJ (Lee and Kim, 2003). The greater presence of orography, land-sea temperature contrasts and sea-surface temperature patterns compared to the SH means that large-scale planetary waves have a greater influence on the NH midlatitude circulation and contribute to larger zonal asymmetries (Held et al., 2002).

Although the SH midlatitudes are more zonally symmetric than the NH (Allen and Kovilakam 2017, Grise et al., 2018), as shown by the similarity between SH Pacific and Atlantic basins in Figure 1.1., important zonal asymmetries exist arising from mean-state tropical and extratropical sea-surface temperature (SST) patterns and Antarctic orography, particularly in the winter (JJA), where there exists a split jet in the South

Pacific (e.g. Patterson et al., 2020), with a more defined EDJ, and a more merged jet over the Indian Ocean (Messori et al., 2021). The Austral summer (DJF) zonal mean EDJ is located at around 40-50°S and is relatively more separated from the weak STJ (Nakamura and Shimpo, 2004; also see Figure 1.1). There are also associated asymmetries in the seasonal SH storm tracks (Hoskins and Hodges, 2005). The annual mean SH EDJ is located around 52°S (Karpechenko and Maycock et al., 2018).

A shifting of the EDJ position is the leading mode of variability in the extratropical zonal mean zonal wind. In the SH, a meridional shifting of the EDJ is strongly associated with the SAM (e.g. Goyal et al., 2021), the dominant mode of large-scale atmospheric variability in the SH (Gong and Wang, 1999; Marshall, 2003; Ho et. al, 2012). The SAM describes an oscillation of atmospheric air mass between mid- and high-latitudes, with its positive-phase corresponding to a low pressure anomaly over high latitudes and high pressure anomaly over the midlatitudes and vice-versa for its negative phase (Marshall, 2003). This shifting air mass is coincident with a shift of the EDJ position, whereby a positive SAM is associated with a more poleward and stronger EDJ and midlatitude storm track and vice versa (e.g. Swart et al., 2015). Many forced EDJ responses project onto the leading mode of variability and thus resemble the SAM, therefore one can infer modifications to the EDJ from studies focussed primarily on the SAM.

1.5 The Response of the Midlatitude Circulation to External Forcing

1.5.1 Historical trends

A wide body of literature points to the midlatitude circulation changing in response to a variety of external forcings, including anthropogenic climate change (e.g. Vallis et al., 2015; Shaw et al., 2016; IPCC AR6 - Lee et al., 2021). Since the mid-1970s, it has been known that upper-tropospheric eddy kinetic energy (EKE) shifts poleward in climate models forced by a doubling of atmospheric CO_2 concentrations (2xCO2; Mannabe and Wetherald, 1975). Since then a theoretical underpinning for a poleward shift of the midlatitude circulation under greenhouse gas (GHG) forcing has been developed, although the understanding of the mechanisms that drive midlatitude circulation shifts is still incomplete (Vallis et al., 2015, Lachmy and Shaw, 2018; Shaw, 2019). There have been robust trends in the extratropical SH circulation in observations and reanalysis datasets of the late 20th-century since satellite observations became available at the end of the 1970s. This includes a poleward shift of the SH EDJ (e.g., Thompson et al., 2011; Swart et al., 2012; Goyal et al., 2021), storm tracks (e.g., Yin, 2005; Chang et al., 2012; Shaw et al., 2016; Priestley et al., 2020) and Hadley cell edge (e.g., Davis and Rosenlof, 2012; Garfinkel et al., 2015), and a positive trend in the SAM (Thompson and Solomon, 2002; Marshall, 2003; Gillett and Fyfe, 2013; Gillett et al., 2013; Swart et al., 2015). These trends are strongest and statistically significant in the Austral summer season (DJF) (e.g. Swart et al., 2015) while smaller trends that cannot be robustly statistically distinguished from internal variability tend to be shown in other seasons. Of the remaining seasons, trends are largest and come closest to significance in Austral autumn (MAM) (Goyal et al., 2021). Some studies also point to a robust strengthening of the SH jet in certain seasons (e.g. Swart and Fyfe, 2012; Goyal et al., 2021). In the first two decades of this century, the poleward SH circulation and more positive SAM trends appear to have paused (i.e. 21st-century trends have no statistical significance), which has been linked to the onset of SH stratospheric ozone recovery (e.g. Banerjee et al., 2020). Further discussion of the attribution of historical trends to climate drivers, including ozone, is provided in section 1.5.3.

Observations for the NH often exhibit greater variability (e.g. Archer and Caldeira, 2008, Allen et al, 2012a; Davis and Rosenlof, 2012, Davis and Birner, 2013, Barnes and Polvani, 2013; Shepherd, 2014; Grise et al., 2019b; Albern et al., 2021) and show greater heterogeneity in regional trends (e.g. Strong and Davis, 2007, Barnes and Polvani, 2013; Simpson et al., 2014) due to asymmetries arising from, for example, land-sea temperature contrasts, orography and Arctic sea ice loss (e.g. Overland and Wang, 2010; Barnes and Screen, 2015; England, 2021), although there are also potentially important SH asymmetries (e.g., Bracegirdle et al., 2013; Waugh et al., 2020; Patterson et al., 2020) that have not been thoroughly investigated.

This study focuses mainly on the Southern Hemisphere and as such the remainder of this section will discuss the SH circulation response.

1.5.2 Mechanisms of Zonal Mean SH Large Scale Circulation Shifts

Several mechanisms have been proposed to explain the observed historical and forecast future shifts in the SH zonal mean circulation (Vallis et al., 2015; Shaw, 2019). Due to the complexity of the system, including the coupling of the atmosphere and ocean, isolating causality and quantifying the contribution of different factors to the overall midlatitude circulation shift is a significant scientific challenge. There remains a lack of consensus regarding the main drivers of observed shifts with several physical mechanisms having been proposed that require further investigation to improve understanding. A number of approaches to this investigation can be taken, including using physical equations, scaling arguments and climate models in a hierarchy of complexity from very simple to fully atmosphere-ocean coupled earth system models (Maher et al., 2019; see section 1.5.2 for more detail on climate modelling).

The poleward shift of midlatitude circulation features such as the EDJ and storm tracks (as discussed above) is coincident with a poleward shift of the maximum eddy momentum flux convergence, which dominates the formation of low-level westerlies in the midlatitudes and plays a dominant role in zonal mean midlatitude circulation shifts (Simpson et al., 2014).

Shaw (2019) describes three barotropic scenarios that result in a poleward shift of eddy momentum flux convergence in response to increased atmospheric CO_2 concentrations and a coincident poleward shift of the EDJ (*Figure 1.2*): the first involves a dipolar eddy momentum flux convergence response resulting from a combination of increased poleward momentum flux and increased equatorward wave propagation in the EDJ region; the second involves increased equatorward momentum flux and poleward wave propagation on the equatorward side of the EDJ, leading to a reduction in eddy momentum flux on the equatorward side of the jet and thus a poleward shift the EDJ position; the third involves a dipole of eddy momentum flux on either side of the EDJ which leads to a poleward shift of the location of maximum eddy momentum flux.

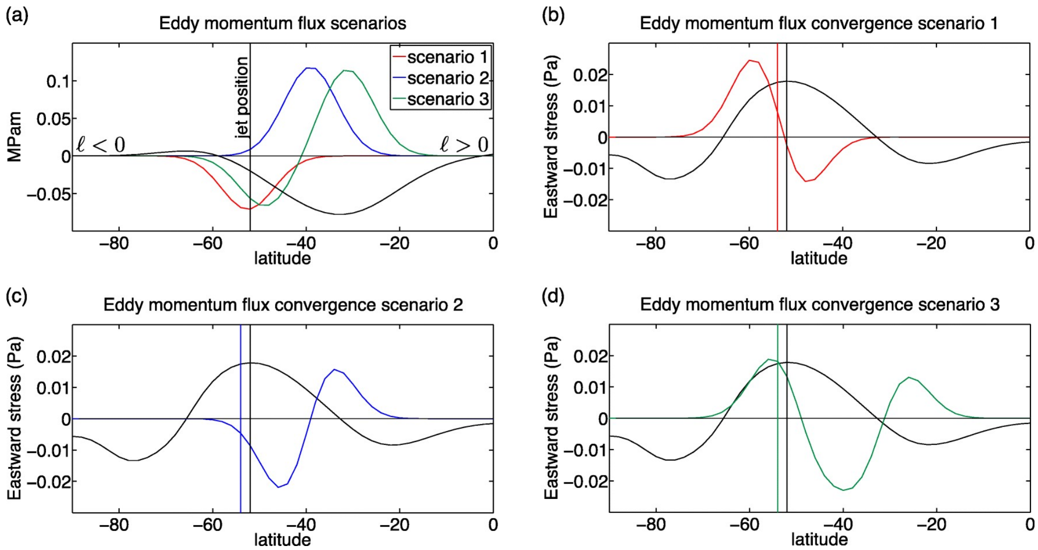

Figure 2.2: Three eddy momentum flux convergence scenarios that result in a poleward shift of the EDJ as described by Shaw (2019). a) vertically integrated eddy momentum flux and b–d vertically integrated eddy momentum flux convergence responses to increased CO_2 relative to climatology (black, divided by 10) in the SH. Vertical coloured lines in b–d indicate the position of maximum eddy momentum flux convergence for the different scenarios relative to the climatology (vertical black line).

Some of the mechanisms that may explain the formation of the scenarios described above and thus a poleward EDJ shift are described here:

In the tropics, increased atmospheric CO_2 concentrations increase latent heat release in the upper troposphere (Vallis et al. 2015). The latent heating increases tropical dry static stability and increases the equator-to-pole upper tropospheric temperature gradient, which strengthens the STJ due to thermal wind balance (*see Section 1.3.2*). These static stability and temperature gradient responses are connected to a poleward shift in the mean potential vorticity (PV) gradient and eddy PV flux on the poleward side of the EDJ (Butler et al, 2010); a net increase in anticyclonic wave breaking (Riviere, 2011); a reduction in the meridional absolute vorticity gradient on either side of the EDJ, resulting in equatorward wave propagation due to increased wave reflection on the poleward side of the EDJ, as well as a poleward shift of eddy momentum flux (Kidson and Vallis, 2012; Lorenz, 2014); and a change in effective diffusivity (Lu et al., 2014). All of these responses result in a poleward shift of the EDJ.

In the extratropics, increased CO_2 concentrations also increase dry static stability while also acting to lift the tropopause height (Vallis et al., 2015). These changes to the vertical atmospheric temperature structure may also drive a poleward EDJ shift through changes to equator-to-pole temperature gradients, eddy momentum flux, eddy length scale and phase speed, and wave propagation (e.g. Kidson et al., 2010; 2011; Chen et al., 2008; Frierson, 2008).

Another mechanism that may contribute to a poleward EDJ shift involves CO_2-induced stratospheric cooling, perhaps due to changes to wave refraction which may enable eddies to propagate further poleward, therefore shifting the EDJ poleward (Wu et al., 2013). Increases in specific humidity may also be driven by increased CO_2 which could change dry and moist static energy fluxes also enabling a poleward EDJ shift (e.g. Shaw and Voigt, 2016b). It is also likely that there is an important role for both longwave and shortwave cloud radiative effects (CRE), due to changes to mean static energy transport, for example (e.g., Voigt and Shaw, 2015; Shaw and Voigt 2016b). It is important to emphasise that causality remains a key question, as some of the proposed drivers of circulation shifts described here may in fact be a response to circulation shifts themselves. There is ongoing research attempting to disentangle and isolate causality.

1.5.3 Climate Modelling

The drivers of the observed shift in the SH midlatitude circulation over the late 20th-century, as discussed above, have been analysed using a suite of general circulation models (GCMs). A large proportion of studies use data from the Coupled Model Intercomparison Project (CMIP) archive, with modern studies using the most recent iterations of the project, namely Phase 5 (CMIP5; Taylor et al., 2012) and Phase 6 (CMIP6; Eyring et al., 2016). These GCMs attempt to reproduce the climate response in the historical period by running simulations with the best estimates of historical forcings, and project future climate change by imposing a range of emissions projections scenarios. In CMIP5 these were known as Representative Concentration Pathways (RCPs), which included emissions scenarios ranging from a future where strong emissions reductions are implemented (e.g. RCP2.6 where a 2.6 Wm^{-2} radiative forcing is present by the end of the 21st century) to a worst-case scenario (RCP8.5 where an 8.5 Wm^{-2} radiative forcing is present by the end of the 21st century) (van Vuuren et al., 2011). In CMIP6, a more comprehensive and sophisticated set of projections were used, known as Shared Socioeconomic Pathways (SSPs), which incorporated potential socioeconomic factors that may contribute to emissions pathways (Gidden et al., 2019). Running identical prescribed-forcing simulations in multiple models enables an assessment of the average and range of expected climate responses and the associated uncertainties arising from inherent model structural differences. Individual models tend to run multiple simulations with the same forcings imposed in an attempt to measure and account for internal climate variability, although these ensemble sizes are generally limited in the CMIP archives primarily due to time and budgetary constraints.

While most GCMs simulate an average poleward shift of the SH midlatitude circulation and shift to a more positive SAM in response to imposed historical forcing, and individual members are able to capture the observed magnitude of the shift, there is a large inter-model spread and the multi-model mean (MMM) in successive generations of CMIP shows smaller poleward shifts than observed (Johanson and Fu, 2009; Barnes and Polvani, 2013; Hu et al., 2013; Allen et al., 2014a; Vallis et al., 2015). However, the observed historical climate is but one realisation of the range of possible trends that may be simulated by models. Therefore, as observed trends lie within the range of model simulated trends, it is plausible that the observations reflect a certain state of internal climate variability and therefore that some models are able to reproduce observed historical forcing. The magnitude of the contribution to historical trends arising from external anthropogenic forcing and internal variability is a key scientific question, as it is likely that both have had some influence, and there is uncertainty and model biases associated with both.

Projections of future poleward shifts in the SH midlatitude circulation also show a large range across different GCMs. For example, annual mean EDJ shifts by the end of the 21st century versus the average position in 1979-2014 under the RCP8.5 or SSP5-8.5 emissions scenario range between around -1° and around 6° poleward in CMIP5 models and -1° to 8° in CMIP6 models (see Figure 2 in Simpson et al., 2021). Identifying and reducing the uncertainty of these projections arising from both internal variability and model structural differences - such as differences in modelled spatial patterns of surface and atmospheric temperature responses (e.g. Murphy et al. 2002, Butler et al. 2010; Harvey et al. 2014, 2015; Ceppi and Shepherd 2017), modelled Antarctic sea ice extent (e.g. Bracegirdle et al., 2018), and circulation climatology, with a potential role for a persistent climatological equatorward EDJ bias in the magnitude of the modelled jet shift (e.g. Kidston and Gerber., 2010; Barnes and Hartmann, 2010, Ceppi et al., 2012; Bracegirdle et al., 2013; Simpson and Polvani, 2016; Simpson et al., 2021), the representation of which has somewhat improved in CMIP6 (e.g. Curtis et al., 2020; Bracegirdle et al., 2020; De et al., 2021) - and thus the development of emergent constraints (ECs) (i.e. finding statistical relationships between climate metrics to constrain future projections) is another key challenge in climate dynamics (e.g. Hall et al., 2019; Simpson et al., 2021).

In addition to historical and future simulations, many CMIP models have performed simulations over the historical period with either individual forcings or a subset of forcings, which can be compared to historical all-emissions forcing simulations to ascertain their contribution to historical climate change (Gillett et al., 2016). These include, for example, GHG-forcing only, anthropogenic aerosol-forcing only, solar-forcing-only, volcanic-forcing-only and stratospheric-ozone-forcing only simulations. Many of these models have also performed single forcing future projections for comparison against all-forcing RCPs or SSPs to assess potential future contributions. Collectively, this type of study is referred to as 'detection and attribution' (D&A) (Bindoff et al., 2013).

Whilst many studies use transient forcing simulations as described above, a separate line of investigation involves the use of idealised forcing experiments, whereby step-change perturbations are imposed. Idealised forcing experiments allow the equilibrium climate response to specific forcing agents to be investigated. For example, a core experiment in CMIP5 and CMIP6 models is an instantaneous quadrupling of atmospheric CO_2 concentrations (4xCO2) relative to pre-industrial (commonly defined as 1850) levels. The perturbed simulations are then compared against a pre-industrial climatology. This allows for a clean analysis of the effects of increased CO_2 while also developing an understanding of the roles of internal climate variability and systematic differences between models, facilitating understanding of sources of modelling uncertainty. Another example of an idealised forcing multi-model intercomparison project relevant to the present study is the Precipitation Driver and Response Model Intercomparison Project (PDRMIP: Myhre et al., 2017) which includes five idealised abrupt single forcing experiments, including CO_2, methane (CH_4), sulphate and black carbon aerosols and solar forcings (see Chapter 2).

1.5.4 Detection and Attribution Studies

1.5.4.1 GHGs

Detection and attribution (D&A) studies have attempted to explain the observed poleward shift of the midlatitude circulation and have suggested a role for anthropogenic GHG forcing (e.g. Hall et al., 1994; Fyfe et al., 1999; Kushner et al., 2001;

Cai et al., 2003; Shindell and Schmidt, 2004; Yin, 2005; Arblaster and Meehl, 2006; Lorenz and DeWeaver, 2007; Butler et al., 2010; Gerber and Son, 2014; Fogt and Marshall, 2020; Morgenstern, 2021). These studies have used several diagnostics of zonal mean or regional mean large scale circulation, including the jet stream (e.g. Kushner et al. 2001; Yin, 2005, Fyfe and Saenko 2006; Kidston and Gerber 2010; Swart and Fyfe 2012; Wilcox et al., 2012; Woollings and Blackburn 2012; Barnes and Polvani 2013; Bracegirdle et al. 2013; Ceppi et al., 2014), the SAM (e.g. Fyfe et al., 1999; Cai et al., 2003; Rauthe et al., 2004; Miller et al. 2006; Previdi and Liepert 2007; Woollings and Blackburn 2012; Gillett and Fyfe 2013; Choi et al., 2019; Fogt and Marshall, 2020; Morgenstern, 2021) and storm tracks (e.g. Yin 2005; Chang et al. 2012, Bender et al., 2012, Harvey et al., 2014, Shaw et al., 2016). GHGs mediate an increase in GSAT, with polar surface temperature amplification, tropical tropospheric warming, and stratospheric cooling (in response to increased CO_2 (Goessling and Bathiany, 2016)) (IPCC, 2014). These heterogeneous temperature responses translate to a strengthening of the mid-to-high latitude meridional temperature gradient, with an increased upper troposphere-lower stratosphere temperature gradient, which is related to a shift of the midlatitude westerlies due to thermal wind balance and changes to baroclinicity (e.g. Ceppi et al., 2012; Harvey et al., 2014).

Scenario-dependent GHG forcing over the 21st century is projected to be associated with further shifts in the SH midlatitude circulation, with GHGs having a large influence on high-forcing scenarios such as RCP8.5 and SSP5-8.5, which project a significant poleward shift by the end of this century, although there is large uncertainty in the magnitude of response across GCMs (e.g. Barnes and Polvani, 2013; Simpson et al., 2021). More realistic lower-forcing scenarios project more subtle effects that are harder to distinguish from internal variability (e.g. Kushner et al., 2001; Previdi and Liepert; 2007; Ceppi and Hartmann, 2013; Barnes and Polvani, 2013).

1.5.4.2 Stratospheric Ozone

The SH midlatitude circulation response to non-GHG forcing is also of great interest. Indeed one of the most robust anthropogenic signals in the SH that has emerged from modelling studies is the influence of stratospheric ozone depletion on the Austral summertime (DJF) poleward midlatitude circulation shift occurring after the peak of the Antarctic ozone hole in the Austral spring (Farman et al., 1985; Thompson and Solomon, 2002; Gillett and Thompson, 2003; Arblaster and Meehl, 2006; Son et al., 2009; Son et al., 2010; Polvani et al., 2011; McLandress et al., 2011; Thompson et al., 2011; Staten et al., 2012; Min and Son, 2013; Eyring et al., 2013; Previdi and Polvani, 2014; Schneider et al., 2015; Waugh et al., 2015; Fogt and Marshall, 2020; Morgenstern, 2021). SH stratospheric ozone loss peaks in October and mediates a stratospheric cooling response. The effect of this stratospheric ozone loss and cooling on the troposphere develops over the following months and generally presents in mid-summer (November / December). The exact mechanism for how this ozone-depletion-induced cooling drives tropospheric midlatitude circulation shifts is not well understood (Kidston et al., 2015, Garcia, 2021). Orr et al. (2012) suggests that ozone depletion causes a strengthening of lower-stratospheric winds and the polar vortex which reduces planetary wave propagation from the troposphere. This initial reduction in wave propagation causes a feedback that further strengthens the polar vortex therefore further restricting upward planetary wave propagation, which leads to the strengthened stratospheric winds being drawn downwards to the tropopause. Poleward heat and momentum fluxes develop in the troposphere from the resulting changes to tropospheric wave activity mediated by their inability to propagate upward into the stratosphere, thus increasing baroclinicity and shifting the midlatitude jet poleward.

Following the implementation of the 1987 Montreal Protocol and subsequent amendments that sought to ban the use of ozone-depleting substances (ODSs) such as chlorofluorocarbons (CFCs), stratospheric ozone levels began to stabilise in the mid-1990s. Ozone is projected to recover to pre-1980s levels over the coming decades (Karpechko and Maycock et al., 2018) which is expected to have an impact on the SH summertime circulation over the remainder of this century (e.g. Perlwitz et al., 2008; Son et al., 2008; Son et al., 2009; Son et al., 2010; Barnes et al., 2014; Banerjee et al., 2020). The poleward forcing from stratospheric ozone loss will be removed and potentially reversed, offsetting the effects of emissions scenario-dependent increases in GHG concentrations over the next decades (e.g. Arblaster et al., 2011; Gerber and Son., 2014; Morgenstern, 2021).

The effect of stratospheric ozone concentrations in seasons other than Austral summer is smaller, with GHGs and other forcings likely to be more influential (Arblaster and Meehl, 2006). However, it has been found that ozone mediated effects might extend into the Austral autumn, with a potentially robust signal found in May that is strongest in the Pacific sector (Ivy et al., 2017).

1.5.4.3 Tropospheric Aerosols

Aerosols such as sulphates and black carbon (BC) play an important role in net effective radiative forcing (ERF) and global and regional surface temperature trends, although there remains substantial uncertainty regarding the magnitude of their impact (Forster et al., 2007; Myhre et al., 2013, Carslaw et al., 2013, Bellouin et al., 2019, Eyring et al., 2021). Differences in aerosol representation in climate models, arising from, for example, uncertainty in aerosol-cloud interactions, is a crucial source of variance in responses between climate models (e.g. Rotstayn et al., 2015). More uncertainty exists regarding the importance of anthropogenic aerosol (AA) emissions for large-scale circulation trends and, while their impact is generally not well understood, there have been suggestions that aerosols mediate a change in the midlatitude circulation (e.g., Allen et al., 2012a, Gillett et al 2013; Allen et al, 2014; Steptoe et al., 2016; Allen and Ajouku, 2016; Rotstayn, 2013; Rotstayn et al., 2014, Choi et al., 2019). Despite the majority of aerosol emission sources and aerosol burden being localised in the NH (Hoesly et al., 2018), an aerosol-forced atmospheric circulation signal may also occur in the SH (e.g. Gillett et al., 2013; Rotstayn, 2013; Steptoe et al., 2016; Wang et al., 2020). There may be a role for NH temperature anomalies in mediating SH circulation shifts, as shown, for example, in an idealised modelling study by Ceppi et al. (2013). However, current understanding of this effect and its importance for observed circulation trends is highly uncertain. The robustness of the AA influence on the SH circulation has been questioned (e.g. Steptoe et al., 2016; Choi et al., 2019) and a substantial gap in knowledge remains.

There are likely to be differences in response depending on aerosol species. Sulphate aerosols cool the atmosphere and surface by scattering incoming solar radiation and increasing planetary albedo, and thus have a negative ERF. In the troposphere, this is in opposition to the effects of increased GHGs therefore increased sulphate concentrations may act to offset GHG-mediated poleward migration of the midlatitude circulation (Fischer-Burns et al., 2009). Conversely, BC has a positive ERF and induces localised tropospheric heating through shortwave absorption (Stjern et al., 2017). The resulting changes to meridional temperature gradients and static stability may significantly influence the midlatitude circulation (e.g. Shen and Ming, 2018; Zhao et al., 2020). BC also reduces albedo at the poles through deposition on snow. However, the overall BC impact on the climate is complex and still uncertain due to factors including cloud, convection and precipitation feedbacks and their representation in GCMs (e.g. Ramanathan and Carmichael, 2008; Bond et al., 2013; Allen et al., 2019;

Johnson et al., 2019). The effect of sulphates and BC in isolation has not been thoroughly studied, with attribution experiments with transient emissions tending to combine all AA forcings together, and most idealised BC forcing experiments tending to be focused primarily on the Northern Hemisphere (e.g. Allen et al., 2012a; Allen et al., 2014; Kovilakam and Mahajan, 2015; Allen and Ajoku, 2016). With substantial reductions in AA emissions projected over the coming decades (Rao et al., 2017) it is vital that their impact is better understood as there is potential for a significant impact on modifications to future atmospheric circulation.

1.5.5 Idealised Forcing Studies

While the major drivers of historical SH circulation changes are broadly known, as identified in D&A studies, the quantification and importance of different forcings, and their contribution to inter-model spread in large-scale circulation responses, is uncertain. Isolating and quantifying the specific large-scale circulation responses to individual climate drivers is essential for accurate modelling and reducing uncertainty in emissions scenario projections, as different drivers are expected to follow different emissions pathways (Gidden et al., 2019). Individual forcers have varying effectiveness in driving GSAT change (Richardson et al., 2019) and may have idiosyncratic effects on other aspects of the climate, including the large-scale circulation (e.g. Zhao et al., 2020). Different aerosol species, for example, are likely to mediate strong, regionally varying responses in precipitation (e.g. Liu et al., 2018).

Idealised forcing experiments allow the climate response to individual forcings to be investigated. Studies of the 4xCO2 experiment in CMIP5 and CMIP6 show a robust poleward shift of the SH EDJ with most models agreeing on the sign of change whilst also showing significance against internal variability (e.g. Barnes et al. 2014; Grise and Polvani 2014a, 2016; Curtis et al., 2020). However, while there is a qualitative understanding of the large-scale circulation response to CO_2 forcing (e.g., Kushner et al., 2001; Kidston and Gerber, 2010; Barnes and Polvani 2013; Ceppi et al., 2014; Grise and Polvani, 2016), there is a large spread of several degrees latitude in the magnitude of EDJ response across models to the same forcing scenario (Barnes and Polvani 2013; Barnes et al. 2014; Grise and Polvani 2014a, Curtis et al, 2020) indicating the difficulty in accurately quantifying the response. Furthermore, the separation of the effect of internal variability and structural model differences and their relative contribution to this spread is difficult to overcome given the relatively small (in statistical terms) datasets currently available.

The response of a GCM to increased atmospheric CO_2 is primarily characterised by the effective (or sometimes 'equilibrium') climate sensitivity (ECS). This is defined as the equilibrium GSAT change in response to a doubling of atmospheric CO_2 concentrations (e.g. Knutti and Hegerl, 2008), which is highly correlated with GSAT change in response to 4xCO2. There is a large model spread in ECS across CMIP5 (Andrews et al., 2012) and CMIP6 (Forster et al., 2020) models. However, while ECS is a useful diagnostic for the planetary surface temperature response to CO_2 forcing and is a simple and effective tool for characterising the general sensitivity of a model (or indeed the Earth) to CO_2 forcing, it is unable to account for the complexity of the model spread in large-scale circulation responses. This was noted by Grise and Polvani (2014b, 2016), who defined the magnitude of a model's large-scale circulation response to CO_2 forcing as its 'dynamical sensitivity' (which Gerber and Son (2014) referred to as 'circulation sensitivity'). They showed the model spread in ECS accounts for around 20% of the model spread in the annual mean SH jet shift on centennial timescales, with the relationship weaker and statistically nonsignificant in winter and spring (<5%) and stronger with statistically significance in summer and autumn (~35%) (Grise and Polvani, 2014b). The weak relationship between ECS and dynamical sensitivity indicates complexities in the spread of dynamical responses that are not captured by differences in ECS. Thus it cannot be assumed that models with larger ECS will also produce a larger midlatitude circulation response and efforts to constrain ECS therefore have limited value for the constraint of midlatitude circulation projections. As the dynamical response to climate change may control potentially drastic changes to aspects of regional climates, such as temperature and precipitation (Zappa, 2019), Grise and Polvani (2014b) argue that understanding dynamical sensitivity is of greater importance in terms of its societal impacts than understanding ECS.

Several studies show the importance of the spatial pattern of surface and atmospheric temperature changes to large-scale circulation changes. This includes changes to upper- and lower-tropospheric temperature gradients (e.g. Murphy et al. 2002; Polvani et al., 2011; Arblaster et al., 2011; Ceppi et al., 2012; Wilcox et al., 2012; Gerber and Son, 2014; Harvey et al., 2014, 2015; Ceppi et al., 2014, Choi et al., 2019), an increase or decrease in static stability (e.g. Butler et al., 2010; Ceppi and Shepherd 2017; Grise et al., 2019b) and changes to sea-surface temperature (SST) patterns (e.g.

Patterson et al. 2020) and SST fronts (Inatsu and Hoskins, 2004; Marshal and Connolley, 2006; Nakamura et al., 2008; Ogawa et al., 2016), all of which may contribute to regional changes in baroclinicity. Cloud radiative effects are also an important source of uncertainty for SH midlatitude circulation trends. Ceppi et al (2012) showed that biases in midlatitude shortwave cloud forcing were correlated with the mean climatological SH EDJ latitude in CMIP5 models. Ceppi et al. (2014) then showed that shortwave cloud feedbacks play a key role in model spread of SH jet latitude responses across CMIP5 models. This finding was corroborated by Grise and Polvani (2014c) and more recent studies (e.g. Voigt and Shaw, 2016; Ceppi and Shepherd, 2017; Voigt et al., 2019; Li et al., 2019; Grise et al., 2019a; Grise and Kelleher, 2021; Voigt et al., 2021) also highlight the contribution of CRE to uncertainty in modelled midlatitude circulation responses. Cloud feedback is also a significant contributor to differences in ECS across models (e.g. Bony and Dufresne, 2005), with cloud responses over the SH extratropics recently found to contribute to ECS differences across GCMs (Zelinka et al., 2020).

The extent to which these factors can explain the model spread varies by region and by season, with the EDJ response to forcing itself also varying by season due to climatological differences (McGraw and Barnes, 2016) and seasonal factors such as stratospheric polar vortex trends (Ceppi and Shepherd, 2019). However, regional and seasonal uncertainty can be even larger than in zonal and annual means (e.g. Shepherd, 2014; Simpson et al., 2014), while the zonal mean can potentially average out opposing regional trends. Therefore, a significant gap in knowledge exists regarding expected changes and the relative sources of uncertainty on smaller temporal and spatial scales.

While the midlatitude response to CO_2 has been well studied due to the prominence of the 4xCO2 experiment in CMIP, responses to other climate forcers, including short-lived climate forcings like methane and aerosols, such as sulphates and black carbon, as well as natural forcings, such as solar irradiance, have received much less attention, especially in the SH. Relatively few datasets exist that include idealised non-CO_2 single forcing experiments, with a low priority for such experiments in CMIP programmes. Another gap thus exists in our understanding of the large-scale circulation response to these forcings, further limiting accuracy and robustness of model projections.

1.5.6 Fast and Slow Responses of the SH Midlatitude Circulation

An emerging topic in studies of the response of the midlatitude circulation to forcing is the relative importance of "direct or rapid adjustments" and "indirect or slow feedback" processes. Rapid adjustments are the direct effects of forcing on the atmosphere without the influence of surface temperature driven feedbacks (Sherwood et al., 2015). Rapid adjustments are an important part of the top-of-atmosphere energy budget, affecting the balance of incoming and outgoing radiation through changes to clouds, humidity, atmospheric thermal structure and surface fluxes (Sherwood et al. 2015; Smith et al. 2018). They are a key component of ERF, distinct from climate feedbacks driven by surface temperature responses (e.g. SSTs) which tend to act on longer timescales. In practice, however, rapid adjustments are isolated in GCMs by fixing SSTs and sea ice (fSST experiments) at a pre-defined climatology (usually either pre-industrial or present day) which allows for a contribution from land surface temperature changes. These changes may be an important component of regional circulation responses to direct forcing (Shaw and Voigt, 2016a), so while rapid adjustments diagnosed from fSST experiments are often assumed to be representative of the direct, atmosphere only adjustments, there exists a caveat that the inclusion of land surface temperature changes may alter results. Despite the use of terms such as 'rapid' or 'fast' and 'slow', the separation of timescales of has not yet been formalised (Sherwood et al., 2015; Smith et al., 2020). It has been suggested that rapid adjustments are a key source of uncertainty in aerosol ERF, especially black carbon due to its strong atmospheric shortwave absorption (e.g. Smith et al., 2020). While a substantial body of work now exists on the contribution of rapid adjustments to ERF, their effect on the large-scale circulation is a nascent field of study.

Deser and Phillips (2009) is an early example of a study of the relative roles of SST and direct atmospheric forcing on the large-scale circulation. They showed that the direct atmospheric effects of observed changes in a combined set of forcings in the CAM3 model were distinct from the effects of observed SST forcing, and contributed to the overall poleward shift of the SH midlatitude jet in austral summer. The large-scale circulation response to 4xCO_2 in the CMIP5 models was decomposed into the rapid adjustment and SST-driven components by Grise and Polvani (2014a). They showed that the overall poleward shift of the EDJ, and differences in the response between hemispheres, was primarily mediated by the SST-driven component, corroborating earlier work highlighting the importance of SSTs in the overall SH midlatitude circulation response (Staten et al., 2012). However, they also found a contribution to the poleward shift from rapid adjustments. Grise and Polvani (2017) then found that the timescale of SH midlatitude circulation response varies by season, with the circulation shifting poleward roughly in line with GSAT change in the austral summer, but showing a quicker response than GSAT change in winter. Ceppi et al. (2018) showed that the majority of the SH EDJ response to 4xCO2 in CMIP5 occurs within the first decade after the forcing is applied. However, only around half of the eventual quasi-equilibrium GSAT change occurs during this time. They investigated this further using a single GCM with experiments where SSTs were either fixed or fully coupled to the atmosphere and allowed to adjust, therefore enabling a separation of rapid adjustments and SST-driven responses. This difference in the timescale of response was explained by the evolving pattern of SSTs. In particular, a more slowly warming Southern Ocean compared to the tropical oceans in the initial period after forcing is applied.

The few studies that exist on the effect of rapid adjustments on the SH midlatitude circulation mostly focus on the response to increasing CO_2. The relative importance of rapid adjustments and indirect feedbacks for the response to other forcings, such as non-CO_2 GHGs and aerosols, as discussed in the previous section, are underexplored and thus less well known. Zhao et al. (2020) found that the location of the SH Hadley cell edge shifts poleward in response to BC forcing and equatorward in response to sulphate forcing, and that rapid adjustments contribute to these shifts. Given the relationship between midlatitude circulation and Hadley cell width on interannual timescales (Kang and Polvani, 2011; Ceppi and Hartmann, 2013; Staten and Reichler, 2014), this suggests a potential role for rapid adjustments on the extratropical circulation response to these forcings.

1.6 The Tropical Width Response to External Forcing

Observations and reanalysis datasets show that the width of the tropics has expanded over the late 20th-century (e.g. Seidel et al., 2008; Johanson and Fu, 2009; Garfinkel et al., 2015; Lucas and Nguyen, 2015; Davis and Birner, 2017; Kim et al., 2017; Grise et al., 2019b). However, the quantification of this expansion has been a persistent problem for more than a decade (Staten et al., 2020b). One issue is methodological uncertainty arising from the use of different metrics to measure tropical width (Waugh et al., 2018), and the degree to which these metrics are correlated with each other (Solomon et al., 2016). Metrics that have been used include those based on meridional mass stream function (Ψ), the subtropical tropopause break, outgoing longwave radiation (OLR), the STJ, the EDJ, precipitation minus evapotranspiration (P - E), near-surface winds (U_{SFC}), and sea level pressure (*Figure 1.3*; Waugh et al., 2018). The zero-crossing of the meridional mass stream function at 500 hPa (Ψ_{500}) is the most common definition of the zonal mean Hadley cell edge and tends to be the benchmark against which other zonal mean metrics are compared. Upper- and lower-tropospheric metrics are only weakly correlated (e.g. Solomon et al., 2016; Davis and Birner, 2017) and it has been suggested that lower-tropospheric metrics are a more appropriate measure for Hadley cell width than those based on the upper troposphere due to their greater correlation with Ψ_{500} (Kang and Polvani, 2011; Waugh et al., 2018). For example, Menzel et al. (2019) showed that the STJ (an upper-tropospheric metric) and Hadley cell respond differently to increased CO_2, whereas Kang and Polvani (2011) showed a relationship between the SH EDJ (a lower-tropospheric metric) and Hadley cell edge (especially in austral summer), and this relationship has been corroborated and further elucidated in subsequent studies (e.g., Ceppi and Hartmann, 2013; Staten and Reichler, 2014). There has been effort in recent years to standardise methodologies to address this issue (e.g. Adam et al., 2018; Waugh et al., 2018). Another issue in quantifying the historical tropical width response is the disagreement across different reanalysis products regarding the rate of expansion and climatological extent, although modern reanalysis products tend to show a smaller spread than early ones (Garfinkel et al., 2015; Grise et al., 2019b). Despite these issues, that the tropics have expanded is a robust finding in both reanalysis and GCMs shown in several of the metrics listed above (Xian et al., 2021). The remaining questions therefore relate to the quantification of the expansion, its attribution to different climate drivers (e.g. Staten et al., 2012), and the role of natural (Mantsis et al., 2017; Allen and Kovilakam, 2017) and internal (Garfinkel et al., 2015) variability including the detection of an anthropogenic-forced signal (e.g. Grise et al., 2019b). Regional differences in Hadley cell expansion is also an emerging topic (e.g. Schwendike et al., 2014, Grise et al., 2018; Staten et al., 2019; Staten et al., 2020a; Martin et al., 2020). The current best estimate of the rate of tropical expansion over the last several decades is around 0.2 - 0.5° decade^{-1} (Grise et al., 2018; Davis and Davis 2018; Staten et al., 2020a).

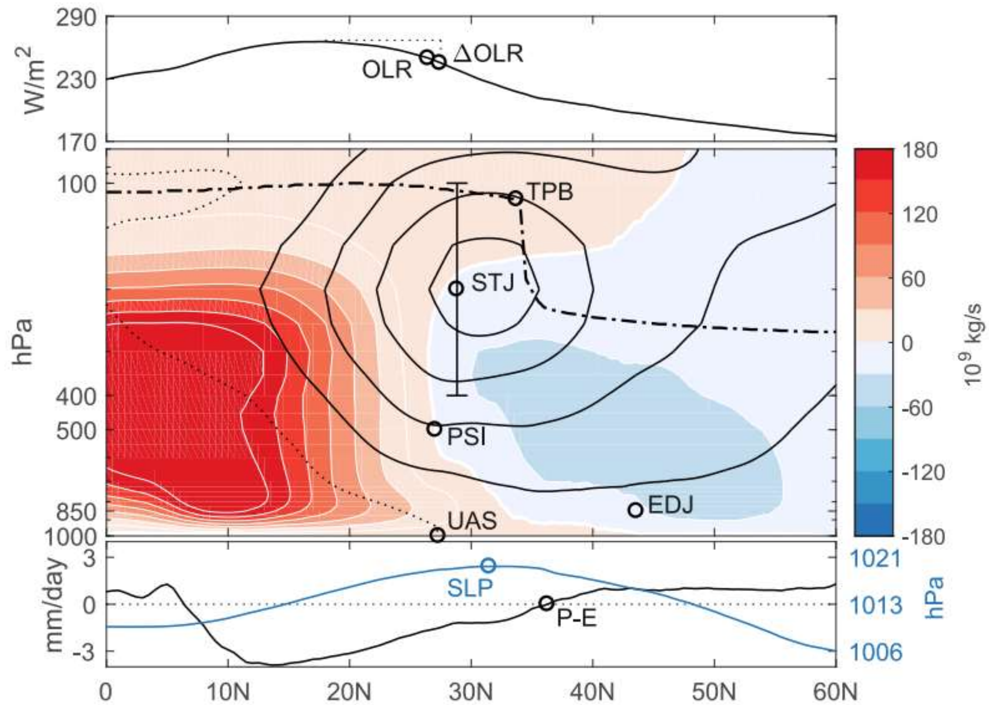

Figure 1.3: Schematic of the NH atmospheric circulation and structure. Tropical width metrics are shown by symbols. In the top panel, OLR is outgoing longwave radiation. In the middle panel, zonal winds are shown by black contours, meridional streamfunction (PSI) is shown by coloured shading and the tropopause is shown by a dot–dashed curve (where TPB = tropopause break). UAS indicates the latitude of near-surface zonal wind zero crossing and EDJ is the latitude of the eddy-driven jet. In the lower panel, sea level pressure (SLP) is shown by the blue curve, and P-E shown by the black curve. Taken from Waugh et al. (2018).

Similarly to the midlatitude circulation discussed in the previous section, the expansion of the SH Hadley cell has been attributed to anthropogenic drivers including GHGs (Lu et al., 2007; Hu et al., 2013; Nguyen et al., 2015; Tao et al., 2016; Grise and Davis, 2020), and stratospheric ozone depletion, mainly in the austral summer (e.g. Min and Son, 2013, Waugh et al., 2015; Grise et al., 2019b). A role for anthropogenic aerosols has also been suggested in both the NH and SH, though this may be metric and seasonally dependent (Ming and Ramaswamy, 2011; Zhao et al., 2020; Grise and Davis, 2020). In the NH, the role of BC has been particularly highlighted as a potential driver of tropical expansion which may be proportionally stronger at mediating a shift than GHGs per unit ERF (Allen et al., 2012a, 2014; Kovilakam and Mahajan, 2015; Shen and Ming, 2018). Shortwave absorption by BC aerosols increase gross static stability which has been hypothesised to shift the zone of baroclinic instability, and thus the Hadley cell edge, poleward (Allen et al., 2012b; Tandon et al., 2013), though this mechanism is still a matter of investigation (Frierson et al., 2007; Kang and Lu, 2012; Vallis et al. 2015).

Scattering aerosols such as sulphates have been hypothesised as a driver of tropical contraction, having the opposite effect on static stability to BC and offsetting GHG-induced expansion trends. Therefore, the reduction of sulphate aerosol emissions over the remainder of this century may contribute to future tropical expansion (Allen and Ajoku, 2016). Yang et al. (2020) suggested a significant role for poleward advancing meridional SST gradients in driving tropical width trends, especially in the SH due to the larger proportion of ocean surface compared to the NH. Compared to the SH, models show a smaller signal from GHGs in the NH and a larger role for internal variability (Grise et al., 2019b). The relative role of these forcings in Hadley cell width trends and whether anthropogenic-forced signals can be separated versus internal variability, is a key question that has yet to be resolved (e.g. Garfinkel et al., 2015; Allen and Kovilakam, 2017; Amaya et al., 2018; Grise et al., 2018, 2019; D'Agostino et al., 2020), especially for the effect of AA (Steptoe et al., 2016; Choi et al., 2019).

1.7 Thesis Aims and Structure

The aim of this thesis is to advance the understanding of the response of the large-scale tropospheric circulation to CO_2 and non-CO_2 climate change drivers and the representation of these changes in GCMs, with a focus on the SH. Through multi-model assessments, this thesis will develop the understanding of dynamical sensitivity, rapid adjustments and SST-mediated large-scale circulation responses, and sources of uncertainty arising from structural differences between models and internal variability.

Chapter 2 contains a study published in Journal of Climate (Wood et al., 2020a) evaluating the SH midlatitude circulation responses to rapid adjustments and SST-driven feedbacks in a set of abruptly perturbed idealised forcing experiments in an ensemble of nine GCMs participating in the Precipitation Driver and Response Model Intercomparison Project (PDRMIP). The study builds on previous work by quantifying both the annual and seasonal SH midlatitude circulation response to CO_2 and non-CO_2 forcings, including methane, sulphate aerosols, BC aerosols and solar irradiance, apportioning the response mediated by rapid adjustments and by SST-driven feedbacks. The study identifies sources of uncertainty in the SH midlatitude circulation response and the degree to which changes to meridional tropics-to-pole temperature gradients can explain the inter-model spread in these circulation responses.

Chapter 3 contains a study published in Environmental Research Letters (Wood et al., 2020b) evaluating the difference in 'fast' SH midlatitude circulation responses to an idealised quadrupling of CO_2 in the two latest generations of multi-model ensembles (CMIP5 and CMIP6) to investigate why the poleward shift of the SH EDJ in CMIP6 is smaller than found in CMIP5, despite a higher range of ECS in CMIP6. Building on previous work showing the importance of evolving SST responses to forcing in the shift of the SH EDJ (Ceppi et al., 2018), new prescribed sea surface temperature simulations are executed in an intermediate complexity climate model (IGCM4), with CMIP5 and CMIP6 multi-model mean SST warming patterns in the first decade after forcing is applied (when the majority of the difference in response is realised) imposed. The study investigates whether differences in SST warming patterns can account for the differences in SH EDJ responses. The study breaks the response down regionally by SH ocean basin, building on previous studies showing that regional responses to forcing can show different behaviours.

Chapter 4 contains a manuscript prepared for submission to Environmental Research Letters that evaluates the tropical width response to AA over the 1950-2005 period. The study first questions whether an AA-forced response signal in three tropical width metrics can be identified and quantified using CMIP5, CMIP6 and two single model initial condition large ensembles (SMILEs). The study then investigates whether the single-forcing ensembles in each model are sufficiently large to robustly detect a signal from AA given the degree of internal variability, and what size of ensemble would be required to detect such a response in GCMs, both in the annual mean and seasonally. The study thus questions whether previous studies that rely on models with few ensemble members are statistically robust, and whether they allow sources of uncertainty due to structural differences between models to be identified. Building on the emerging topic of regional tropical widening, the study then evaluates the annual and seasonal, regional ocean basin responses in the SMILE models in both the NH and SH to identify any robust seasonal and regional AA-forced signals that may be masked in zonal mean measures. The study also attempts to identify potential structural differences between the two SMILE models examined.

Finally, Chapter 5 synthesises the main findings of the previous three chapters, places them into the wider context of current scientific understanding, and suggests avenues for future work. Appendices include supplementary information for Chapters 2-4.

Chapter 1 References

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Chapter 2

The Southern Hemisphere Midlatitude Circulation Response to Rapid Adjustments and Sea Surface Temperature Driven Feedbacks

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Published in Journal of Climate (2020)

DOI: 10.1175/JCLI-D-19-1015.1

Chapter 3

Role of Sea Surface Temperature Patterns for the Southern Hemisphere Jet Stream Response to CO_2 Forcing

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Published in Environmental Research Letters (2020)

DOI: 10.1088/1748-9326/abce27

Chapter 4

Aerosol-forced tropical expansion in the late 20th century: robust anthropogenic signal or internal variability?

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Prepared for submission to Environmental Research Letters

4.1 Abstract

Previous studies based on multi-model ensembles have shown uncertainty in the contribution of anthropogenic aerosols (AA) to historical tropical width trends. Here we use two single model initial condition large ensembles (SMILEs) to investigate the role of AA-forcing for tropical width trends over 1950-2005. We show that single-forcing ensembles with only a few members, typical of Coupled Model Intercomparison Projects, are insufficient to robustly identify tropical width trends from AA and mix uncertainty from internal variability and structural model uncertainty. As a consequence, the small multi-model mean tropical width trends due to AA are found to be highly sensitive to model choice. The two SMILEs reveal structural differences in the tropical width response to AA, for example CESM1 shows no tropical width trend due to AA while CanESM2 shows a significant annual mean contraction. This is because of seasonal differences in trends, in particular CESM1 shows NH Hadley cell expansion in boreal autumn which cancels the Hadley cell contraction in spring and winter also found in CanESM2. The results motivate producing single forcing large ensembles for multiple climate models to enable a robust isolation of the response to individual drivers and associated structural uncertainty.

4.2 Introduction

Observation and reanalysis datasets show the tropical belt widened by around 0.2° to 0.5° decade^{-1} between 1980 and the mid-2000s (Grise et al., 2018; Davis & Davis 2018; Staten et al., 2020). Hadley cell (HC) expansion has been attributed to anthropogenic drivers and internal climate variability (IV; Grise et al., 2019; Allen and Kovilakam, 2017). In the Southern Hemisphere (SH), climate models show that increased atmospheric greenhouse gas (GHG) concentrations (Lu et al., 2007; Hu et al., 2013; Nguyen et al., 2015; Tao et al., 2016; Grise and Davis, 2020) and stratospheric ozone depletion (e.g. Min and Son, 2013) have contributed to HC expansion (Grise et al., 2019; Garfinkel et al., 2015). In the Northern hemisphere (NH), models show a smaller signal from GHGs and a larger role for IV in recent tropical width trends (Grise et al., 2019). Some studies have suggested a role for anthropogenic aerosols (AA) in driving tropical width trends and meridional shifts in the extratropical circulation (Allen et al., 2012; Gillett et al., 2013; Zhao et al., 2020, Wood et al., 2020a), but the magnitude and robustness of this effect over the historical period remains unclear (Steptoe et al., 2016; Choi et al., 2019). Grise and Davis (2020; henceforth GD20) identified an AA-driven contraction of the NH HC edge during 1950-2005 in Coupled Model Intercomparison Project phase 5 models (CMIP5), but the response in phase 6 (CMIP6) models is metric-dependent, with a surface zonal wind metric showing a nonsignificant trend. In the SH, GD20 found a significant AA-forced contraction of the annual mean HC edge in both CMIP5 and CMIP6 models, but no significant trend in austral summer (DJF), suggesting the AA impact on the SH HC is seasonally dependent. Zhao et al. (2020) found that idealised sulphate and black carbon (BC) aerosol perturbations were proportionately stronger than CO_2 at driving tropical width trends than GHGs relative to their respective effective radiative forcing.

The impact of anthropogenic drivers on historical climate trends is commonly determined using climate model simulations that isolate individual forcings (e.g. Gillett et al., 2016). Such simulations were performed for the major anthropogenic climate drivers within CMIP5/6, but these typically comprise only a few ensemble members for each model. Small ensembles have limited use for robustly identifying externally-forced responses of highly variable aspects of climate and for isolating structural differences in responses between models (Milinski et al., 2020; McKenna and Maycock, 2021). Steptoe et al. (2016) examined the impact of AA on the SH circulation using AA-only experiments from 10 CMIP5 models. However, 9 of the 10 models provided only ≤5 ensemble members. The AA-only results of GD20 are based on only 8 CMIP5 and 9 CMIP6 models, with all models providing ≤5 ensemble members in CMIP5 and ≤10 members in CMIP6. Therefore, their results likely mix uncertainty from IV with uncertainty from structural model differences.

Single Model Initial Condition Large Ensembles (SMILEs; Deser et al., 2020) offer a way to separate forced signals from IV and to differentiate forced responses between models (e.g., Maher et al., 2021a). This study assesses the contribution of AA to historical tropical width trends using SMILEs performed by two models. We place equivalent results from CMIP5/6 models into the context of the SMILEs. We specifically address: 1) how does the intermodel spread in tropical width trends in CMIP5/6 AA experiments compare with the inter-member spread and intermodel spread in forced response of the SMILEs? 2) Is it possible to detect structural differences in the tropical width response to historical AA forcing between CMIP5/6 models with small ensembles? 3) How many members are needed to detect structural differences between models?

Many studies concentrate on zonal mean tropical width trends, but there is a growing emphasis on regional heterogeneity of both AA and HC trends (e.g. Grise et al., 2018; Staten et al., 2019; Wood et al., 2020b; Waugh et al., 2020; Yang et al., 2020, 2021; Diao et al., 2021). Therefore, we also assess regional tropical width trends.

4.3 Methods

4.3.1 Datasets

4.3.1.1 CMIP5 and CMIP6

We use 11 and 14 models from CMIP5 (Taylor et al., 2012) and CMIP6 (Eyring et al., 2016), respectively, that performed historical AA-only (histAA and histAER) experiments (Gillett et al., 2016) and a pre-industrial control (piControl) experiment. The models and number of ensemble members are shown in Table S1.

4.3.1.2 Single Forcing SMILEs

CESM1

We use data from the CESM1-CAM5 large ensemble (CESM1-LE; Kay et al., 2015) single forcing experiments (Deser et al. 2020). Three experiments are used:

1. HIST: As in CMIP5 historical experiment [40 members]
2. XAER: Industrial and energy/transportation AA emissions (including both sulphates (Sul) and black carbon (BC)) fixed at 1920 levels. Other forcings including biomass burning emissions follow HIST (Diao et al., 2021). [20 members]
3. XBMB: Anthropogenic and natural biomass-burning emissions held fixed at 1920-levels. Other forcings follow HIST. [15 members].

CanESM2

The CanESM2 large ensemble (CanESM2-LE; Gagné et al., 2017; Oudar et al., 2018) performed 50 member ensembles for:

1. HIST: As in CMIP5 historical experiment.
2. AER: AA emissions including industrial and energy/transportation AA emissions and anthropogenic biomass burning as in HIST. All other forcings (GHGs, land use, natural aerosols) fixed at 1950 levels. The AER simulations are spun off from the five CMIP5 histAA CanESM2 ensemble members in 1950 and thus are identical in design.
3. NAT: Natural forcings as in HIST. All other forcings fixed at 1950 levels.

In addition, the piControl simulations from the two SMILE models are used to quantify IV. In CanESM2, the ERF from sulphate aerosol is very similar to that for AA over the historical period (Boucher et al., 2013 - Table 7.5). The sulphate and black carbon aerosol burdens in the SMILE experiments are shown in Fig. S1.

4.3.1.3 Comparability of Ensembles

Following GD20, we analyse trends for 1950-2005 when the global AA burden increased strongly (Boucher et al., 2013).

There are experimental differences between the single forcing SMILEs. The impact of industrial and energy/transportation AA in CESM1-LE is inferred by subtracting XAER from HIST (likewise for biomass-burning aerosols (BMB)), whereas in AER in CanESM2-LE (and the histAA experiment in CMIP5) aerosols are left while all other forcings are removed. This leaves the possibility of differences between CESM1-LE and CanESM2-LE due to potential non-linearities between aerosol and GHG interactions (England, 2021; Deng et al., 2020; Gettelman et al., 2016). There is also a difference in the combination of forcings included in XAER and AER. XAER does not remove anthropogenic biomass burning emissions while AER does include this as part of AA. Despite these differences, we consider the experiments sufficiently similar to warrant comparison and contend that they reflect complementary lines of evidence to build a picture of the tropical width response to AA forcing.

4.3.2 Tropical Width Metrics

The latitude of the HC edge in each hemisphere is calculated using the Tropical-width Diagnostics package (TropD; Adam et al., 2018). Three metrics are used from the 'lower tropospheric' group of metrics that most closely relate to the HC (Waugh et al., 2018). These are (see Supplementary Information):

1. ϕ_{UAS} - the latitude of near-surface zonal wind zero crossing;
2. ϕ_{EDJ} - the latitude of the eddy-driven jet (EDJ);
3. $\phi_{\Psi500}$ - latitude at which the zonal mean meridional mass streamfunction at 500 hPa crosses zero.

Monthly mean fields are first zonally averaged (by hemisphere or by region). Metrics are then calculated from zonal or regional mean wind fields on a monthly basis and then time-averaged to give annual and seasonal means. Figures show poleward expansion as a positive trend.

4.3.3 Regional Ocean Basins

Regional HC edges are calculated for the North Atlantic (NA; 60°W - 0) and North Pacific (NP; 120°E - 140°E) NH ocean basins (Oudar et al., 2020), and the South Atlantic (SA; 40°W - 20°E), Indian Ocean (Ind; 30°E - 120°E) and South Pacific (SP; 150°E - 290°E) SH ocean basins (Bracegirdle et al., 2013).

4.3.4 Statistical Methods

4.3.4.1 Trend detection

Where models have more than two ensemble members, ensemble mean trends are said to be statistically significantly different from zero if the 95% confidence interval of a bootstrapped distribution of the ensemble mean does not include zero. The number of bootstrap samples is taken to be the number of unique combinations, s, of N-members randomly sampled with replacement from an N-member ensemble, defined by:

$$s = \frac{(2N-1)!}{N!\,(N-1)!} \tag{1}$$

The bootstrap method is used due to the small number of members in the models, with even the 50-member CanESM2-LE ensemble distribution of trends being non-Gaussian. When only a single member exists, the trend is said to be statistically different from zero when outside of the range of trends due to IV, calculated as the 2.5 - 97.5th percentile range of the distribution of overlapping 55-year trends from the piControl simulation.

Multi-model mean (MMM) ensemble means are said to be significantly different from zero following the same bootstrapping methodology described above, but where N is the number of models. CMIP5 and CMIP6 distributions are compared with a Welch's t-test and are said to be significantly different when $p < 0.05$.

4.3.4.2 Ensemble size for detecting a response

Following Screen et al. (2014) and McKenna and Maycock (2021), we estimate the minimum ensemble size (N_{min}) required to detect a statistically significant forced tropical width trend given a certain IV and trend magnitude. We use the rearranged equation for a two-sided Student's t-test for a difference in means:

$$N_{min} = 2{t_c}^2 \times \left(\frac{\sigma}{X}\right)^2 , \tag{2}$$

where t_c is the t-statistic for p = 0.025 and $2N_{min}$ - 2 degrees of freedom, σ is the standard deviation of 55-year trends in the piControl simulation, and X is the magnitude of forced response.

We use the SMILEs to calculate the probability that a statistically significant ensemble mean trend from a given CMIP model would result by chance due to sparse sampling. The trends in the CESM1-LE XAER and CanESM2-LE AER ensembles are first pooled by subtracting their respective ensemble means. The distribution is then randomly sampled taking N-sized samples where N is the ensemble size of a given CMIP model and these are averaged to give an ensemble mean; this is repeated 10^4 times. The percentile of this distribution corresponding to the CMIP model trend is then the likelihood of finding a trend of that magnitude due to internal variability despite a zero forced trend. This method assumes the pooled LE standard deviation is roughly equivalent to the CMIP model IV.

4.4 Results

4.4.1 Annual and Zonal Mean Tropical Width Trends

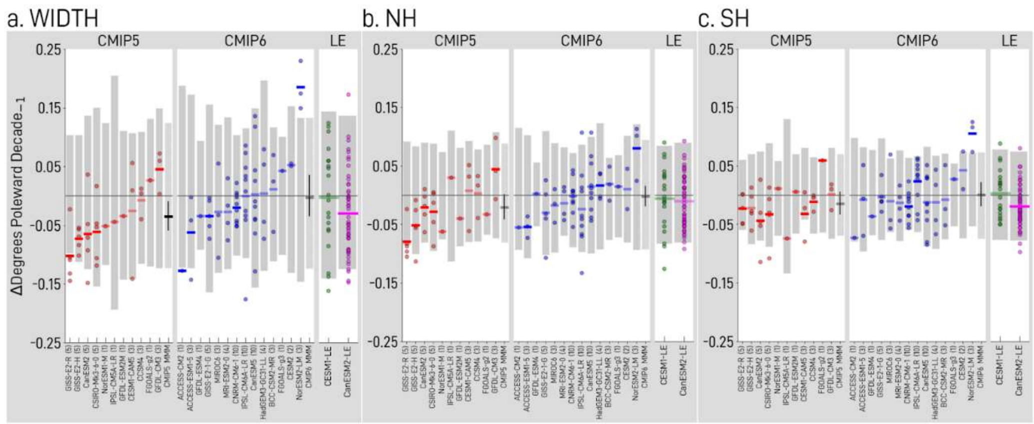

Figure 1: Zonal and annual mean ϕ_{UAS} trends [° decade^{-1}] over 1950-2005 in the histAA (CMIP5: red), histAER (CMIP6: blue), HIST - XAER (CESM1-LE: green) and AER (CanESM2-LE: magenta) experiments for a) tropical width, b) NH and c) SH trends. Circles show ensemble members and horizontal lines show ensemble means. Opaque lines show statistically significant ensemble mean trends (p<0.05) and semi-transparent means nonsignificant trends. Black horizontal lines adjacent to CMIP5 and CMIP6 models show the multi-model means with 95% confidence intervals. Grey bars show piControl internal variability (IV) in each model (see Methods). The median IV of CMIP5/6 models is shown by the grey bar in the MMM column.

The CMIP5/6 models show large inter-model spread in ensemble mean ϕ_{UAS} trends in both hemispheres (σ_{NH} = 0.036°dec^{-1}, σ_{SH} = 0.038°dec^{-1}) as shown in Figure 1 which shows zonal and annual mean ϕ_{UAS} 1950-2005 trends due to AA in the CMIP5 and CMIP6 models and the CESM1-LE and CanESM2-LE. Corresponding trends in $\phi_{\Psi500}$ and ϕ_{EDJ} are shown in Supplementary Figure S2. The inter-model spread is broadly comparable to the inter-member spread in CESM1-LE and CanESM2-LE (σ_{NH} = 0.056°dec^{-1} and 0.036°dec^{-1}; σ_{SH} = 0.040°dec^{-1} and 0.044°dec^{-1}, respectively). Given the CMIP5/6 ensemble sizes are generally small, with 14 of 25 models having ≤3 members, there is likely to be a substantial contribution to the intermodel spread from IV. Consistent with this idea, around two thirds of CMIP5/6 models show statistically nonsignificant ϕ_{UAS} trends due to AA in both hemispheres.

A minority of CMIP5/6 models do appear to show a statistically significant ϕ_{UAS} trend, but there is little agreement in those that show significant ϕ_{UAS} trends in both hemispheres (three models in CMIP5 and one model in CMIP6). The 5-member CMIP5

CanESM2 ensemble appears to show a significant equatorward ϕ_{UAS} trend in the NH, but the CanESM2-LE shows a nonsignificant trend. We calculate a 16% chance of finding a spurious statistically significant trend for a 5-member ensemble drawn from a pooled and detrended CESM1-LE and CanESM2-LE distribution. Other studies have shown the response to AA in a 5-member CanESM2 ensemble may not be representative of a SMILE ensemble (Oudar et al., 2018). In the SH, the CanESM2-LE does show a significant ensemble mean ϕ_{UAS} trend, but the magnitude found in the 5-member CMIP5 ensemble is around double the SMILE mean, showing this is not a good estimate of the forced response. Small ensembles can therefore give spurious estimates of forced HC width trends. If a forced trend is sufficiently large, it may be detectable in a small ensemble. For example, the SH poleward ϕ_{UAS} ensemble mean trend in NorESM2-LM lies outside the simulated piControl variability and the inter-ensemble range of both SMILEs, and is therefore likely to represent a forced response to AA. However, we cannot exclude the possibility that the relatively large SH poleward ϕ_{UAS} trends in FGOALS-g2 and ACCESS-CM2 are a consequence of single members sampling extreme states of IV (compare ACCESS-CM2 with ACCESS-ESM1-5).

The CMIP5/6 multi-model mean (MMM) ϕ_{UAS} trends represent a small residual averaged over large inter-model spread, with nonsignificant MMM ϕ_{UAS} trends in both hemispheres for CMIP5 and CMIP6. The overall tropical width contraction due to AA is $\leq 0.03°$ decade^{-1}, which is around an order of magnitude smaller than the estimated total historical tropical widening trend of 0.2° to 0.5° decade^{-1}. Based on the model simulations, AA is therefore a minor contributor to historical tropical width trends and may have made a small offset to the widening from IV, stratospheric ozone and GHGs (Grise et al., 2019). GD20 found a significant annual mean NH ϕ_{UAS} trend in CMIP5 models and a significant contraction in both hemispheres in CMIP5/6 measured by $\phi_{\Psi500}$, whereas these results show nonsignificant $\phi_{\Psi500}$ trends in both hemispheres in CMIP6 (Fig. S2d and S2g) demonstrating the MMM trend is highly sensitive to model and ensemble member selection. Other tropical width metrics show generally similar behaviour, with $\phi_{\Psi500}$ trends being mostly comparable to ϕ_{UAS}, although $\phi_{\Psi500}$ CMIP5 MMM trends are also significant in each hemisphere, and ϕ_{EDJ} showing smaller and nonsignificant trends in the SMILEs and CMIP5/6 (Fig. S2).

Given the limitations of small ensembles for identifying forced tropical width trends, we now assess the minimum ensemble size (N_{min}) required to detect a robust ϕ_{UAS} trend given a certain magnitude of IV. Figure 2 shows N_{min} for annual and seasonal trends computed using IV estimated from both SMILEs. To robustly detect an annual mean tropical width trend of the magnitude found in CanESM2 (~ -0.03° decade^{-1}) requires around 50 members. On seasonal timescales, where IV is larger, this would require >100 members. A forced seasonal tropical width trend would need to be at least 0.04° decade^{-1} in SON and 0.05° decade^{-1} in DJF to be robustly detected in the CanESM2-LE, and at least 0.06° decade^{-1} and 0.08° decade^{-1} in the 20-member CESM-LE. A forced annual mean trend in each hemisphere would need to be ≥|0.03° decade^{-1}| in CESM1-LE, and around half that in CanESM2-LE. The IV in the SMILEs is roughly equal to the median CMIP model. These findings reinforce that very large ensembles are required to isolate a forced signal that is small compared to IV and that most CMIP5/6 models do not provide sufficiently large ensembles to achieve this.

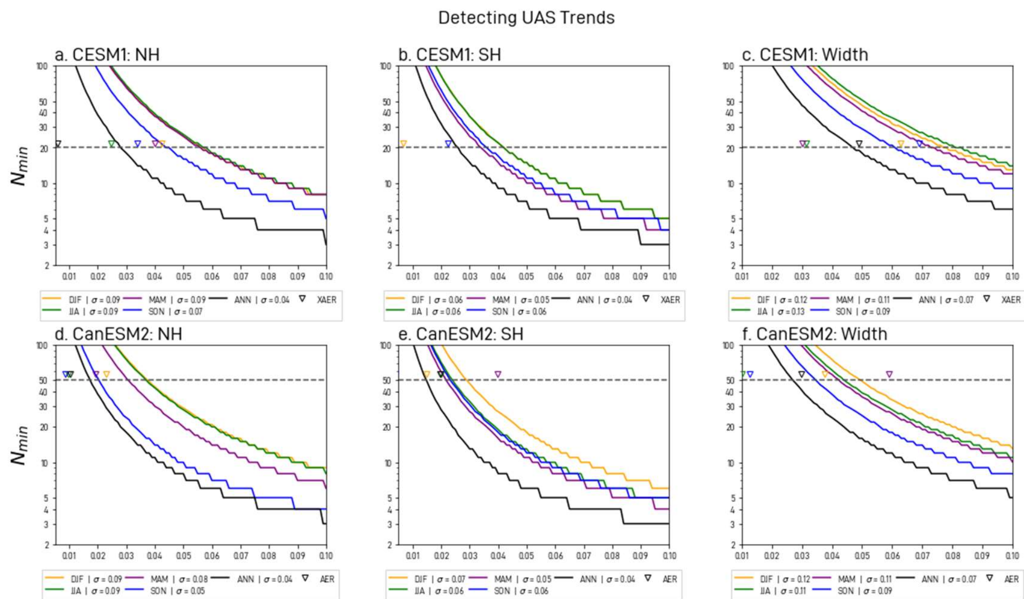

Figure 2: Minimum number of ensemble members (N_{min}; y-axis; note the log scale) required to detect a robust ϕ_{UAS} trend of a given magnitude (x-axis) for (a-c) CESM1-LE, (d-f) CanESM2-LE IV. N_{min} curves are shown for annual (black) and seasonal (colours - see legend). Horizontal dashed lines show the number of ensemble members available in the respective SMILE experiments, with ensemble mean trends for the respective experiments shown as symbols on the corresponding line.

Based on the above findings, we therefore conclude that the CMIP5/6 dataset is insufficient to identify whether climate models simulate robust tropical width trends due to AA between 1950-2005, and for distinguishing structural differences in responses between models since N_{min} for a difference between two forced responses is larger than for identifying a forced response in one model (McKenna and Maycock, 2021). In the remainder of the paper, we examine the SMILEs in more detail, where the ensemble mean trends give a more accurate estimate of the response to AA in those models.

4.4.2 Seasonal Tropical Width in SMILEs

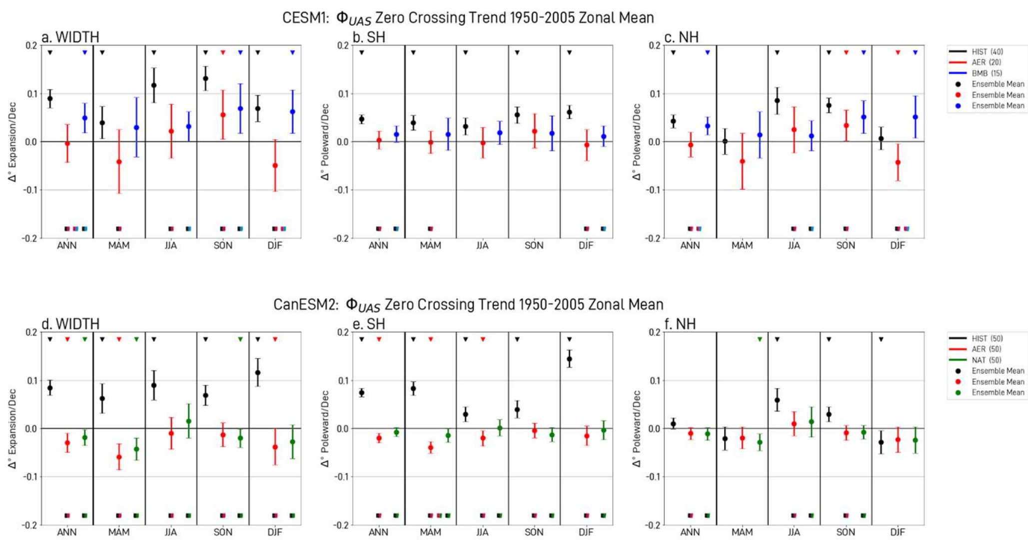

Figure 3: *Annual and seasonal mean ϕ_{UAS} trends for 1950-2005 in (a-c) CESM1-LE HIST (black), AER (HIST - XAER; red) and BMB (HIST - XBMB; blue), and (d-f) CanESM2-LE HIST (black), AER (red) and NAT (green), for (a, d) tropical width, (b, e) SH and (c, f) NH trends. Ensemble means are shown by large dots with 95% confidence intervals shown by whiskers. At the top, downward pointing triangles show where the ensemble mean is significant from zero. Coloured squares at the bottom indicate where experiment ensembles are significantly different from each other e.g. a red/black square indicates significant difference between HIST and AER distributions ($p < 0.05$).*

The large-scale circulation response to many external forcing agents exhibits seasonality, including CO_2 (e.g., Watt-Meyer et al., 2019), stratospheric ozone (e.g., Karpechko and Maycock et al., 2018), stratospheric water vapour (Maycock et al., 2013) and anthropogenic aerosols (Zhao et al., 2020). Figure 3 shows ensemble mean annual and seasonal mean tropical width, SH and NH ϕ_{UAS} trends over 1950-2005 ($\phi_{\Psi500}$ and ϕ_{EDJ} in Supplementary Figures S3 and S4). While the CESM1-LE AER experiment shows no significant annual mean trends in either hemisphere, there is a significant NH poleward shift in autumn (SON; 0.034°dec^{-1}) and a contraction in winter (DJF; -0.042°dec^{-1}) (Fig. 3c). The significant NH autumn poleward shift is distinct from studies using idealised, large amplitude global sulphate aerosol perturbations which show NH tropical contraction in all seasons (Zhao et al., 2020). In summary, the results suggest that AA contributes around half of the simulated historical ϕ_{UAS} trend in SON in the CESM1-LE and contributes substantially to overall seasonal differences in historical width trends.

The BMB experiment shows tropical expansion in the annual mean and all seasons except boreal spring (Fig. 3a), which comes mainly from poleward expansion in the NH (Fig. 3c). In NH winter, industrial and energy/transportation AA and biomass burning aerosols induce approximately equal and opposite ϕ_{UAS} trends thereby contributing to the lack of trend in HIST. In contrast, in boreal autumn BMB drives NH HC expansion, enhancing the AER trend. The results suggest biomass burning aerosol contributes around half of the total annual mean tropical widening trend in the CESM1-LE over 1950-2005. In boreal autumn, the sum of ensemble mean ϕ_{UAS} trends in AER and BMB is comparable to the total trend in HIST suggesting aerosols are a major driver of this NH expansion in CESM1-LE.

In contrast to the results for CESM1-LE, the CanESM2-LE AER experiment shows no significant NH trend in any season despite the larger ensemble size. This may be a consequence of structural differences between CESM1 and CanESM2 and, in some seasons, potential cancelling effects of industrial and biomass burning aerosol species from anthropogenic sources on tropical width as these are not separated in CanESM2-LE. In the SH, AER shows a significant contraction in austral autumn and winter (Fig. 3e). This is qualitatively consistent with the negative Southern Annular Mode trend due to AA found in CMIP5 models (Gillett et al., 2013); however, those models also showed SH contraction in austral summer which is not evident in CanESM2-LE.

Significant tropical width contraction is found in MAM and DJF in CanESM2-LE (Fig. 3d). Similar ensemble mean trends are found in these seasons in CESM1-LE (Fig. 3a), but they are nonsignificant owing to the smaller ensemble size compared to CanESM2-LE. There is therefore good qualitative agreement between the SMILEs, but a larger CESM1-LE ensemble would be required to deduce a robust AA-driven tropical width response in MAM and DJF across both models. The contraction is offset in CESM1-LE by a significant expansion trend in SON resulting in zero annual mean trend (Fig. 3a). In contrast, CanESM2-LE shows no such significant SON trend and therefore no offset of the MAM and DJF trends, resulting in a significant annual mean tropical contraction (Fig. 3d). This indicates that structural differences in the SON response are responsible for the structural differences in annual mean width response between the models.

In summary, the results show there are structural differences between the CESM1 and CanESM2 models in the zonal mean tropical width response to AA that arise from important differences in seasonality.

4.4.3 Regional Tropical Width in SMILEs

Aerosol forcing is spatially heterogeneous (Hoesly et al., 2018) and the dynamical processes that shape the circulation response to forcing differ regionally. This section examines basin-scale HC edge responses to AA and puts the zonal mean results in Section 3.2 into a regional context.

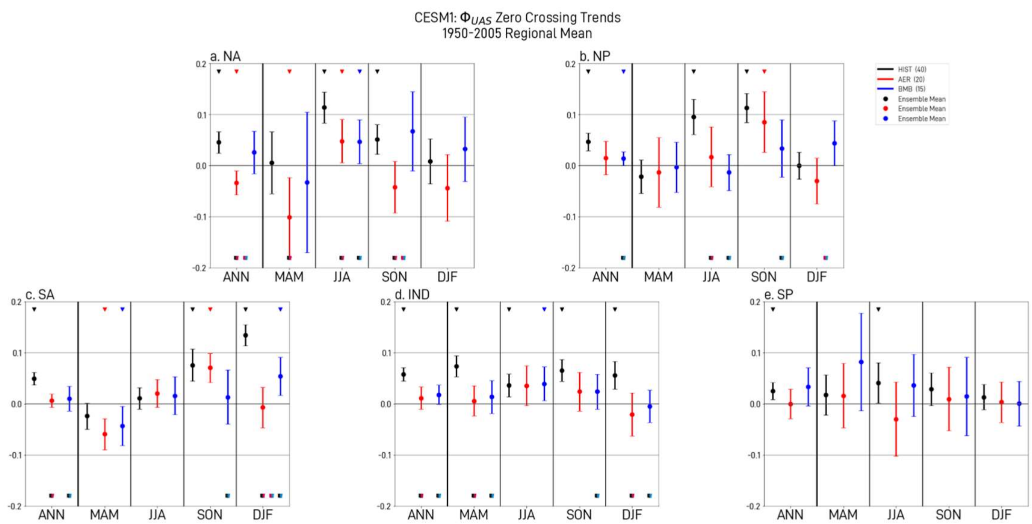

Figure 4: As in Figure 3, but for CESM1-LE regional mean ϕ_{UAS} trends. Northern hemisphere regions are a) North Atlantic (NA) and b) North Pacific (NP) and Southern hemisphere regions are c) South Atlantic (SA), d) Indian (IND) and e) South Pacific (SP). Basins defined as in Wood et al. (2020b).

Figure 4 shows ϕ_{UAS} trends in the CESM1-LE for the North Atlantic (NA), North Pacific (NP), South Atlantic (SA), Indian Ocean (IND) and South Pacific (SP) basins (see Methods). Corresponding trends in $\phi_{\Psi500}$ and ϕ_{EDJ} are shown in Supplementary Information Figures S5 and S6. The NH zonal mean shows no ϕ_{UAS} trend due to industry/energy aerosols (Fig 3c). However, the regional perspective shows a significant contraction in the NA (Fig. 4a), which is opposite to the trend in the historical experiment, indicating AA offset part of the NA ϕ_{UAS} expansion from other drivers in CESM1-LE. The significant annual mean NA contraction in AER shows seasonal variation, with relatively large contraction in MAM (-0.1° decade^{-1}) opposed by a smaller expansion in JJA. In the CanESM2-LE (Fig. 5a), there is also an equatorward trend in ϕ_{UAS} in the NA in XAER. However, the seasonality is different with a contraction in boreal summer, opposite to CESM1-LE. Therefore, even if the models agree in their annual mean NA response, they do not agree on the seasonality indicating model structural differences. There are also model structural differences in the NP, where CESM1-LE shows a significant poleward ϕ_{UAS} trend in SON in the historical experiment, with a substantial contribution from energy/industry aerosols (Fig. 4b), and CanESM2-LE shows a smaller historical ϕ_{UAS} trend in SON by around a factor of 6 (Fig. 5b), and a nonsignificant effect from AA.

In the SH, CESM1-LE shows no significant annual mean zonal mean ϕ_{UAS} trends in any experiment (Fig. 2) and this is also the case for individual basins (Fig. 4c-e). In the SA, this masks opposing significant trends in MAM (equatorward) and SON (poleward). CanESM2-LE does not show an equatorward SA ϕ_{UAS} trend in MAM, but like CESM1-LE shows a poleward trend in SON. CanESM2-LE shows significant contraction in the SP throughout the year which is opposite to the historical trends, with both having relatively larger amplitudes. In contrast, in CESM1-LE neither the historical nor AER simulations show significant SP ϕ_{UAS} trends year-round. SH regional heterogeneity in historical low-level zonal wind trends has been hinted at in previous studies (e.g. see Fig. 3 in Waugh et al., 2020). These results indicate a potentially significant role for industrial/transport AA in this difference.

In BMB, both hemispheres show a significant poleward trend in the annual mean (Fig. 2). In the NH, regional averaging shows this is only significant in the NP, but as this trend is small, the zonal mean significance is likely formed from the residual of NP and NA trends. There is also a significant BMB trend in JJA in NA, with no trend in NP. The BMB and AER NA summer trends are both roughly half the overall HIST trend, implying the combined aerosol forcing contributed substantially to the historical summer poleward trend in that region.

In the SH there are significant BMB trends in SA in MAM (equatorward) and DJF (poleward), and IND in JJA. The BMB effect on summer SA ϕ_{UAS} makes the HIST trend significant versus IV, implying that while historical ozone forcing is accepted as the largest contributing factor to overall summertime poleward HC expansion

(Karpechko and Maycock et al., 2018), there may also have been a significant contribution from BMB aerosols in the SA.

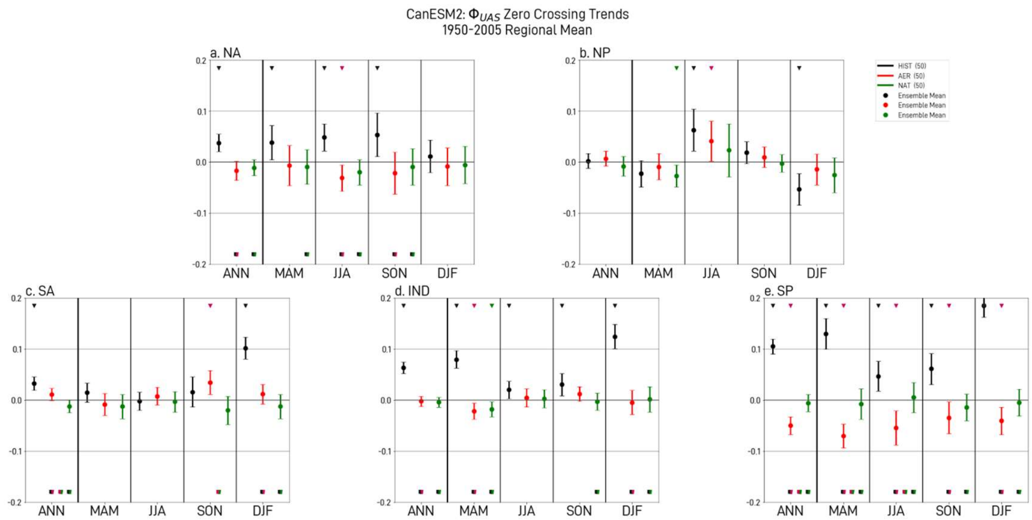

Figure 5: As in Figure 4, but for CanESM2-LE.

4.5 Discussion and Conclusions

This study has assessed the impact of anthropogenic aerosols (AA) on late 20th-century tropical width trends in CMIP5/6 and two SMILE models. Our results advance the understanding of AA forced tropical width trends in the following ways:

The results show that, in most cases, the ensemble size of historical AA-only experiments in CMIP5/6 are too small to detect a late 20th-century forced trend in tropical width and to differentiate forced HC edge trends between models. Even in the one third of CMIP5/6 models that show a trend that is significant from zero, there is little agreement in the trend and significance of trends by hemisphere. Our analysis of two large ensembles that isolate AA forcing shows the tropical width trend estimated from a small ensemble can be biased due to large internal variability. We show the multi-model mean CMIP5/6 tropical widening response to AA is highly sensitive to model selection, with our analysis showing nonsignificant responses while previous work has shown statistical significance (Grise and Davis, 2020).

The two large ensembles show different effects of AA on late 20th-century zonal and annual mean tropical width trends, indicating structural differences in the response to aerosols between the models. CESM1 shows no zonal mean tropical width trend due to AA, whereas CanESM2 shows a contraction of around 0.03°dec^{-1} which comes mainly from the Southern hemisphere. In CanESM2 the AA-driven contraction can be compared to the overall tropical expansion in the historical experiment (0.08°dec^{-1}).

We show potentially important regional and seasonal tropical width responses to AA in the SMILEs. In the NH, both CESM1 and CanESM2 show HC contraction in boreal spring and winter. CESM1 also shows HC expansion in autumn which is absent in CanESM2 and cancels the contraction shown in the aforementioned seasons resulting in a zero annual mean trend. In CESM1 this AA-driven expansion contributes around half of the simulated historical ϕ_{UAS} trend in SON.

There are also regional differences in HC edge trends, especially in the NH, such as a contraction in the North Atlantic annual mean that offsets tropical expansion forced by other drivers. In the SH, we show that the zonal mean masks regional seasonal trends, such as opposing spring and autumn trends in the South Atlantic that offset each other in the annual mean, and significant equatorward trends in the South Pacific throughout the year in CanESM2. These South Pacific trends are not found in CESM1, again indicating structural differences between the models.

We show that an ensemble size of around 50 is required to robustly identify a typical tropical width trend of the magnitude found in CMIP5 and CanESM2 (~0.03°dec^{-1}). Seasonal trends of similar magnitude require even larger ensemble sizes due to larger seasonal IV.

Future aerosol trends are uncertain and different socio-economic pathways show different aerosol trends during this century (e.g. Gidden et al., 2019); therefore, the role of AA for future tropical width trends will be scenario-dependent. We also identify structural differences between the representation of AA-induced trends that contribute to modelling uncertainty.

The results support other recent work showing the small ensemble sizes available in CMIP5/6 models have limited use in both robustly identifying externally forced responses that are small relative to IV, and diagnosing sources of structural uncertainty between models (McKenna and Maycock, 2021). Other studies that have investigated model structural differences using CMIP5/6 models should therefore be viewed with caution (e.g., Steptoe et al., 2016). The large ensembles available indicate important regional AA impacts which should be investigated further. For example, our results add to evidence of SH regional heterogeneity in low-level zonal wind trends (e.g., Waugh et al., 2020) and indicate a potentially significant role for AA in this difference. The results motivate the further production of single forcing large ensembles to isolate the contribution of drivers to historical climate and to robustly identify structural differences in responses between models.

Acknowledgements

TW was funded by a PGR studentship from the NERC SPHERES Doctoral Training Partnership (NE/L002574/1). ACM was funded by the European Union's Horizon 2020 Research and Innovation Programme under grant agreement No. 820829 (CONSTRAIN project) and the Leverhulme Trust. We acknowledge the World Climate Research Programme, which, through its Working Group on Coupled Modelling, coordinated and promoted CMIP5 and CMIP6. We thank the climate modelling groups for producing and making available their model output, the Earth System Grid Federation (ESGF) for archiving the data and providing access, and the multiple funding agencies who support CMIP6 and ESGF. We also thank CEDA and JASMIN for archiving CMIP data and providing access to high-performance computing facilities. We acknowledge the CESM1-CAM5 modelling group and the CanESM2 modelling group and in particular those involved in the Single Model Initial-condition Large Ensemble (SMILE) project for making their data openly available. We are grateful to Dr Christine McKenna for useful comments on an earlier version of this manuscript and to Dr Kevin Grise and Dr Sean Davis for sharing data that were useful in the initial stages of this study.

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Chapter 5

Conclusions and Recommendations

5.1 Summary and Conclusions

It is well understood that the large-scale circulation responds to external forcing. Many studies exist regarding the modelled SH midlatitude circulation response to increased atmospheric CO_2 concentrations. However, large uncertainties in the quantification of the response in GCMs still exist, inhibiting the ability to accurately reproduce observed historical circulation shifts and make projections of future changes under anthropogenic forcing scenarios. The relative role of model structural differences and internal variability in the overall uncertainty, as well as sources of such uncertainty, are open research questions. Moreover, the midlatitude circulation response to non-CO_2 forcing is even less certain, with relatively few studies addressing the impact of different types of anthropogenic aerosols in particular. The relative roles of direct atmospheric rapid adjustments and sea-surface temperature driven feedbacks in these responses is also not well understood. There is also a lack of understanding of the tropical width response to anthropogenic aerosols with the detection of a forced signal amidst relatively large internal variability a persistent problem. Furthermore, the large majority of studies of tropical width focus on the zonal average response to forcing, potentially missing important regional changes that could drive substantial localised climate changes. This thesis has addressed some of these open questions and advanced the understanding of the large-scale circulation response to external forcing.

Chapter 2 addressed the relative roles of rapid adjustments and SST-mediated feedbacks for the SH midlatitude circulation response to doubled CO_2 (2xCO2), tripled methane (3xCH4), a five-fold increase in sulphate aerosols (5xSO4), a ten-fold increase in black carbon aerosols (10xBC) and a 2% increase in the solar constant (2%Sol), particularly focusing on the seasonal response of the SH EDJ. The results showed that in austral summer (DJF), the magnitude of the rapid adjustment component of the EDJ shift is around 20-30% of the fully coupled response for most of the perturbations (other than 10xBC), corroborating results for CO_2 forcing found in previous studies (e.g., Grise and Polvani, 2014) and showing that a similar magnitude of rapid adjustment is mediated by methane, solar and sulphate forcing despite differences in governing radiative mechanisms (e.g. the extent of stratospheric heating or cooling). These findings showed that while SST-mediated feedbacks are the most important component, rapid adjustments also contribute a substantial fraction of the midlatitude circulation response. However, the MMM response in DJF was shown to be small compared to internal variability, casting doubt on whether a rapid adjustment signal in the SH midlatitude circulation could be observed. Despite the small overall signal, the results showed strong sign agreement across models, indicating that the responses are unlikely to be a result of internal variability alone. The response to 10xBC was markedly different to the other forcings examined, with the rapid adjustment found to be around 75% of the coupled EDJ shift, owing primarily to strong, direct localised tropospheric heating anomalies mediated by shortwave absorption by BC. The results therefore showed that aerosol forcing could influence SH midlatitude circulation changes in projected emissions scenarios, despite the majority of the forcing being concentrated in the NH, although large uncertainty regarding this effect remains.

Further analysis in Chapter 2 showed that up to 80-90% of the intermodel spread in local midlatitude 850-hPa zonal wind responses (U_{850}) in DJF can be explained by the spread in the lower tropospheric tropics-to-pole temperature gradient responses across the models. This finding was true of all five perturbations, indicating that the response of meridional temperature gradients is an important source of uncertainty in lower-troposphere zonal wind responses. By showing that the gradient of the relationship between zonal winds and meridional temperature gradient for the rapid adjustment was around double that for the coupled experiment, the results also suggest that processes captured by the rapid adjustment, such as changes to clouds and land surface temperatures, mediate effects on the midlatitude circulation in a manner that cannot be quantitatively captured by simple large-scale temperature gradient change measures, and demonstrate why ECS is not generally a useful measure for determining regional midlatitude climate change.

Chapter 3 built on the work of Chapter 2 by further exploring the effects of evolving SST patterns on the SH midlatitude circulation. In Chapter 3, the SH EDJ response to 4xCO2 was compared in the CMIP5 and CMIP6 generations of GCMs. Corroborating contemporary findings by Curtis et al. (2020), the results revealed a muted MMM poleward SH EDJ shift in CMIP6 compared to the previous generation, despite there being a higher average ECS than in CMIP5. This provided further evidence that ECS is not a useful measure for constraining midlatitude circulation responses. Extending Curtis et al. (2020), the study also showed that, as found for CMIP5 in Ceppi et al. (2018), the majority of the poleward EDJ shift occurs in the first decade after forcing is applied in CMIP6. Importantly, the majority of the difference between CMIP5 and CMIP6 EDJ responses also occurred in this 'fast' period. Motivated by the importance of evolving SST responses for the SH EDJ shift shown by Ceppi et al. (2018), new fixed-sea surface temperature (fSST) simulations were executed in an intermediate complexity climate model (IGCM4), with CMIP5 and CMIP6 multi-model mean 'fast' SST warming patterns imposed, both globally and for the SH midlatitudes only. The results showed that in both sets of experiments, IGCM4 was able to quantitatively reproduce many of the differences in U_{850} responses between CMIP5 and CMIP6, with a significant pattern correlation between the difference in U_{850} responses in IGCM4 and the difference in U_{850} responses between CMIP5 and CMIP6 in three SH ocean basins (Atlantic, Indian and Pacific) in most seasons. The findings demonstrated that differences in SST patterns can explain a substantial fraction of regional differences in the response of the SH midlatitude circulation to CO_2 forcing between CMIP5 and CMIP6, and thus provided an alternative hypothesis to that forwarded by Curtis et al. (2020), who attributed the difference in extended austral winter responses to a reduction in the equatorward bias in climatological EDJ position (Bracegirdle et al., 2020; De et al., 2021), as found in successive CMIP generations (e.g. Kidston and Gerber., 2010; Ceppi et al., 2012), which may be an emerging constraint on midlatitude circulation responses (e.g. Simpson et al., 2021). While it is possible that part of the extratropical dipole pattern of SST difference between CMIP5 and CMIP6 models could be driven by differences in the surface winds themselves (Ferreira et al., 2015), further analysis showed that the magnitude of the difference in SST patterns was larger than could be explained by surface wind anomalies alone, and therefore that the difference in SST warming patterns was likely to be related to other processes such as ocean circulation responses or Southern Ocean cloud feedbacks, which have been shown to also be important for differences in ECS between the model generations (Zelinka et al., 2020).

Motivated by the findings in Chapter 2 indicating that anthropogenic aerosols (sulphates and black carbon) can mediate changes to the large-scale circulation, Chapter 4 examines the contribution of AA to late-20th century (1950 - 2005) tropical width trends in the CMIP5 and CMIP6 models and in two Single Model Initial Condition Large Ensembles (SMILEs; Deser et al., 2020) - CESM1 and CanESM2. This study expanded on work by Grise and Davis (2020) and other studies (e.g. Allen et al., 2012; Gillett et al., 2013; Steptoe et al., 2016; Zhao et al., 2020) suggesting a role for AA in changes to tropical width. Chapter 4 also examined whether it is possible to detect a tropical width response to AA in CMIP5 and CMIP6 models, and structural differences between the modelled responses, given their small ensemble sizes and magnitude of internal variability, motivated by previous and contemporary studies utilising large ensembles to examine large-scale circulation responses (e.g., Deser et al., 2012, McKenna and Maycock, 2021). The results showed that in most cases, ensemble sizes in the CMIP5 and CMIP6 models are too small to detect a robust late 20th-century forced trend in tropical width and diagnose structural differences in response between models. Moreover, trends in the small CMIP ensembles can be biassed due to relatively large internal variability, while statistical significance of MMM responses are highly sensitive to model selection. The results showed that, on the timescale considered here, an ensemble size of around 50 members is required to robustly identify a tropical width trend of a typical magnitude found in CMIP5 and the CanESM2 SMILE ($\sim 0.03° \text{ dec}^{-1}$), while even larger ensembles are required to detect seasonal trends due to larger seasonal internal variability.

Comparing the two SMILEs did reveal structural differences in annual mean tropical width trends, but these could only be distinguished with the large ensemble sizes. Furthermore, potentially important regional and seasonal AA-forced tropical width trends were found. For example, in the NH both models showed a contraction in boreal winter and spring. In CESM1, a tropical expansion in boreal autumn, which is around half of the overall historical trend, cancels the contraction in the aforementioned seasons, resulting in a net zero annual mean trend. This was not found in CanESM2, however, indicating an important structural difference between the models. Regional differences in tropical width trends included a contraction in the North Atlantic annual mean that offsets tropical expansion forced by non-AA drivers, and regional seasonal trends in the SH, such as opposing spring and autumn trends in the South Atlantic, that are masked in the zonal mean. Structural differences between models were also indicated in the regional responses. For example, significant trends were found throughout the year in the South Pacific in CanESM2, but not in CESM1.

The results of Chapter 4 are an important consideration when drawing conclusions from the results of Chapters 2 and 3. Chapter 2 suggests a role for black carbon and sulphate aerosols in mediating a shift of the SH midlatitude circulation. However, the results should be read as a first investigation of the effect of these forcings, with the large idealised perturbations indicating the qualitative response of the SH midlatitude circulation, and providing a quantitative estimation of the relative responses, based on evidence such as the degree of consistency regarding the sign of the modelled response and the MMM. The study therefore motivates further investigation of the effects of these forcings in more realistic single forcing historical ensembles, which would complement and extend these results. Caution should always be exercised when using MMMs in model intercomparison studies to provide 'best guess' estimates of the forcing response (Knutti et al., 2010), as demonstrated by the sensitivity of the MMM to model selection shown in Chapter 4. These issues are primarily related to small ensemble sizes, in terms of both the error on the mean associated with the number of models included in the calculation, i.e. whether this number is large enough to ensure that models which have by chance sampled extreme states of internal variability and are therefore biased do not overly skew the mean; and also the number of ensemble members within each model, i.e. whether the number of ensemble members is large enough to ensure the ensemble mean captures the true forced signal versus internal variability. Chapter 4 suggests that CMIP model ensembles do not meet this threshold for the assessment of small forced signals relative to the magnitude of internal variability present in several large-scale circulation measures. Therefore, previous studies that have investigated structural differences among CMIP5 or CMIP6 models should be viewed with caution (e.g., Steptoe et al., 2016). That being said, in regards to the anthropogenic aerosol forced response of the SH midlatitude circulation, the balance of multiple lines of evidence presented in this thesis, including idealised abrupt forcing experiments (Chapter 2) and single forcing historical attribution experiments, both in CMIP5 and CMIP6, and in two SMILE models (Chapter 4), points towards a contribution of AA on both the SH EDJ and in the SH HC width, adding to evidence of such an effect in previous studies (e.g., e.g. Gillett et al., 2013; Rotstayn, 2013; Steptoe et al., 2016; Wang et al., 2020). The evidence for a contribution from AA is especially strong regionally and seasonally, adding to previous evidence of the importance of regional heterogeneity in SH low-level zonal wind trends (e.g. Bracegirdle et al., 2013; Waugh et al., 2020;

Patterson et al., 2020; Yang et al., 2020, 2021; Diao et al., 2021) and pointing to the need to more carefully address these regional responses in future work. Some recent studies, such as Yang et al. (2021), have begun to further develop understanding in this direction. However, although one might conclude that it is likely that anthropogenic aerosol forcing has affected historical SH large-scale circulation trends and will play a role in future changes, especially regionally, due to the large expected changes in anthropogenic aerosol emissions over the next decades (Gidden, 2019), the magnitude and robustness of these responses, and their relative importance versus the role of internal variability on various temporal and spatial scales, remains an open question and a key challenge in climate science. Caution must also be exercised when using MMMs for the assessment of regional responses, due to the danger of averaging out important regional features across multiple models, highlighted by Knutti et al. (2010) and Zappa (2019). Caution should also be exercised when assessing the potential structural differences in regional tropical width trends identified between the SMILE models, given that only two models are currently available in the ensemble and there are large uncertainties between models in their representation of aerosol forcing.

Both Chapter 2 and Chapter 3 identified a prominent role for SST warming patterns in the SH midlatitude response to forcing, and Chapter 3 identified that the response in the 'fast' period is crucial. These findings corroborate previous work by e.g. Grise and Polvani (2014) and Ceppi et al. (2018). Chapter 2 also indicated a substantial role for rapid adjustments in the SH EDJ response to forcing, especially for the response to black carbon. However, Chapter 3 showed that there was only a small difference in rapid adjustments between CMIP5 and CMIP6 models, and this does not account for the differences in zonal wind responses between the ensembles, adding to evidence that SST responses are dominant in this difference. As discussed above, the regional heterogeneity of the SH circulation response to forcing is also of critical importance, and the role of SST-driven feedbacks in regional SH midlatitude zonal wind responses was further shown in Chapter 3. However, the results of Chapter 3 are based on a small number of prescribed-SST simulations performed by IGCM4, and thus the results may suffer from similar issues regarding the detection of signals versus internal variability as in Chapter 2, although internal variability is inherently reduced when SSTs are fixed.

5.2 Recommendations for Future Work

Whilst this thesis has progressed the understanding of the large-scale circulation response to external forcing, many questions remain. Here, recommendations are made for potential future research.

- It has been shown that small ensemble sizes are unable to robustly capture large-scale circulation responses to individual climate forcings (or groups of forcings) that may be small relative to internal variability, but are crucial to understanding historical trends and for projecting changes under future emissions scenarios. Therefore, the lack of models that have run single forcing large ensembles hampers the ability to make quantitative assessments of large-scale circulation responses to external forcings - especially to anthropogenic aerosol forcing, which has particularly large associated uncertainties - and the separation of the roles of forced signals and internal variability. The lack of single forcing large ensembles also hampers the ability to make robust assessments of uncertainties related to structural differences between GCMs, impeding model development. Thus, the reduction of the uncertainty in those responses is currently very difficult. Although there exists a suite of models that have run large ensembles of historical all-forcing experiments (the multi-model large ensemble archive (MMLEA) project (Deser et al., 2020)), and two of these models (CESM1 and CanESM2) have run large ensembles with anthropogenic aerosol forcing separated from e.g. GHGs, these experiments combine the forcing from multiple aerosol species that have been shown to mediate different large-scale circulation responses (Chapter 2), each with their own inherent uncertainties. Large ensembles of idealised, single forcing experiments, of the type performed as part of PDRMIP (Chapter 2), do not currently exist. Due to the complexity of the large-scale circulation system, methods to mitigate small ensemble sizes such as climate model emulators or emerging artificial intelligence and deep learning techniques have limited application in identifying sources of uncertainty and quantifying responses of large-scale circulation measures. Therefore, to move forward, large ensembles of idealised single forcing simulations must be performed. As recommended in Chapter 4, the ensemble sizes should be at least 50 and ideally larger (100+) in order to investigate signals on seasonal timescales where internal variability is larger. Following the principles employed by PDRMIP, the idealised perturbations should aim to produce roughly equivalent ERFs to allow responses to different forcings to be compared. This would improve understanding considerably and facilitate a much deeper understanding of the relative roles of different forcings in the climate system versus internal variability, enabling statistically robust conclusions that are currently lacking. This suggestion echoes the recent 'Lighthouse Activity Science Plan' by the World Climate Research Programme (WCRP), published in June 2021 (WCRP, 2021). The second point in this plan is to:

> *"Build an integrated operational capability to attribute and predict multi-annual to decadal changes in the climate system and provide quantitative attribution statements to support WMO forecasts and State of the Climate reports … a key component [of which] will be* ***large ensemble single forcing simulations*** *of the historical period. These do not currently exist…". (WCRP, 2021)*

- The results in Chapter 3 are also based on a small number of simulations executed in IGCM4. The robustness of these results could be improved by also running large ensembles of prescribed-SST experiments (although internal variability is inherently reduced by fixing SSTs), with SSTs in different ocean basins isolated to investigate the effect of regional SST anomalies. These experiments could further improve the understanding of the source of differences between CMIP model generations, particularly considering the alternative hypotheses regarding the source of this difference forwarded by Curtis et al. (2020). It is possible that both suggested effects are important and the two hypotheses may therefore be complementary. If so, an investigation of the relative importance of these effects in constraining SH midlatitude circulation responses is necessary.

- Chapter 3 speculated that there may be processes related to ocean circulation responses and/or Southern Ocean cloud feedbacks that are influencing differences in patterns of SST warming responses to CO_2 forcing between the CMIP5 and CMIP6 generation of GCMs, that were found to explain a substantial fraction of regional differences in SH midlatitude circulation responses. Given the importance of cloud radiative effects on the SH midlatitude circulation (e.g., Ceppi et al., 2012; 2014; Ceppi & Shepherd, 2017; Grise and Kelleher, 2021; Voigt et al., 2021) and considering that differences in Southern Ocean cloud feedback have been shown to contribute to differences in ECS between the model generations (Zelinka et al., 2020), it is plausible that similar phenomena also influence the differences in SH midlatitude circulation responses shown in Chapter 3 (Grise and Kelleher, 2021). Future work could investigate these effects further, for example by using a cloud locking method to isolate the role of cloud-radiative interactions (e.g., Albern et al., 2019). This would help elucidate reasons for the differences between model generations and the development of emergent constraints that could reduce uncertainty of SH midlatitude circulation projections.

- Chapters 2 and 3 suggested a significant role for midlatitude SSTs for determining the response of the SH midlatitude circulation to forcing. Given the importance of changes to midlatitude SST fronts to the midlatitude circulation in the SH, particularly seasonally, due to their influence on baroclinicity (e.g. Nakamura, 2004; Nakamura and Shimpo, 2004; Ogawa et al., 2016), future work could develop understanding by examining the dependence of SH EDJ strength and positional changes on changes to regional and seasonal SST fronts. These future studies could involve experiments whereby either observed, prescribed or idealised SST fronts (or a combination) are imposed into an intermediate complexity climate model, building on previous studies that have used more simple aquaplanet models (e.g., Ogawa et al., 2016). However, the model resolution would need to be fine enough to resolve the sharp meridional temperature gradients that are characteristic of SH midlatitude SST fronts. This study would build on the findings of Chapter 3 in particular.

- Chapter 2 highlighted the importance of upper and lower tropospheric meridional tropics-to-pole temperature gradients in the SH EDJ response to forcing. Changes to gross static stability have also been implicated in both midlatitude circulation responses and tropical widening (e.g. Allen et al., 2012; Zhao et al., 2020). The relative importance of regional atmospheric heating anomalies in these effects is an open question. Chapter 2 showed that localised tropospheric heating anomalies mediated by shortwave absorption by black carbon is associated with a significant poleward shift of the SH EDJ. To further investigate these effects, idealised tropospheric localised heating anomalies could be imposed into an intermediate complexity model, building on work by e.g. Butler et al. (2010). This type of study could also make use of atmospheric nudging methods.

- While most existing studies of tropical width tend to focus on zonal mean measures, there is a growing emphasis on the importance of regional Hadley cell responses to external forcing (e.g. Grise et al., 2018; Staten et al., 2019; Staten et al., 2020a; Martin et al., 2020). Chapter 4 showed potential regional tropical widening mediated by anthropogenic aerosols. A deeper understanding of regional Hadley cell responses could be developed by examining the externally forced response of locally partitioned Hadley cell measures (Schwendike et al., 2014). Some recent studies (e.g. Hurr et al., 2021) have begun to take this approach.

Chapter 5 References

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Appendix A

Supplementary Information for Chapter 2:

Wood et al. (2020a)

The Southern Hemisphere Midlatitude Circulation Response to Rapid Adjustments and Sea Surface Temperature Driven Feedbacks

Published in Journal of Climate (2020)

Appendix B

Supplementary Information for Chapter 3:

Wood et al. (2020b)

Role of Sea Surface Temperature Patterns for the Southern Hemisphere Jet Stream Response to CO2 Forcing

Appendix C

Supplementary Information for Chapter 4:

Wood and Maycock (2022)

Aerosol-forced tropical expansion in the late 20th century: robust anthropogenic signal or internal variability?

Supplementary Tables

CMIP5		CMIP6	
Model	**Ensemble Members**	**Model**	**Ensemble Members**
CanESM2	r1i1p4 r2i1p4 r3i1p4 r4i1p4 r5i1p4	ACCESS-CM2	r1i1p1f1
CSIRO-Mk3-6-0	r1i1p4 r2i1p4 r3i1p4 r4i1p4 r5i1p4	ACCESS-ESM1-5	r1i1p1f1 r2i1p1f1 r3i1p1f1
GFDL-ESM2M	r1i1p5	BCC-CSM2-MR	r1i1p1f1 r2i1p1f1 r3i1p1f1
GISS-E2-H	r1i1p107 r2i1p107 r3i1p107 r4i1p107 r5i1p107	CanESM5	r1i1p1f1 r2i1p1f1 r3i1p1f1 r4i1p1f1 r5i1p1f1 r6i1p1f1 r7i1p1f1 r8i1p1f1 r9i1p1f1 r10i1p1f1
GISS-E2-R	r1i1p107 r2i1p107 r3i1p107 r4i1p107 r5i1p107	CESM2	r1i1p1f1 r3i1p1f1
NorESM1-M	r1i1p1	CNRM-CM6-1	r1i1p1f2 r2i1p1f2 r3i1p1f2 r4i1p1f2 r5i1p1f2 r6i1p1f2 r7i1p1f2 r8i1p1f2 r9i1p1f2 r10i1p1f2
FGOALS-g2	r2i1p1	FGOALS-g3	r1i1p1f1
IPSL-CM5A-LR	r1i1p3	GFDL-ESM4	r1i1p1f1
GFDL-CM3	r1i1p1 r3i1p1 r5i1p1	GISS-E2-1-G	r1i1p1f1 r2i1p1f1 r3i1p1f1 r4i1p1f1 r5i1p1f1
CCSM4	r1i1p10 r4i1p10 r6i1p10	HadGEM3-GC31-LL	r1i1p1f3 r2i1p1f3 r3i1p1f3 r4i1p1f3
CESM1-CAM5	r1i1p10 r2i1p10 r3i1p10	IPSL-CM6A-LR	r1i1p1f1 r2i1p1f1 r3i1p1f1 r4i1p1f1 r5i1p1f1 r6i1p1f1 r7i1p1f1 r8i1p1f1 r9i1p1f1 r10i1p1f1
		MIROC6	r1i1p1f1 r2i1p1f1 r3i1p1f1
		MRI-ESM2-0	r1i1p1f1 r2i1p1f1 r3i1p1f1 r5i1p1f1
		NorESM2-LM	r1i1p1f1 r2i1p1f1 r3i1p1f1

Table S1: CMIP5 and CMIP6 models and their anthropogenic aerosols forcing experiment members.

Supplementary Figures

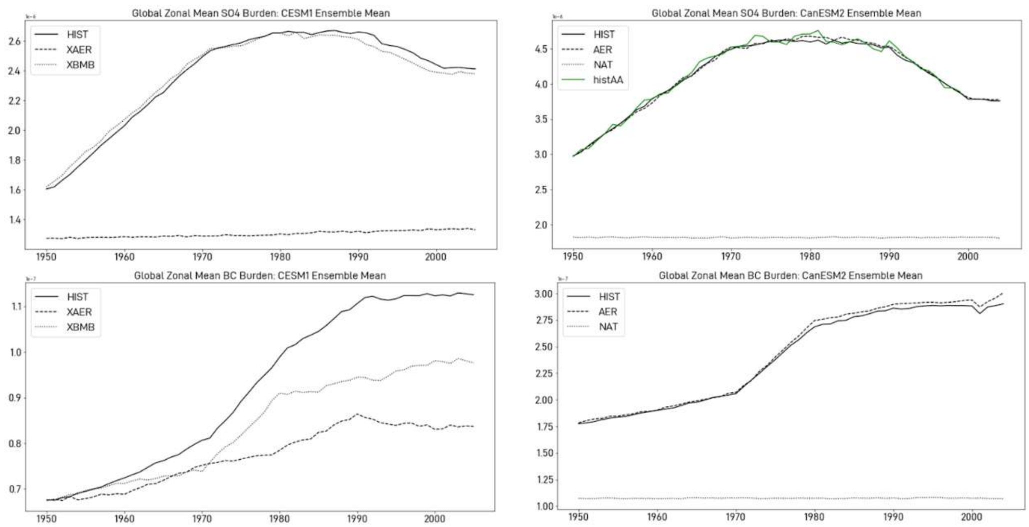

Figure S1: Global mean SO_4 and black carbon burdens for CESM1-LE and CanESM2-LE for the experiments analysed in this study.

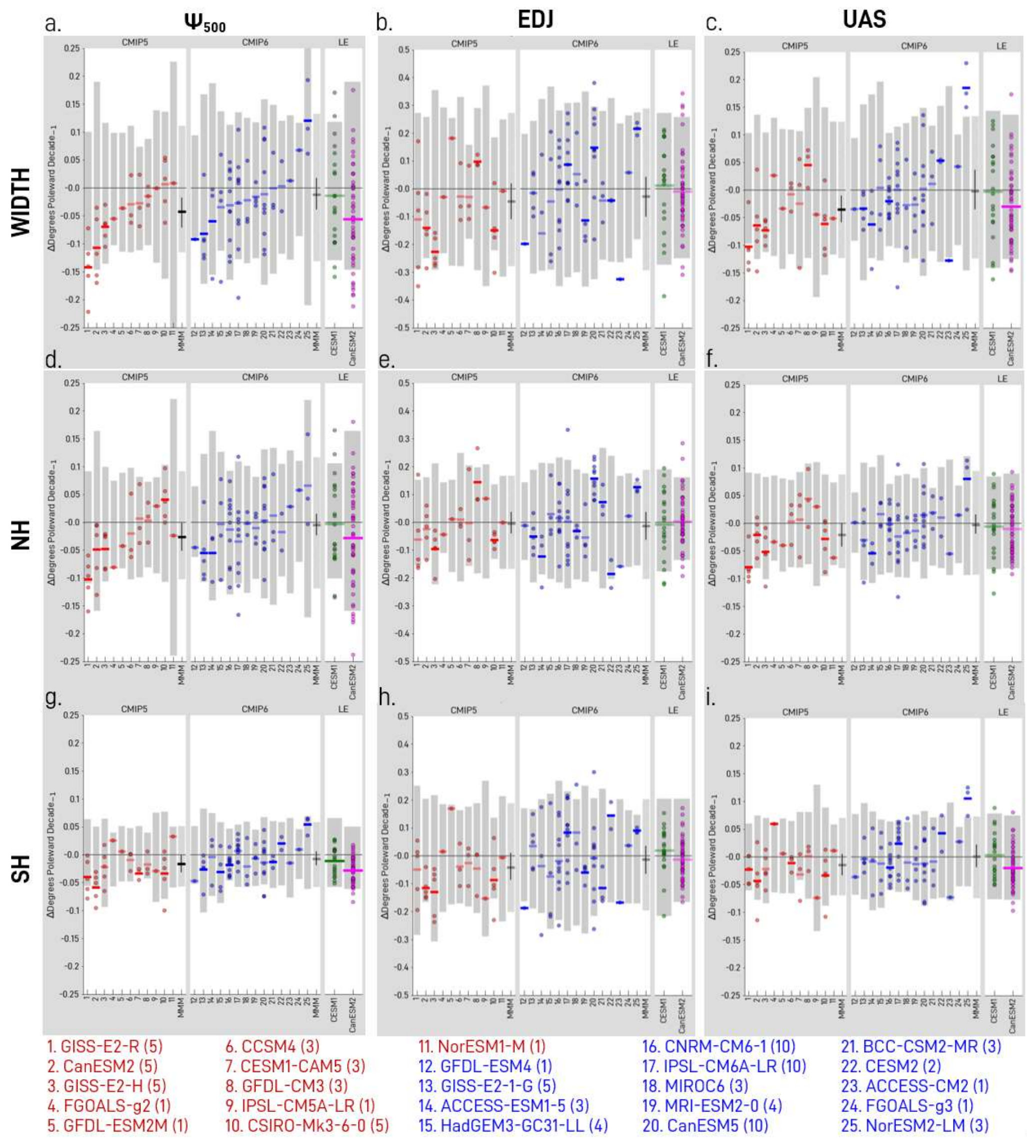

Figure S2: Zonal and annual mean tropical width metrics ($\phi_{\Psi500}$, ϕ_{EDJ} and ϕ_{UAS}) trends for the period 1950-2005 in the histAA, histAER, HIST - XAER and AER experiments for CMIP5 (red), CMIP6 (blue), CESM1-LE (green) and CanESM2-LE (magenta) respectively for (a-c) Width, (d-f) Northern Hemisphere (NH) and (g-i) Southern Hemisphere (SH). Grey bars show the 2.5 to 97.5 percentile range (i.e. the 95% confidence interval (CI)) of 55-year piControl trends i.e. the internal variability (IV) in each model. Circles show individual ensemble members in each model, the number of which is shown in brackets adjacent to the respective numbered model names at the bottom. Coloured horizontal lines show ensemble means. Ensemble mean lines are solid if the ensemble mean anomaly is statistically significant (*see Methods*), and semi-transparent otherwise. Black horizontal lines adjacent to the CMIP5 and CMIP6 models show the multi-model mean (MMM), which are solid where the MMM is significant from zero, with CI shown by vertical black error bars. The median IV of the respective CMIP ensembles is shown as a grey bar in the MMM column. Note the different scale in the EDJ column.

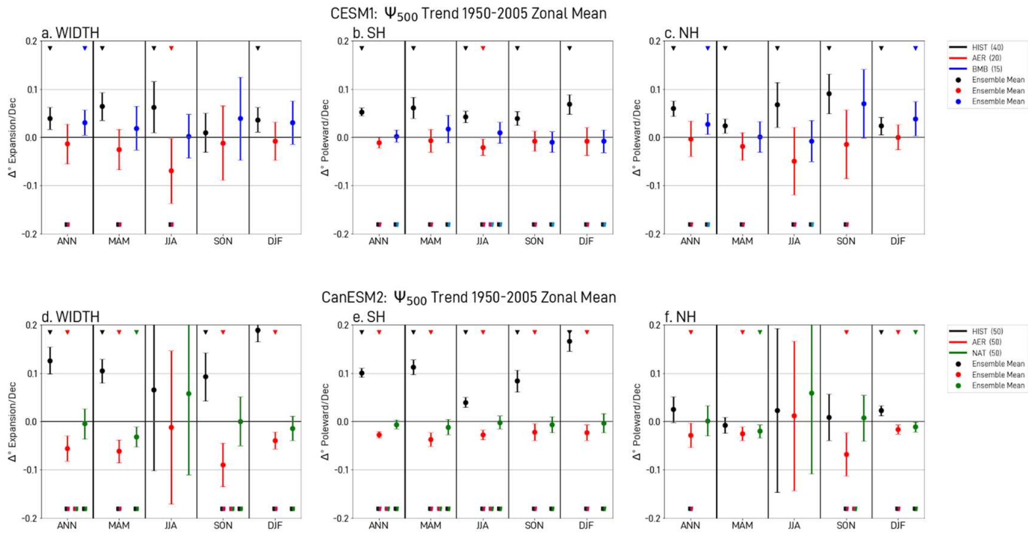

Figure S3: Annual and seasonal mean $\phi_{\Psi 500}$ trends for 1950-2005 in (a-c) CESM1-LE HIST (black), AER (HIST - XAER; red) and BMB (HIST - XBMB; blue), and (d-f) CanESM2-LE HIST (black), AER (red) and NAT (green), for (a, d) SH, (b, e) NH and (c, f) tropical width. Ensemble means are shown by large dots with 95% confidence intervals shown by whiskers. At the top, downward pointing triangles show where the ensemble mean is significant from zero. Coloured squares at the bottom indicate where experiment ensembles are significantly different from each other e.g. a red/black square indicates significant difference between HIST and AER distributions ($p < 0.05$).

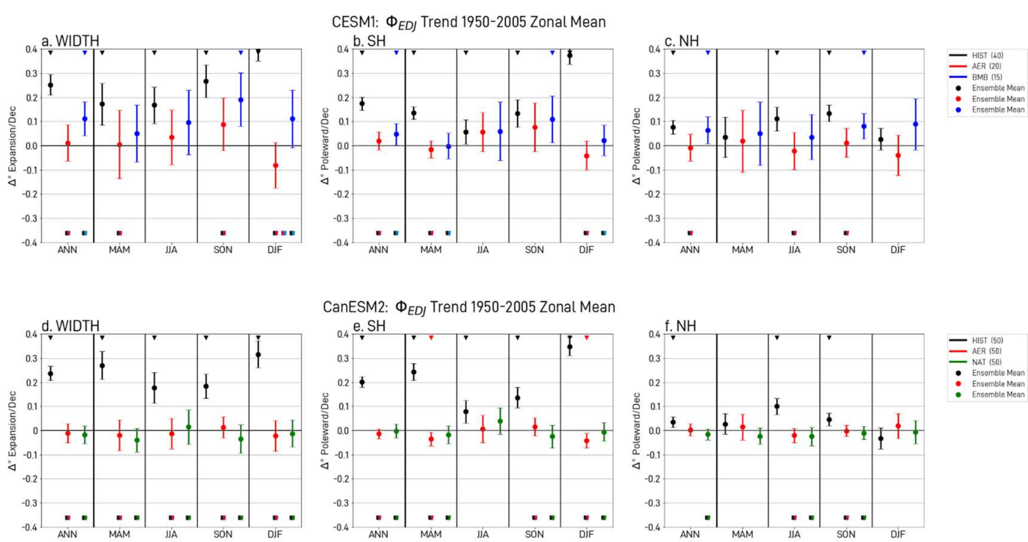

Figure S4: As in Figure S3 but for ϕ_{EDJ}.

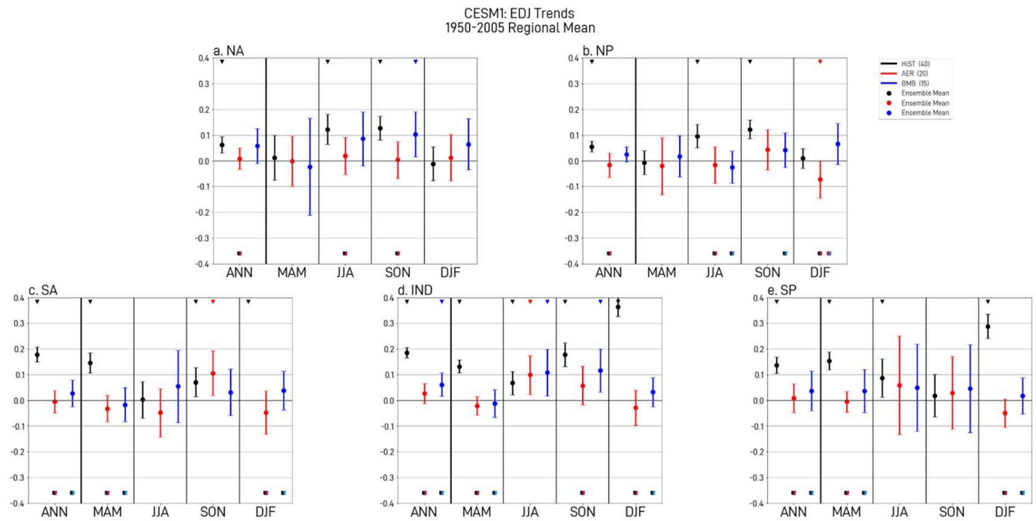

Figure S5: As in Figure 3, but for CESM1-LE regional mean ϕ_{UAS} trends. Northern hemisphere regions are a) North Atlantic (NA) and b) North Pacific (NP) and Southern hemisphere regions are c) South Atlantic (SA), d) Indian (IND) and e) South Pacific (SP). Basins defined as in Wood et al. (2020b).

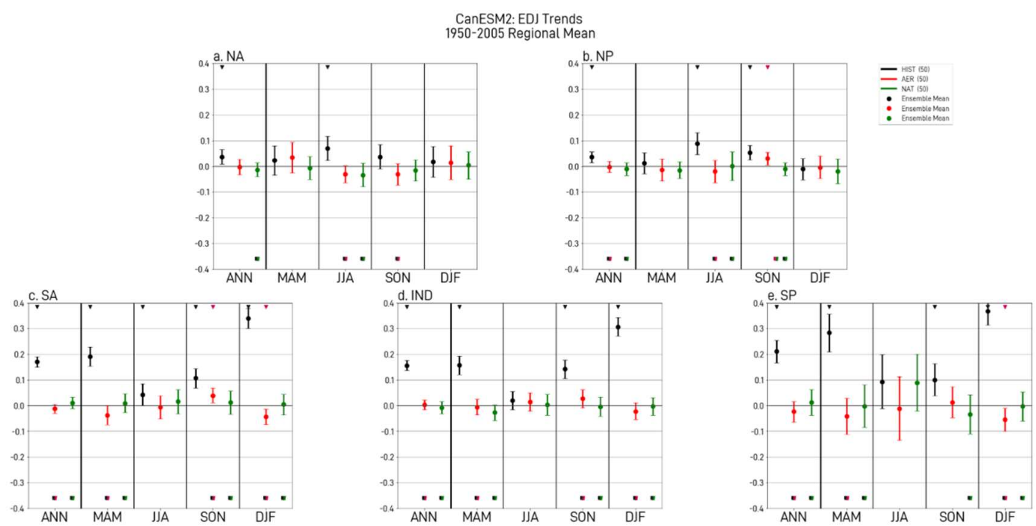

Figure S6: As in Figure S5 but for CanESM2.

Supplementary Methods Information:

Near-Surface Zonal Wind Zero Crossing (UAS)

The UAS metric captures the latitude where there is a change in sign of near surface zonal wind between tropical easterlies and extratropical westerlies (ϕ_{UAS}). It is calculated as the zero-crossing of the near-surface wind polewards of the subtropical minimum and equatorward of 60° (Adam et al., 2018).

Eddy-Driven Jet (EDJ)

The latitude of the eddy-driven jet (EDJ), ϕ_{EDJ}, is defined as the latitude of the centroid of 850 hPa zonal wind between 15° and 70°:

$$\phi_{\text{EDJ}} = \frac{\int_{15^\circ}^{70^\circ} \phi \bar{U}^n d\phi}{\int_{15^\circ}^{70^\circ} \bar{U}^n d\phi},$$

where ϕ is latitude and U is zonal mean zonal wind at 850 hPa. Following Waugh et al. (2018), n = 30 in the SH ('peak' method in TropD) and n = 6 in the NH ('max' method in TropD) as these choices have been found to locate the EDJ in each hemisphere.

Meridional Mass Stream Function (Ψ_{500})

$\phi_{\Psi 500}$ is the latitude at which the zonal mean meridional mass streamfunction at 500 hPa crosses zero. This is considered the edge of the Hadley circulation and is a widely used metric for tropical width (e.g. GD20; Hu and Fu, 2007).